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**GLOBAL CLIMATE FEEDBACKS
PROCEEDINGS OF THE BROOKHAVEN NATIONAL LABORATORY WORKSHOP
JUNE 3-6, 1990**

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FOREWORD

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The unusually hot summers of 87-88 did much to raise the level of consciousness among the public world-wide about the "greenhouse effect". In the popular media there was a great range of opinion from apocalyptic jeremiads to panglossian reassurances. Despite significant disagreement on many specific issues, in the scientific community there was widespread acknowledgment of the need to investigate more thoroughly and expeditiously the scientific aspects of this problem in order to provide policy makers with the best possible basis for decision-making.

The production, conversion, and use of energy constitute the basis for possible significant global climate changes. In July 1989, the President of the United States directed the Secretary of Energy to prepare a National Energy Strategy that explicitly addresses the environmental implications of energy production and use. The National Energy Strategy - still being crafted and due to be delivered to the President before the end of the year - states as its goals: to assume adequate and secure supplies of energy at a reasonable cost, and to improve the environmental quality. Other governments in the industrialized and developing worlds have taken similar positions.

This world-wide concern for the environmental implications of energy policies is reflected in the agenda of the Intergovernmental Panel for Climate Control (IPCC). In the United States, the relevant scientific and policy issues are being coordinated and led by the Committee for Earth and Environmental Sciences (CEES) that has been constituted from several governmental agencies. The Department of Energy, through its National Laboratories

and its sponsored research at academic institutions, has a resource base that is an important component of the international armamentarium that can be arrayed to understand and solve this problem.

Assessing climate alterations and understanding the changes that may be imposed on the earth's energy budget (forcings) and the reactions of the climate system to climate changes (feedbacks), present complex scientific and technological problems, requiring careful and extensive measurements, compilation of an enormous collection of data, and building and validation of predictive models. Temperature increase is an obvious parameter of global climate alteration, although it is not easy to quantify; precipitation and planetary radiation budgets are equally complex. Similarly, with climate forcings, whether they are natural (variations in solar radiation and volcanic dust) or anthropogenic (release of greenhouse gases, aerosols, deforestation), as well as with climate feedbacks (atmospheric water vapor, clouds, ocean heat-exchange processes, terrestrial vegetation), the scientific issues involved require considerable multidisciplinary and long-range efforts if they are to be fully understood.

On behalf of the Department of Energy and in cooperation with other national laboratories and academic institutions, the Department of Applied Science at Brookhaven National Laboratory has made significant contributions to the understanding of problems of energy and the environment. With the recent emphasis at the Department of Energy on global environmental aspects of the use of energy - through such initiatives as Atmospheric Radiation Measurements (ARM) and Computer Hardware, Advanced Mathematics and Model Physics (CHAMMP) - the Department of Applied Science has sought to bring its many pertinent strengths to the solution of the problem, again in cooperation with the other national laboratories.

In this context, the Department of Applied Science convened this WORKSHOP ON CLIMATE FEEDBACKS. We are grateful for the participation of many prominent scientific leaders, and we are convinced that their deliberations, summarized in this volume, will have a significant role in delineating the pertinent issues. We also acknowledge with thanks the support of A. Patrinos of the Department of Energy; the strong encouragement of the Director of Brookhaven National Laboratory, N. Samios; and the indefatigable efforts of the organizers, led by B. Manowitz (Editor of this volume).

ACKNOWLEDGMENTS

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B. Manowitz
Workshop Chairman
October, 1990

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INTRODUCTION TO BROOKHAVEN WORKSHOP ON GLOBAL CLIMATE FEEDBACKS
JUNE 3-6, 1990

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The important physical, chemical, and biological events that affect global climate change occur on a mesoscale - requiring high spatial resolution for their analysis. Only computer-based modeling is likely to be able to provide reliable predictions and many such general circulation models (GCMs) are presently being used.

With the participation of universities and national laboratories, the Department of Energy has formulated two major initiatives under the U.S. Global Change Program: ARM (Atmospheric Radiation Measurements), and CHAMP (Computer Hardware Advanced Mathematics and Model Physics). ARM is designed to use ground and air-craft based observations to document profiles of atmospheric composition, clouds, and radiative fluxes. With research and models of important physical processes, ARM will delineate the relationships between trace gases, aerosol and cloud structure, and radiative transfer in the atmosphere, and will improve the parameterization of global circulation models.

The present GCMs do not model important feedbacks, including those from clouds, oceans, and land processes. The purpose of this workshop is to identify such potential feedbacks, to evaluate the uncertainties in the feedback processes (and, if possible, to parameterize the feedback processes so that they can be treated in a GCM), and to recommend research programs that will reduce the uncertainties in important feedback processes.

