Report: Ocean-Atmospheric Linkages

Stephen R. Rintoul

This chapter focuses on the role of the ocean in the global carbon cycle on the time scale of decades to centuries. The input rate of CO_2 to the atmosphere due to fossil fuel burning and deforestation has continued to increase over the last century. To balance the global carbon budget, a sink is required whose magnitude is changing on similar time scales. We have sought to identify aspects of the ocean system that are capable of responding on decadal time scales, to examine our present ability to model such changes, and to pinpoint ways in which this ability could be improved. Many other important aspects of the ocean's role in global change are not addressed, including the importance of oceanic heat transport and thermal inertia to the climate system, biogeochemical cycling of elements other than carbon, and the importance of the ocean as a source or sink of trace gases.

Estimates of Present Oceanic Uptake of CO_2

The chemical buffering of the ocean carbon system and the large size of the oceanic carbon reservoir relative to that of the atmosphere place a fundamental limit on the uptake of CO_2 by the ocean. The Revelle factor is a measure of the change in ocean pCO_2 required for a given change in atmospheric pCO_2 to establish a new equilibrium. This number is around 10 for most of the surface ocean, expressing the fact that for a given change in atmospheric CO_2 the expected change in oceanic pCO_2 is an order of magnitude smaller.

A second constraint on the extent to which the ocean can be a net sink of atmospheric CO_2 is due to the inherent negative feedback of

the ocean-atmosphere system. In a system in equilibrium, the net oceanic sources and sinks of CO_2 will be in balance. If the equilibrium is disturbed, for example, by circulation changes at high latitude resulting in more efficient CO_2 uptake by the ocean, atmospheric pCO_2 will decrease. In turn, the oceanic source regions will become more effective and the sinks less so, tending to restore atmospheric pCO_2 toward its equilibrium value.

The flux of CO_2 across the air-sea interface is equal to the product of the gas transfer coefficient and the difference in pCO_2 between the atmosphere and the surface ocean. Direct estimates of the air-sea exchange of CO_2 based on observed δpCO_2 suggest that the ocean is taking up 1.6 Gt/yr (Tans et al., 1990); however, the uncertainty in this estimate is large (range of 0.8–2.4 Gt/yr; see Bolin and Fung, this volume). Expressions for the dependence of the gas transfer coefficient on wind speed vary by a factor of two. A more serious problem is the lack of pCO_2 measurements in the surface ocean. Over large portions of the globe, particularly in the Southern Hemisphere, there are no observations at all. Compounding the sampling problem is the fact that oceanic pCO_2 is highly variable, both in space and in time. Accurate direct estimates of the net exchange of CO_2 across the air-sea interface will require at a minimum seasonal data at all latitudes.

A variety of models, ranging in complexity from relatively simple box-diffusion models to three-dimensional general circulation models (GCMs), have been employed to simulate the uptake of CO_2 by the ocean (Oeschger et al., 1975; Bolin, 1981; Siegenthaler, 1983; Peng and Broecker, 1985; Maier-Reimer and Hasselmann, 1987; Sarmiento et al., 1992). The model calculations suggest that the ocean is taking up 26–44% of the fossil fuel CO_2 input to the atmosphere. However, this range of values probably does not reflect the true uncertainty in the CO_2 uptake by the ocean. The ocean circulation in the models used so far to estimate the oceanic sink of CO_2 has been in steady state and much more diffusive in nature than that of the real ocean. In addition, the circulation and mixing have generally been calibrated by fitting to the same "clock," based on ^{14}C observations. In the simpler models, there are usually insufficient degrees of freedom to simultaneously fit more than one tracer distribution (Bolin, 1986). Given these similarities, it is not surprising that all of the models result in a similar uptake of the CO_2 transient. However, in box models in which geostrophy is included, imposing on the system a different "clock" that is based on ocean dynamics, the uptake of CO_2 is considerably more efficient (Moore, this volume).