To clarify the terminologies, let me introduce a simple globally averaged energy-balance model.

The change in temperature ΔT from a quasi-equilibrium state due to a heating perturbation ΔQ is given by:

$$C \frac{d\Delta T}{dt} + \lambda \Delta T = \Delta Q$$

where C is the heat capacity of the system, t is time, and λ is the feedback strength.

In the absence of feedbacks ($\lambda = 0$), ΔT would increase indefinitely. However, the long-wave cooling to space will always produce a negative feedback.

At the next quasi-equilibrium state:

$$\Delta T = \frac{\Delta Q}{\lambda}$$

If we assume a forcing function due to a doubling of CO_2 of 4 Wm^{-2} , black body cooling provides a feedback strength of 3.7 Wm^{-2} , and:

$$\Delta T = \frac{4}{3.7} = 1.1^\circ\text{K}$$

Note, that by this convention a positive λ implies a negative feedback.

Using this terminology λ s can be considered as cumulative and other feedbacks are simply added to the black body feedback increasing or decreasing λ .

In another convention found in the literature λ' is defined as the reciprocal of λ , as given here. When λ' is used, the cumulative λ' s are not additive. Each λ' has to be interpreted as an amplification or deamplification factor. This latter approach is probably more rigorous because many of the feedback factors are coupled and interactive.

There are other feedbacks from many components in the system that will be affected, either by the temperature rise, or by the increased concentration of greenhouse gases. A 1°K increase in global surface temperature will increase the concentration of water vapor by about 6%, leading to a strong positive feedback between temperature and water vapor. Veerabhadran Ramanathan will probably tell you that we know the feedback strength of water vapor with a relatively high degree of certainty. However, there is a relatively high degree of uncertainty associated with most other feedbacks. What are the other feedbacks? Can we quantify them? Can we describe the uncertainties of individual feedbacks? Can we suggest research that will help resolve these uncertainties? The organization of this workshop is designed to answer some of these questions.

In the first session on Atmospheric Feedbacks, Robert Cess will say that cloud feedbacks are not yet represented in models, and that there is some

uncertainty, even in the sign of the feedback. David Randall will update us on cloud parameterization. Then, there will be comments on the formal presentations: the commentators will include modelers, who will discuss the structure of advanced models, and observational scientists who will describe the reality of the assumptions and constraints in the models.

With regard to Ocean Interactions, we know that the climatic response will be slowed by the large thermal inertia of the oceans, and that the time taken to reach a substantive fraction of the equilibrium response will be inversely proportional to the feedback strength. Kirk Bryan will talk about the role of the ocean in driving climatic change, and other presenters and commentators will elaborate on chemical transients and biological interactions.

Next, the workshop will treat Terrestrial and Cryosphere Interactions with a similar interdisciplinary approach.

Subsequent to the open session, three panels will convene to formulate better descriptions of the individual feedbacks and their uncertainties, and to recommend research programs to reduce these uncertainties.

OBSERVATIONAL DETERMINATION OF THE GREENHOUSE EFFECT*

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Satellite measurements are employed to quantify the atmospheric greenhouse effect G , defined here as the longwave (or IR) radiation energy trapped by the atmospheric gases and clouds. We find G increases significantly with sea-surface temperature. The rate of increase gives compelling evidence for the positive feedback between surface temperature, water vapor amount and the greenhouse effect. The magnitude of the feedback is consistent with that hypothesized by climate models. This study demonstrates an effective method for directly monitoring, from space, future changes in the greenhouse effect.

In its normal state, the earth-atmosphere system absorbs solar radiation and maintains global energy balance by re-radiating this energy to space as infrared or longwave radiation. The intervening atmosphere absorbs and emits the longwave radiation, but since the atmosphere is colder than the surface, it absorbs more energy than it emits upward to space. The energy which escapes to space is significantly smaller than that emitted by the surface. The difference, the energy trapped in the atmosphere, is popularly referred to as the greenhouse effect (G). Some very interesting questions about the greenhouse effect involve feedbacks between the various elements of the climate system (1). These feedbacks tend to enhance or diminish small external perturbations in the radiation balance. One of the most important is the H_2O feedback, proposed by Manabe and Wetherald (2). Consider an initial increase in the global temperature due to the direct IR trapping resulting from, say, an increase in CO_2 . As the surface and the atmosphere warm, the saturation vapor pressure increases exponentially with temperature,

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**Presented by V. Ramanathan

and more H_2O can evaporate into the atmosphere. Since water vapor is a strong greenhouse gas, the extra water vapor traps more longwave radiation, and drives the temperature even higher. This feedback can amplify the original perturbation by as much as a factor of two (1,2).

The greenhouse effect and its feedback on the climate have been explored by hundreds of theoretical and model studies. Satellite radiation budget measurements have also been used to examine the radiative feedbacks in the climate system (3). Yet to our knowledge the radiative trapping by the atmosphere has not been examined in depth with observed data (with the exception of unpublished work by Cess). The lack of such observational work provides the fundamental motivation for this study, which addresses three critical issues: (1) The magnitude of G in the present atmosphere; (2) The quantitative nature of the feedback between surface temperature and G ; and (3) The change in G with time. We provide observational estimates for the former two, and discuss how our findings bear on the third issue. A possible change in G is of urgent concern because of the rate at which human activities have been adding greenhouse gases into the atmosphere.

We obtain G by subtracting longwave radiation escaping to space from estimates of the radiation emitted by the ocean surface. The radiation escaping to space is measured since 1985 by the spaceborne Earth Radiation Budget Experiment (ERBE). Under cloudless skies, the atmosphere traps 146 W/m^2 while clouds increase G by 33 W/m^2 . G increases significantly with surface temperature (T_s), even when it is normalized by the surface emission. The rate of increase of G , which is $3.3 \text{ W m}^{-2}\text{K}^{-1}$, is consistent with the magnitude of the H_2O -greenhouse feedback inferred from climate models and line-by-line radiative transfer calculations. Furthermore, the observed variation of total atmospheric water content (W) with T_s suggests that the normalized greenhouse effect (g) is strongly correlated with W . The observations build a compelling case for the H_2O feedback. The data also reveal a rapid rise in G at the higher temperatures ($T_s > 298 \text{ K}$), the causes of which are not apparent.

DETERMINATION OF THE GREENHOUSE EFFECT

If E is the longwave flux emitted by the surface at a certain location, and F is the flux leaving the top of the atmosphere (TOA) directly above that location, then the greenhouse effect G for that location can be defined as $G = E - F$.

The TOA fluxes, F , have been measured with good accuracy by scanning radiometers on board the spaceborne Earth Radiation Budget Experiment (ERBE) (4). ERBE consists of three satellites carrying identical scanners which detect both longwave (IR) and shortwave (solar) radiation. The ERBE scanners, when viewing directly down, can resolve a region of about $35 \times 35\text{-km}$. The scanners have inflight calibration which assures an accuracy of 1% for the satellite-altitude readings. Those readings are converted to TOA fluxes on a global grid of 2.5° latitude $\times$ longitude. Data for the months of April, July and October 1985 and January 1986 have been processed thus far. A

unique feature of ERBE is that it separates the fluxes for clear skies (cloudless conditions) from fluxes for cloudy skies (mixture of clear and overcast scenes) (5). Hence we can obtain the greenhouse effect of the atmosphere and that of clouds separately. The emission, E , from the surface is estimated from the Stefan-Boltzmann black-body law ($E = \sigma T^4$) with observed surface temperatures. Since, in the infrared, the ocean surface approximates a blackbody to within 1% (6), and since the diurnal surface temperature variation is small in the ocean's mixed layer (less than a few degrees K), the uncertainties in the computed values of E are minimum over oceans. We restrict this study to the open oceans; i.e., we ignore continents and ocean regions covered with sea-ice.

The sea surface temperatures (SSTs) are obtained from the National Meteorological Center daily data sets archived at the National Center for Atmospheric Research (NCAR). The SST values are derived by blending in-situ (ship and buoy) data with satellite temperature retrievals (7).

The SSTs used during the period of this study have an estimated error of about ± 1.0 K (7). The error in the surface emitted flux is simply $\Delta E = 4\sigma T^3 \Delta T$, or about ± 5 W/m². ERBE TOA longwave fluxes have an estimated error of about ± 5 W/m² for monthly averaged values (4,5). Since the errors in E and F are uncorrelated, the error in the the monthly averaged greenhouse effect for a single location is ± 7 W/m². When these values are averaged around a latitude belt, the random error will diminish significantly. Clear-sky fluxes, however, can have systematic errors due to cloud contamination. The magnitude of this error was examined (5) by sorting out homogeneous cloudless regions, i.e., cloudless regions which are surrounded by other cloudless regions and for which the spatial standard deviation of the longwave flux is less than <1% of the clear-sky flux. This stringent homogeneity criterion yielded fluxes which were smaller than the operational values used here by only 4 W/m², suggesting minimal cloud contamination (5). In addition, the clear-sky fluxes when compared with state-of-the-art radiation models agreed within 1% (5). Overall, we expect the error (systematic plus random) in the monthly and regional mean values of G to be between 5 to 10 W/m².