Thus there remains a large uncertainty in both direct and model-based estimates of the amount of fossil fuel CO_2 presently entering

the ocean. Additional observations that resolve the seasonal cycle are critically needed, especially in the Southern Hemisphere. Accurate estimates based on observations would provide a critical test of the models we hope to use to predict the climatic response to increasing atmospheric CO_2 concentrations. The atmosphere and the upper ocean equilibrate with respect to CO_2 on a time scale of about one year (Broecker and Peng, 1982). The rate-limiting step in the oceanic uptake of CO_2 is thus the exchange between the surface ocean and the ocean interior (Oeschger et al., 1975; Broecker et al., 1980; Sarmiento et al., 1992). Reliable predictions of the future oceanic sink therefore depend critically on accurate treatment of the physical and biological mechanisms that transport carbon from the sea surface to the deep ocean.

Physical Transport of Carbon into the Ocean Interior

As mentioned above, the models so far used to estimate the oceanic uptake of CO_2, whether box models or coarse-resolution GCMs, are heavily diffusive. The penetration of tracers into the ocean interior thus goes as the square root of time. In reality, we believe the pathways between the upper ocean and the interior are primarily advective. Sites of strong mixing are sparse and of limited extent, with the end products of active vertical exchange at these sites carried into the interior by flow along isopycnals. The penetration of the tracer in this case is linear in time. Either type of model can be calibrated to a transient tracer such as ^{14}C at a particular time, but the time evolution in each case will be very different. Recent simulations of tritium penetration into the subtropical gyre of the North Atlantic confirm that higher-resolution models with less diffusive circulations result in more efficient tracer injection into the interior (F. Bryan, personal communication).

A variety of vertical mixing processes act to transfer physical and chemical properties between the mixed layer and the ocean interior. Each vertical exchange mechanism is characterized by different time and space scales. The mixing processes which represent the exchange between the thermocline and deeper layers of the ocean play the greatest role in oceanic uptake on decadal time scales. In particular, we are interested in the potential for variability in the vertical exchange mechanisms on this time scale.

The Thermohaline Circulation

The global thermohaline circulation, or "conveyor belt," plays a major role in carrying CO_2 and tracers from the sea surface to the deep interior. Wust (1935), for example, concluded from deep oxygen measurements that North Atlantic deep water was the primary

agent responsible for the ventilation of the deep ocean. More recent observations of tracers such as tritium and freon confirm that the formation and export of intermediate and deep water masses provide an efficient conduit from the sea surface to the deep ocean (Jenkins and Rhines, 1980; Weiss et al., 1985).

Although the importance of the thermohaline circulation in ventilating the deep ocean has been recognized for many years, fundamental aspects of its operation are still not well understood. We do not have a good conceptual model of the forces driving the conveyor belt or the factors determining the transport of the overturning cell. We know even less about how the global thermohaline circulation is closed. Consequently, we are unable to predict how the conveyor belt will change in response to changes in atmospheric forcing.

Model studies have suggested that the thermohaline circulation is delicately balanced, and may be characterized by significant variability. Stommel (1961) and Bryan (1986) demonstrated the possibility of multiple solutions for the thermohaline circulation in simplified ocean models. Coupled ocean-atmosphere models can also lead to nonunique solutions. For example, Manabe and Stouffer (1988) showed that for the same external forcing a coupled model exhibited two possible climates, one with an active thermohaline circulation in the North Atlantic and another with almost no Atlantic thermohaline overturning.

If an atmospheric GCM is forced by the observed sea surface temperature distribution, it is possible to determine from the solution the surface fluxes of heat and moisture over the ocean required for a balanced equilibrium. If an ocean model is forced by the observed sea surface temperature and salinity fields, another set of surface fluxes of heat and moisture is required for equilibrium. At present, the fluxes derived from atmospheric and oceanic models do not agree with each other. As a result, when such models are coupled, the model climate is apt to drift rapidly away from the observed climate toward another equilibrium state with a very different ocean circulation and sea surface temperature distribution. In order to make progress in the face of this inconsistency, ad hoc constant corrections to the surface fluxes have been used to prevent the model climate from drifting away from that observed (Sausen et al., 1988; Manabe and Stouffer, 1988). This approach has been applied by Stouffer et al. (1989) in a coupled model simulation of greenhouse warming. In a control experiment, the coupled model oscillates about a climate with a sea surface temperature distribution close to what is observed for an extended integration of 150 years.

A weakness of this approach is that it is not clear that it can be used to simulate climates that are very different from the present

one, excluding many interesting applications to paleoclimate or to extreme greenhouse warming scenarios. Most of the flux corrections required in the Stouffer et al. (1989) calculation appear to be due to inadequacies in the cloud model used and to the poor resolution of heat and salinity transport by ocean currents. These are both areas where active efforts are being made to improve the models.