GLOBAL GREENHOUSE EFFECT OF THE ATMOSPHERE AND CLOUDS

Estimates of G over the open oceans for a cloudless atmosphere (clear sky) reveal significant regional variations (Fig. 1b). Globally, for April 1985, the surface emission is 421 W/m² while the TOA emission is 243 W/m², resulting in a total trapping of 178 W/m². Of this, the atmospheric gases (H₂O, CO₂, and O₃ among others) trap 146 W/m², while clouds trap the remaining 33 W/m². The greenhouse effect for each of the months studied is:

Month	Clear Sky	Clouds	Cloudy Sky*
April 85	146.3	32.1	178.4
July 85	147.7	31.3	179.0
October 85	146.9	32.9	179.8
January 86	143.7	34.1	177.8
4 Month Mean	146	33	179

Since the months are taken from the four seasons, we will assume the four month mean to be representative of the annual average. Then the annual average value of G (179 W/m^2) is comparable to the global mean solar absorption, about 237 W/m^2 . The global mean latent heat released in the atmosphere is about 100 W/m^2 . The greenhouse effect of a doubling of CO_2 is 4 W/m^2 and that of human activities during the past century (8) is about 2 W/m^2 . The greenhouse effect of the atmosphere, next to solar absorption, is the largest factor in maintaining the present climate.

Why does the atmosphere trap longwave radiation? The TOA flux, F , is given by the radiation transfer equation:

$$F = B(T_s) - \int_0^1 A(x) [dB/dx] dx \quad (1)$$

This equation is written in a normalized pressure co-ordinate, $x = p/p_s$, where p_s is the surface pressure. The blackbody emission, B , is simply $\sigma T^4 = \int B_\lambda(T) d\lambda$, where B_λ is the monochromatic Planck function evaluated at atmospheric temperature T . The effective absorptivity, A , is the integral of the monochromatic absorptivity A_λ weighted with dB_λ , and normalized by dB (2). Thus $A(x)$ is the absorptivity between the TOA ($x=0.0$) and the pressure level x . T_s is the surface temperature. Since $G = E - F$, with $E = B(T_s)$, we obtain for G :

$$G = 4\sigma \int_0^1 A(x) T^3 [dT/dx] dx \quad (2)$$

where we have let: $(dB/dx) = 4\sigma T^3 (dT/dx)$. Most of the contribution to A comes from the troposphere (the bottom 10 km) where the bulk of the absorber mass resides. (Although CO_2 is uniformly mixed in the atmosphere, the major absorbers, water vapor and clouds, are almost entirely restricted to the troposphere.) In the troposphere temperature generally decreases with height. Hence $(dT/dx) > 0.0$, and G is positive. This reveals that for G to be positive, the temperature must decrease with altitude in the region of the absorbing gas.

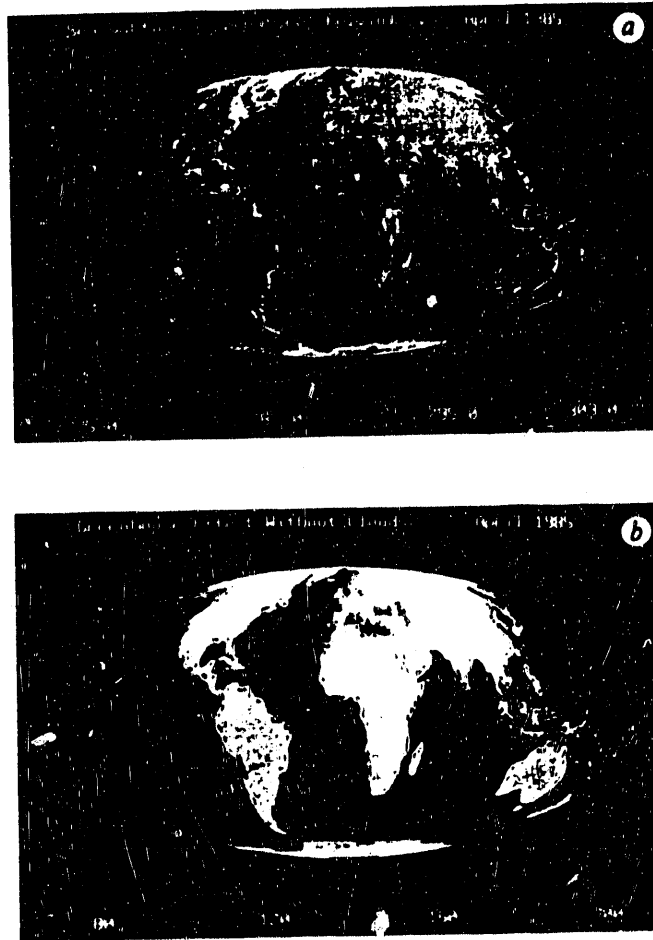


Fig. 1. a: Monthly mean map of observed sea surface temperature for April 1985. Values are in degrees Kelvin. 1b: Same as Fig. 1a but for the greenhouse effect (G) of atmosphere under clear skies. Values are in Watts/m². The major inference from these figures is that the greenhouse effect of the atmosphere is strongly correlated with surface temperature. This correlation can explain the equator to pole decrease in G as well as the east to west variations. For example, G is maximum over the relatively warm western equatorial Pacific Ocean (Indonesian region) and is very low over the cold oceanic upwelling regions in the central and eastern equatorial Pacific and off the west coasts of Africa and South America.

EVIDENCE OF H₂O FEEDBACK

The clear-sky G (Fig. 1b) generally decreases from equator to pole with some exceptions. The trapping due to clouds (not shown here) does not vary systematically with latitude, but peaks in the convectively disturbed regions of low latitudes and in the vicinity of the mid-latitude jet stream regions. We find that the clear-sky G has a strong positive correlation with sea-

surface temperature (compare Fig. 1a with Fig. 1b). Our thesis is that this correlation gives direct evidence of the H₂O feedback.

We show this in three steps. First, we show the correlation of G with T_s, implies that the longwave absorptivity of the atmosphere increases with surface temperature. Second, the variation of water vapor in the atmosphere is determined largely by the increase in the saturation vapor pressure with temperature, as shown by other observational (2,9) and modeling (10) studies. Third, if the observed variation of water vapor causes the sort of G-T_s coupling revealed by the data, the atmospheric absorptivity must scale logarithmically with the water content. This is not only indicated from observations but is also consistent with theories of radiative transfer.

Step 1. Since the surface emission increases as the fourth power of T_s, the trapping should also increase with T_s. In order to remove the T_s⁴ dependence, we define a normalized trapping $g = G/\sigma T_s^4$. Both the clear-sky g (Fig. 2) and the cloudy g (not shown) increase non-linearly at warm temperatures (T_s > 298 K), but otherwise g increases smoothly with T_s. The grid point values of g for the individual months and the two hemispheres reveal similar dependencies on T_s. Because of the robustness of this feature, we conclude that g and the surface temperature are strongly coupled.

The increase in g is primarily due to the increase in the absorptivity (A) with T_s. In the troposphere the lapse rate $\Gamma = -dT/dz$ is approximately constant with pressure. We can then express the atmospheric temperature as: (T/T_s) = x^α, where α = ΓR/g_s. Inserting this in Eq. (2),

$$g = G/\sigma T_s^4 = 4\alpha \int_0^1 A(x)x^\beta dx \quad (3)$$

where β = (4α - 1). We first see that g does not explicitly depend on T_s. The g-T_s coupling has to come about because either the lapse rate (and hence α) or A depend on T_s. The observed change in lapse rate (11) over the range of T_s considered here can affect g by at most 10% (based on flux sensitivities to lapse rates given in Coakley (12)). The only other variable is the atmospheric absorptivity A. For a fixed amount of H₂O, A varies by only 5% (2) for the range of T_s considered here. The temperature dependence of A also cannot explain the rate of increase of g with T_s. Therefore, the amount of water vapor must increase with surface temperature.

Step 2. The increase in water vapor with T_s, which in turn causes an increase in A, is sufficient to explain the g-T_s coupling. The saturation vapor pressure (e_s) of H₂O is given by the Clausius-Clapeyron equation $e_s = C \exp(-(LR^{-1})/T)$. Here, C is a constant, L is the latent heat of vaporization, and R is the gas constant for water vapor; LR⁻¹, at the global mean temperature, is on the order of ~5400 K. Using the observed global mean lapse rate and relative humidity profile (2) we numerically compute the total water content, W, in a vertical column in the troposphere for a range of values of T_s. W increases exponentially with a rate of increase d(ln W)/dT_s = 6.7 x 10⁻² K⁻¹. Compare this to the actual rate of increase (5.5 x 10⁻²

K^{-1}), obtained by satellite microwave observations (13) of W (Fig. 3a). They agree to within 20%; hence, W is largely determined by the surface temperature, in agreement with the findings of Prabhakara et al (9) and Stephens (accepted, J. of Climate). This coupling between W and T_s arises from the temperature dependence of the saturation vapor pressure.