The major shifts in the thermohaline circulation (e.g., reversal of the overturning cell in the North Atlantic) seen in some of the model results require approximately 1000 years to occur. However, the transport or properties of the thermohaline circulation may be modulated on shorter time scales. There is some observational evidence that such changes may have occurred in recent decades. A layer of anomalously fresh water capped the subpolar North Atlantic in the 1970s and may have prevented the convective formation of deep water (Schlosser et al., 1991). Roemmich and Wunsch (1985) observed a shift in the depth of the deep branch of the North Atlantic thermohaline circulation between 1959 and 1981, but no change in the volume transport.

The deep variability noted by Roemmich and Wunsch may be due to changes in the thermohaline characteristics of the upper ocean such as the salinity anomaly observed in the 1970s. However, our models of the thermohaline circulation are inadequate to predict the change in properties or transport of deep water in response to a change in atmospheric forcing. More realistic modeling of deep convection will require a more detailed treatment of the upper ocean than currently appears in oceanic GCMs. Suggestions for such a model are discussed below.

Thermocline Ventilation

The thermohaline conveyor belt is not the only mechanism of vertical exchange that plays a role in the oceanic uptake of CO_2 on decadal time scales. Both convective formation of intermediate waters and Ekman pumping by the wind inject water masses from the sea surface into the thermocline. The relative importance of deep and intermediate water mass formation in supplying tracers to the ocean interior is not well known. However, several studies suggest that the major input is through ventilation of the thermocline. Stuiver (1978), for example, concluded at the time of the Geochemical Ocean Sections Study (GEOSECS) expedition that 80% of the bomb-produced radiocarbon was found in the thermoclines of the subtropical gyres. Atmospheric inverse calculations that calculate the sources and sinks of CO_2 implied by the atmospheric distribution also suggest that the primary oceanic sink is at midlatitudes (Enting and Mansbridge, 1989). Coupled model studies show that the ocean warming resulting

from a doubling of atmospheric CO_2 is concentrated in the subtropical gyres at midlatitude (Schlesinger et al., 1988).

The volume and physical and chemical properties of the water transferred from the sea surface to the interior depend on the fluxes of heat and moisture and the wind stress at the sea surface. Variability in the forcing thus leads to variability in the net uptake of tracers by the ocean. We know that the forcing varies on the time scale of years to centuries, but we do not have observations of the resulting changes in the vertical exchange driven by these processes, except in a few instances.

To model such changes, we need to improve the treatment of the upper ocean in GCMs. No existing ocean model, for example, does a good job of reproducing the large-scale sea surface temperature distributions. An ocean model that is appropriate to couple to an atmospheric GCM will require higher vertical resolution in the upper ocean and must be able to reproduce the observed temperature and salinity when driven with realistic heat and moisture fluxes at the sea surface. In addition, the model should include a physically based mixing parameterization (e.g., shear- or stability-dependent vertical mixing, lateral mixing along isopycnals). For short-term climate predictions (approximately 10–100 years), resolving the deep ocean is not necessary. A climate model that includes a highly resolved mixed layer and thermocline patched to a coarse-resolution deep ocean would be most appropriate for studying the uptake of CO_2 on decadal time scales. Such models are now being developed.

A critical test of improved models of the upper ocean will be how well they reproduce the seasonal cycle and interannual variability of temperature and salinity in the upper several hundred meters. However, the existing observations are inadequate for verifying models on the basin scale. The primary source for measurements of these properties is ships of opportunity, supplemented by moored and drifting buoy systems away from shipping routes. The observational network must be substantially expanded to supply the global data sets needed. Satellite measurements of sea ice and sea surface temperature provide valuable complements to the in situ measurements.

Biologically Mediated Carbon Transport to the Ocean Interior

The export of organic matter from the euphotic zone and its regeneration at depth play a major role in determining the large-scale nutrient and carbon distributions (Broecker and Peng, 1982). Changes in ocean productivity at high latitude may have contributed to the large changes in atmospheric CO_2 that accompanied

the glacial-to-interglacial transition (Siegenthaler and Wenk, 1984; Knox and McElroy, 1984; Sarmiento and Toggweiler, 1984). However, the capacity of the biological system to change on decadal time scales is limited. Carbon, nitrogen, and phosphorus occur in marine organic matter in fixed ratios (i.e., Redfield ratios; Redfield et al., 1963). Export of organic carbon is thus accompanied by an export of nitrogen and phosphorus. When organic matter is decomposed at depth, the nutrients and carbon are remineralized, again in the same ratio. Mixing of the regenerated nutrients into the euphotic zone can support the formation of more organic matter, but the regenerated carbon is mixed up as well. Thus, in the equilibrium case the biological cycling results in no net flux of carbon into the ocean interior. For the biological pump to transfer carbon from the surface ocean to the interior, the stoichiometric constraint imposed by the Redfield ratios must be broken.

Dissolved Organic Matter

Recent observations of a much larger pool of dissolved organic carbon (DOC) than was previously believed to exist challenge several long-held assumptions of biological and chemical oceanography (Sugimura and Suzuki, 1988; Toggweiler, 1989). If the measurements so far obtained in the North Pacific and North Atlantic are representative, the DOC reservoir in the ocean is comparable in magnitude to the terrestrial carbon reservoir. Small changes in the residence time of such a large carbon reservoir, perhaps in response to changes in temperature, might have a significant impact on atmospheric pCO_2. DOC can be transported laterally for long distances between its formation and decomposition, so that the traditional one-dimensional view of particle formation in the euphotic zone, rapid sinking, and regeneration at depth may not be applicable (Toggweiler, 1989). If the C:N:P ratio of DOC is not in the Redfield proportion, then our understanding of the biological pump needs to be substantially revised.

However, very little is known at this point about the nature of the "new" DOC. In particular, we do not know the mechanisms responsible for the formation and destruction of this material, or how rapidly these processes operate. The ratio of nitrogen and phosphorus to carbon in the pool of dissolved organic matter (DOM) is also unknown. Some measurements suggest that most of the oxygen consumption occurring in the ocean is due to the oxidation of DOM (Sugimura and Suzuki, 1988), which seems to conflict with the observed decrease in organic particle flux with depth, and which also seems roughly consistent with observed AOU (Martin et al., 1987). Until these issues are resolved it is difficult to evaluate the

role of DOM in the ocean's biogeochemical cycles. The analogy to the terrestrial DOC pool (e.g., humic and fulvic acids) should be explored more thoroughly to see whether additional insight can be gained into the character of the oceanic DOC reservoir. For example, it is known that aromatics in terrestrial DOC are resistant to ultra-violet oxidation but can be oxidized by high-temperature methods such as those used by Sugimura and Suzuki (1988).

Iron Limitation

Over large regions of the ocean, in particular the Southern Ocean, there are significant concentrations of the major nutrients nitrate and phosphate in the mixed layer, suggesting that some other factor is limiting growth. Traditionally the limiting factor has been assumed to be either light, temperature, or grazing by zooplankton. An alternative hypothesis, that the absence of the trace nutrient iron is limiting plankton growth, has been put forward by Martin and co-workers (Martin and Fitzwater, 1988; Martin et al., 1989, 1990). This has led to the suggestion that iron fertilization of the high-latitude ocean could lead to an increase in productivity and a net uptake of CO_2 from the atmosphere.

The effectiveness of such a strategy is a topic of active debate (Peng and Broecker, 1991; Peng et al., this volume; Joos et al., 1991; Keir, 1991). The species that dominate in incubation studies that have been spiked with iron are not the same as those that tend to dominate in the natural system. In the open ocean, aspects of the marine ecosystem that are not well represented in closed incuba-tions, such as herbivory, may limit the net fertilization effect. The effect may also be constrained by limitations imposed by other nutrients, such as phosphorus. Perhaps more important is the inherent negative feedback of the coupled ocean-atmosphere system mentioned above. Experiments with a simple box-diffusion model (see Peng et al., this volume) suggest that as atmospheric CO_2 decreases, the low-latitude oceanic sources of CO_2 become more effective, acting to damp the initial response. The equilibrium atmospheric pCO_2 is only slightly lower than the initial value. Fur-ther experiments with more sophisticated models are needed before the effectiveness of iron fertilization can be evaluated.