Step 3. Since we are ignoring variations in the lapse rate or in spectroscopic parameters with temperature, we can write dg/dT_s as a product of a partial and a total derivative:

$$\frac{dg}{dT_s} = \frac{\partial g}{\partial \ln W} \cdot \frac{d \ln W}{dT_s} \quad (4)$$

The first term is the observed coupling of the normalized greenhouse effect with T_s . The last term represents the fundamental source of the H_2O feedback - the water content increases with T_s . The second term represents the 'return loop' in the feedback - the absorptivity increases with water content. We infer from Eq. (4) that if g varies linearly with T_s (which it does for $T_s < 298$ K in Fig. 2), and if $\ln W$ varies linearly with T_s (as suggested from theory and from the satellite microwave data in Fig. 3a), then g should increase logarithmically with W . In fact, theoretical calculations (2) (also see Fig. 9.23 of Goody (14)) have shown that, for the range of W considered in Fig. 3, the absorptivity scales logarithmically with the absorber amount. A plot of g versus $\ln W$ (Fig. 3b) shows this scaling. Numerically also the analysis is consistent, since dg/dT_s (from Fig. 2) is $3.4 \times 10^{-3} K^{-1}$, on a global average, while the product of $\partial g/\partial(\ln W)$ (from Fig. 3b) and $d(\ln W)/dT_s$ (from Fig. 3a) is $3.2 \times 10^{-3} K^{-1}$. This comparison may be used in an illustrative sense only, since the microwave soundings of W are only available for the period 1979-83, while the ERBE measurements are not available prior to 1985. However, a comparison of zonal averages (as shown in the g vs. $\ln W$ plot in Fig. 3b) minimizes the variation across these two time periods.

The analysis strongly indicates the presence of the H_2O feedback. Three further points can be made in favor of this interpretation. First, the clear-sky greenhouse effect and its temperature dependence is consistent with state-of-the-art radiative transfer calculations (15) (Fig. 4). These calculations employ observed humidity and temperature profiles for each climate type (sub-polar winter, tropical summer, etc.). Hence, their agreement with the observed G indicates that the 'return loop' in the H_2O feedback, $\partial g/\partial(\ln W)$, behaves as expected from theory.

Second, the sensitivity dG/dT_s of the NCAR Community Climate Model (CCM) (16) agrees closely with the observed sensitivity (Fig 4). The CCM is a 3-dimensional general circulation model. The model ignores the radiative effects of trace gases such as CH_4 and N_2O (which contribute about 5 to 7 W/m^2 to the clear-sky G) and is also drier than the observed climate. (Kiehl and Ramanathan, accepted J. Geophys. Res. 1989.) Hence it systematically underestimates G by 10 to 15 W/m^2 . It has, however, a latitudinal sensitivity dG/dT_s of $3.1 W m^{-2} K^{-1}$, which compares well with the value of $3.3 W m^{-2} K^{-1}$

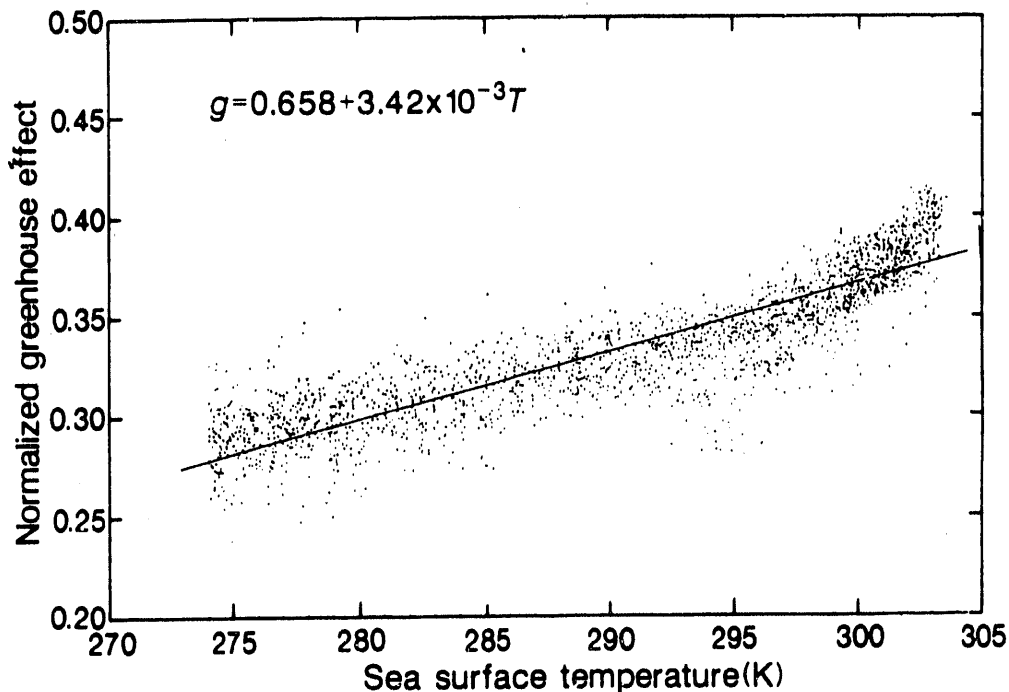


Fig. 2. Clear sky greenhouse effect normalized by IR surface emission (g) vs. surface temperature, April 1985. Values are monthly averages for each $2.5^\circ \times 2.5^\circ$ region. The solid line indicates a least-squares fit through all the points. Fits done individually through Northern and Southern hemisphere points do not deviate by more than 3%.

obtained from ERBE, averaged annually and over the whole range of sea surface temperatures.