Role of Calcareous Organisms

Organisms forming calcium carbonate structures or tests lead to vertical fractionation of carbon and nutrients in the water column. Soft organic tissue is decomposed rapidly in the upper ocean, regen-erating both carbon and nutrients. The calcium carbonate shells, on the other hand, sink to the deep sea before dissolving, carrying car-

bon but not nitrogen or phosphorus to the deep ocean. Therefore an increase in the relative abundance of calcareous organisms, such as coccoliths, could result in increased sequestering of carbon at depth. An increase in the formation and export of calcium carbonate from the upper ocean also decreases surface alkalinity and thus increases oceanic pCO_2. If the climatic changes associated with an increase in atmospheric CO_2 favored calcareous organisms for some reason, there would be a potential for a positive feedback. However, we do not understand the factors controlling community structure well enough to predict what environmental changes could lead to a relative increase in the population of calcareous organisms.

Changes in Community Structure and Bloom Frequency

More general changes in the community structure of marine ecosystems in response to changing environmental conditions may affect the uptake of CO_2 by the ocean. A general feature of ocean ecosystems is that stable environmental conditions lead to a community of multiple, coadapted species which is effective at recycling nutrients, while episodic environments are dominated by blooms of one or a few species followed by large export of organic matter from the euphotic zone (Williams and von Bodungen, 1989). Therefore, changes in the variability of environmental variables relevant to the biological system (e.g., mixed-layer depth or nutrient supply) can lead to changes in the export of organic carbon.

A net increase in the sequestering of carbon in the deep sea would result if changes in the circulation or mixing occurred in regions of the ocean where unutilized nutrients exist in the surface layers, allowing for production of new biomass. We know that the physical environment can change on decadal time scales, but we are not yet able to predict how such changes may lead to changes in community structure and thus net export.

Perhaps the best evidence for a biological response to physical forcing is at the mesoscale (e.g., Nelson et al., 1989). Satellite observations and field campaigns in the 1980s showed that the cores of eddies associated with western boundary currents evolve differently from the waters of origin and the waters surrounding the eddy. The fundamental explanation for this lies in the different conditions for convection and restratification that exist within the eddy. These result both from the change in air-sea interaction occurring as the core of the eddy is displaced from the region of origin, and from the vertical motion of density surfaces as the eddy spins down. Studies within the cores of such eddies have also provided the opportunity to observe the direct biological response to storm-induced mixing. The net effect of all of these processes is to enhance the upward flux of nutrients into

the euphotic zone of the eddy, leading to increased rates of production and a shift in community composition to large diatoms, which tend to export a larger fraction of the total production.

Marine Productivity and Variability in Surface Ocean pCO_2

One way in which marine biology does affect the uptake of CO_2 on short time scales is in modifying the chemical properties of the surface waters introduced to the interior by physical processes. High productivity at the start of the spring bloom can draw down surface pCO_2 by more than 100 ppm in a few weeks (Codispoti et al., 1982). The physical processes exchanging surface water with deeper layers, such as deep convection or Ekman pumping, also vary on seasonal and shorter time scales. Therefore it is essential to account for the interaction between physical transport and biologically induced variability in pCO_2 in models and observational studies.

Sources and Sinks of Trace Gases

The ocean plays an important role as a source or sink of various trace gases that have an impact on the radiative or chemical properties of the atmosphere (Table 1). Both the production and exchange of these species across the air-sea interface are in many cases mediated by biological processes. Although this chapter cannot discuss this issue in detail, we feel it is important to mention some of the climatically important trace gases in the present context, since it may be that changes in the strengths of the sources or sinks of these species are the most important way in which ocean biology can affect climate. To date there have been no attempts to model the biologically mediated sources or sinks of these gases, or their potential variability in response to changes in ocean circulation or atmospheric forcing.

Improved Coupled Physical-Biological Models

Most coupled physical-biological ocean models have focused on a particular small-scale phenomenon (Rintoul, this volume). Recently, work has begun to couple simple food web models to basin-scale ocean GCMs (Sarmiento et al., 1989). However, ocean GCMs perform most poorly in the regions of most relevance to the biology: the mixed layer and the ocean margins. The mixed-layer–thermocline climate model discussed in the previous section would also be well suited to coupling to a biological model.

The biological component of these coupled models needs refinement so that climatically relevant changes in community structure can be simulated. A promising approach may be to focus on func-

Table 1: Important trace gas species with a marine source or sink

Species	Atmospheric Lifetime	Importance	Controls
Sources			
CH_3Cl	Long	Natural source of stratospheric chlorine	Surface bio source (?), monolayer/gas exchange
DMS	Short	Source of CCN, sulfate aerosols, sulfonic acids	Surface bio source, monolayer/gas exchange
Hydrocarbons	Short	"Remote" source of reactive hydrocarbons	Surface bio source, monolayer/gas exchange
N_2O	Long	Possibly 10-20% of global budget	Rapid upwelling, denitrification
Chloroforms, Bromoforms	Short	Reactive halogens in boundary layer	Surface bio source, monolayer/gas exchange
Sinks			
CH_3CCl_3	Long	Uptake and loss may be 25% of total	Exchange, solubility, hydrolysis (loss)
CFCs	Long	Long-term reservoir (100-yr time scale)	Exchange, solubility, deep circulation

tional groups defined by biogeographic provinces or some other descriptor of the physical environment. Such a model requires observations of the physical and biological characteristics as a function of time so that the functional groups and the rules governing transitions between them can be defined. Mesoscale observational and modeling studies at representative sites are necessary to develop the understanding of the essential processes required to design realistic parameterizations for large-scale models. On the global scale, the ship-of-opportunity observations should be expanded to include measurements of optical depth and fluorescence, pH, total CO_2 (TCO_2), alkalinity, pCO_2, nitrate, and silicate. These measurements can now be done continuously while the ship is under way or on stored samples. Optical instruments on moored or drifting buoys and satellite measurements of ocean color would be valuable in filling the gaps between shipping routes.

Summary and Recommendations

This chapter has focused on the role of the ocean in the uptake of CO_2 on decadal time scales. Direct estimates of the oceanic CO_2

sink are very uncertain due to the lack of surface ocean pCO_2 measurements. Simple models predict that the ocean is taking up 26–44% of the fossil fuel input (Tans et al., 1990). However, all of the models are excessively diffusive and are calibrated by the same tracer, ^{14}C. The small range of present estimates probably reflects the similarity of the models used rather than the true uncertainty in the magnitude of the oceanic sink.

Several factors combine to determine the magnitude and timing of oceanic uptake of CO_2. The coupled air-sea system has an inherent negative feedback: If atmospheric pCO_2 decreases, perhaps due to an increase in high-latitude biological production, the oceanic source regions will become more effective and the oceanic sinks less so, tending to restore the system toward equilibrium. The Revelle factor also imposes a fundamental limitation on the capacity of the ocean to respond to atmospheric pCO_2 changes.

The ultimate factor determining the rate of oceanic uptake of CO_2 is the transfer of carbon to the interior by physical and biological processes. The physical transport pathways are primarily advective, connecting sparse sites of active "vertical" exchange. Both the exchange processes and the circulation between them have the capacity to change substantially on the 10–100-year time scale. However, present models do not treat the upper ocean well enough to predict how the vertical exchange processes will change in response to changes in atmospheric forcing.

The potential for changes in the biology to lead to significant changes in the oceanic sink of CO_2 on decadal time scales is probably small. Possible exceptions include changes in the size or residence time of the dissolved organic matter pool, an increase in the relative importance of calcareous organisms, and changes in the community structure in regions of high surface nutrient concentrations. Our present understanding of the dynamics of these pools or processes is so poor that we do not know whether changes in them can lead to changes in the oceanic uptake of carbon on these time scales.

Present models are not able to predict how the ocean circulation or biology will change in response to changes in the atmospheric forcing, or the impact of such changes on climate or the carbon cycle. We recommend that the following steps be taken in the design of more realistic models:

- Ocean circulation models must do a better job of simulating the primarily advective nature of the pathways between the mixed layer and the ocean interior. This will require higher-resolution models with physically based mixing parameterizations (e.g., shear- or stability-dependent mixing, isopycnal mixing).

- Ocean models must be forced with surface fluxes of heat and moisture, rather than restoring sea surface temperature and salinity to observed climatology, if they are to be used for predicting the response to changing atmospheric forcing.