Finally, the General Circulation Model (GCM) designed by S. Manabe at the Geo-physical Fluid Dynamics Laboratory shows a climate sensitivity consistent with the analysis of ERBE flux observations. Wetherald and Manabe calculated (17) that a 2% increase in the solar constant increased the GCM's outgoing longwave flux by 6.3 W/m^2 , while the surface temperature increased by 3.0 K . Thus the longwave trapping G of the GCM has a sensitivity to the global mean temperature of $dG/dT_s = 4\alpha T_s^3 = dF/dT_s$, or $3.7 \text{ W m}^{-2}\text{K}^{-1}$. A 2% decrease in the solar constant results in a higher sensitivity, of $4.0 \text{ W m}^{-2}\text{K}^{-1}$. The cloud cover is held constant in this model, hence the cloud-cover feedback is zero. The observed sensitivity is within 15% of the GCM values. This implies that the climatological sensitivity of the GFDL GCM is in good agreement with the latitudinal sensitivity of the present atmosphere. Such behavior is necessary if the change in the equilibrium climate is determined largely by the H_2O -greenhouse feedback.

The analysis up to this point is very encouraging, in that independent climate data sets show an overall consistency that can be explained in terms

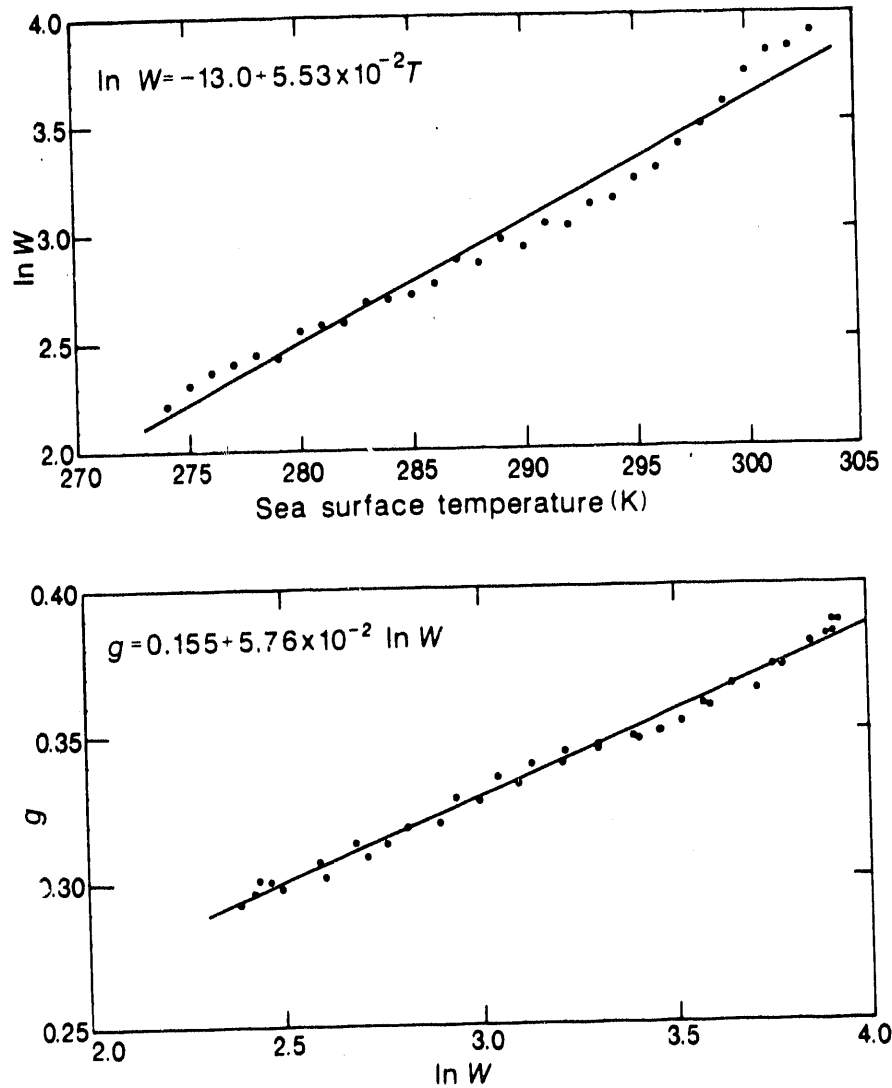


Fig. 3. a: Natural logarithm of the atmospheric water content (W) vs. surface temperature, April 1982, monthly averages for $3^\circ \times 5^\circ$ regions. Values are in $\ln \text{ kg/m}^2$. The water content was derived from microwave satellite soundings (13) available for 1979-1983. The error in W is about 10% (13). The strong positive correlation indicates that the behavior of W follows simple thermodynamic laws; the slight upward and then downward deviations are consistent with the latitudinal variations in relative humidity (2) and lapse rate (11) which are governed by the dynamics of the atmosphere. b: Normalized clear sky greenhouse effect (g) for April 1985 vs. W for April 1982. ERBE data is not available prior to 1985 while W is not available after 1983. In order to minimize the effect of year to year variations in W , we have employed zonally averaged values.