- For coupling ocean circulation models to atmospheric GCMs or to marine ecosystem models, model treatment of the upper ocean must be improved. A model with a highly resolved mixed layer and thermocline and a coarsely resolved deep ocean would be appropriate for climate studies on decadal time scales. A critical test of such a model will be reproducing the observed seasonal cycle and interannual variability of upper ocean temperature, salinity, and biological properties.

- Improved upper ocean models therefore require basin-scale, long-term observations of the seasonal cycle of temperature and salinity in the upper several hundred meters. For ocean biology models, measurements of nitrate, silicate, carbon species (pH, TCO_2, alkalinity, pCO_2), and optical depth/fluorescence (as a measure of biomass) are required. These observations could be obtained under way from ships of opportunity. Satellite measurements of sea surface temperature, sea ice extent, and ocean color would also be useful for verifying such models. The use of moored or drifting buoys with temperature and salinity chains and optical instruments would help fill the data gaps away from shipping routes.

- Marine ecosystem models appropriate for basin scales need to be developed. In particular, the model must be able to simulate climatically relevant changes in community structure or functional groups in response to changes in the physical environment. Progress is most likely to come from building on the experience gained from mesoscale observational and modeling studies.

- More observational work is urgently needed on the role of the ocean as a source/sink of radiatively or chemically active trace gases.

References

Bolin, B. (ed.). 1981. *Carbon Cycle Modeling*. SCOPE 16, John Wiley and Sons, Chichester, England.

Bolin, B. 1986. Requirements for a satisfactory model of the global carbon cycle and current status of modeling efforts. In *The Changing Carbon Cycle: A Global Analysis* (J.R. Trabalka and D.E. Reichle, eds.), Springer-Verlag, New York, 403-424.

Broecker, W.S., and T.-H. Peng. 1982. *Tracers in the Sea.* Eldiglo Press, Lamont-Doherty Geological Observatory, Palisades, New York.

Broecker, W.S., T.-H. Peng, and R. Engh. 1980. Modeling the carbon system. *Radiocarbon 22,* 565–598.

Codispoti, L.A., G.E. Friederick, R.L. Iverson, and D.W. Hood. 1982. Temporal changes in the inorganic carbon system of the southeastern Bering Sea during spring 1980. *Nature 296,* 242–244.

Enting, I.G., and J.V. Mansbridge. 1989. Seasonal sources and sinks of atmospheric CO_2: Direct inversion of filtered data. *Tellus 41B,* 111–126.

Jenkins, W.J., and P.B. Rhines. 1980. Tritium in the deep North Atlantic Ocean. *Nature 286,* 877–880.

Joos, F., J. Sarmiento, and U. Siegenthaler. 1991. Estimates of the effect of Southern Ocean iron fertilization on atmospheric CO_2 concentrations. *Nature 349,* 772–775.

Keir, R.S. 1991. Ironing out greenhouse effects. *Nature 349,* 198.

Kempe, S. 1979. Carbon in the rock cycle. In *Carbon Cycle Modeling* (B. Bolin, ed.), SCOPE 16, John Wiley and Sons, Chichester, England, 343–377.

Knox, F., and M. McElroy. 1984. Changes in atmospheric CO_2: Influence of biota at high latitudes. *Journal of Geophysical Research 89,* 4629–4637.

Maier-Reimer, E., and K. Hasselmann. 1987. Transport and storage of carbon dioxide in the ocean, and in organic ocean-circulation carbon cycle model. *Climate Dynamics 2,* 63–90.

Manabe, S., and R. Stouffer. 1988. Two stable equilibria of a coupled ocean-atmosphere model. *Journal of Climate,* 841–866.

Martin, J.H., and S.E. Fitzwater. 1988. Iron deficiency limits phytoplankton growth in the north-east Pacific subarctic. *Nature 331,* 341–343.

Martin, J.H., G.A. Knauer, D.M. Karl, and W.W. Broenkow. 1987. VERTEX: Carbon cycling in the northeast Pacific. *Deep Sea Research 34,* 267–285.

Martin, J.H., R.M. Gordon, S. Fitzwater, and W.W. Broenkow. 1989. VERTEX: Phytoplankton/iron studies in the Gulf of Alaska. *Deep Sea Research 36,* 649–680.

Martin, J.H., R.M. Gordon, and S. Fitzwater. 1990. Iron in Antarctic waters. *Nature 345,* 156–158.

Oeschger, H., U. Siegenthaler, U. Schotterer, and A. Gugelmann. 1975. A box diffusion model to study the carbon dioxide exchange in nature. *Tellus 27,* 168–192.

Pearman, G.I., and P. Hyson. 1986. Global transfer and inter-reservoir exchange of carbon dioxide with particular reference to stable isotopic distributions. *Journal of Atmospheric Chemistry 4*, 81–124.

Peng, T.-H., and W.S. Broecker. 1985. The utility of multiple-tracer distributions in calibrating models for uptake of anthropogenic CO_2 by the ocean thermocline. *Journal of Geophysical Research 90*, 7023–7035.

Peng, T.-H., and W.S. Broecker. 1991. Dynamical limitations on the Antarctic iron fertilization strategy. *Nature 349*, 227–229.

Redfield, A.C., B.H. Ketchum, and F.A. Richards. 1963. The influence of organisms on the composition of sea water. In *The Composition of Sea Water: Comparative and Descriptive Oceanography* (M.N. Hill, ed.), The Sea, Vol. 2, Interscience Publishers, New York, 26–77.

Roemmich, D., and C. Wunsch. 1985. Two transatlantic sections: Meridional circulation and heat flux in the subtropical North Atlantic Ocean. *Deep Sea Research, 32*, 619–664.

Sarmiento, J.L., and J.R. Toggweiler. 1984. A new model for the role of the oceans in determining atmospheric pCO_2. *Nature 308*, 621–624.

Sarmiento, J.L., J.C. Orr, and U. Siegenthaler. 1992. A perturbation simulation of CO_2 uptake in an ocean circulation model. *Journal of Geophysical Research 97(3)*, 3621–3647.

Sausen, R., K. Barthel, and K. Hasselmann. 1988. Coupled ocean-atmosphere models with flux correction. *Climate Dynamics 2*, 145–163.

Schlesinger, M.E., W.L. Gates, and Y.-J. Han. 1988. The role of the ocean in CO_2-induced climate change: Preliminary results from the OSU coupled atmosphere-ocean general circulation model. In *Coupled Ocean-Atmosphere Models* (J.C. Nihoul, ed.), Elsevier, Amsterdam, 447–478.

Schlosser, P., G. Bonisch, M. Rhein, and R. Bayer. 1991. Reduction of deep water formation in the Greenland Sea during the 1980s: Evidence from tracer data. *Science 251*, 1054–1056.

Siegenthaler, U. 1983. Uptake of excess CO_2 by an outcrop-diffussion model of the ocean. *Journal of Geophysical Research 88(C6)*, 3599–3608.

Siegenthaler, U., and T. Wenk. 1984. Rapid atmospheric CO_2 variations and ocean circulation. *Nature 308*, 624–625.

Stommel, H. 1961. Thermohaline convection with two stable regimes of flow. *Tellus 13*, 224–230.

Stouffer, R.J., S. Manabe, and K. Bryan. 1989. Interhemispheric asymmetry in climate response to a gradual increase of atmospheric carbon dioxide. *Nature 342*, 660–662.

Stuiver, M. 1978. Atmospheric carbon dioxide and carbon reservoir changes. *Science 199*, 253–258.

Sugimura, Y., and Y. Suzuki. 1988. A high-temperature catalytic oxidation method for the determination of non-volatile dissolved organic carbon in seawater by direct injection of liquid samples. *Marine Chemistry 24*, 105–131.

Tans, P.P., I.Y. Fung, and T. Takahashi. 1990. Observational constraints on the global atmospheric CO_2 budget. *Science 247*, 1421–1438.

Toggweiler, J.R. 1989. Is the downward dissolved organic matter (DOM) flux important in carbon transport? In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), John Wiley and Sons, New York, 65–83.

Weiss, R.F., W.S. Bullister, R.H. Gammon, and M.J. Warner. 1985. Atmospheric chlorofluoromethanes in the deep equatorial Atlantic. *Nature 314*, 608–610.

Williams, P.J.leB., and B. von Bodungen. 1989. Export productivity from the photic zone. In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 99–115.

Wust, G. 1935. Schichtung und Zirkulation des Atlantischen Ozeans: Die Stratosphäre. Reprinted in *The Stratosphere of the Atlantic Ocean*, W.J. Emery, ed. Amerind, New Delhi, 1978, 112 pp.