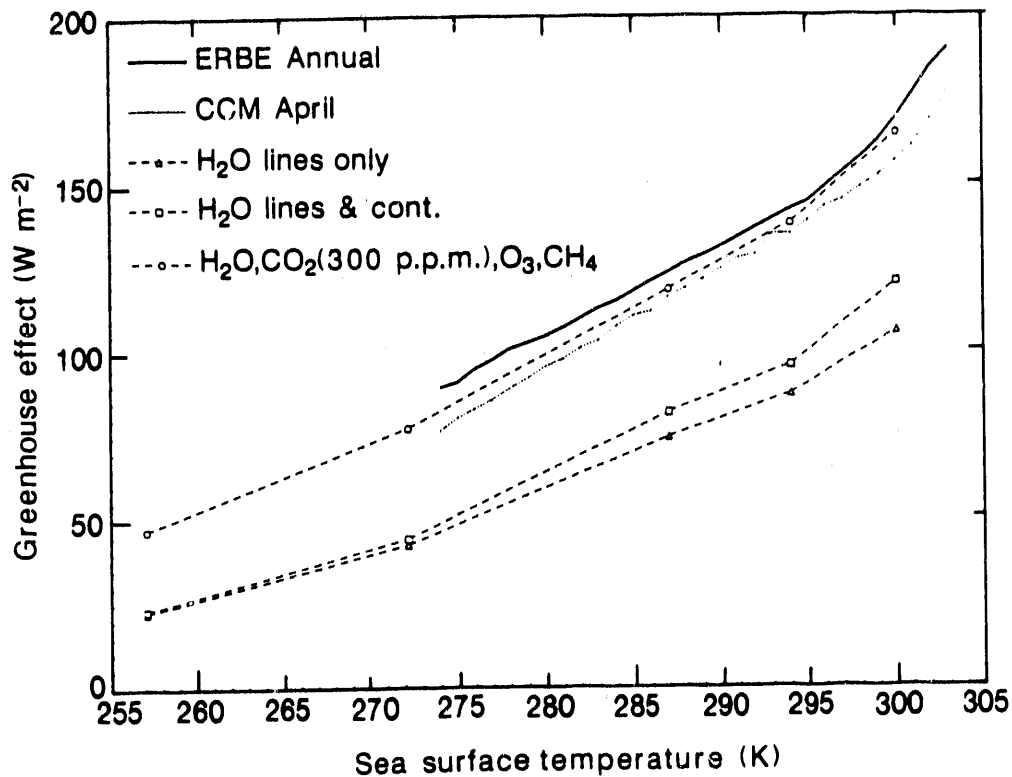


Fig. 4. Comparison of G vs. surface temperature, retrieved from three sources: bold line, ERBE annual values, obtained by averaging April, July and October 1985 and January 1986; thin line, three-dimensional climate model simulations for a perpetual April (NCAR-CCM); dashed lines, line-by-line radiation model calculations by Dr. A. Arking (15) for individual H_2O spectral lines, as well as the continuum absorption. The five points employ climatological winter and summer atmospheric profiles for sub-arctic and mid-latitude conditions, and a tropical profile. A comparison of the computed G with and without the continuum absorption (bottom two dashed lines) indicates a non-linear contribution from the continuum. This may in part account for the non-linear rise at the highest temperatures in the ERBE derived G . When CO_2 , O_3 , and CH_4 are added the results come close to ERBE values.

of thermodynamics and radiation physics; this is perhaps one of the best indicators of the accuracy of the ERBE data. We find compelling evidence that the H_2O feedback is detectable from ERBE observations, and that it can account for most (at least 80%) of the observed variation of G with surface temperature. Other atmospheric processes, such as the variation of relative humidity or lapse rate with latitude, are needed to explain the remaining variation.

IMPLICATIONS TO THE GLOBAL WARMING PROBLEM

We conclude that the coupling between surface temperature and the greenhouse effect will lead to a strong positive feedback on any perturbations to the present climate. One significant implication is that we might detect the greenhouse effect of human activities. The rise in trace gases, if it continues unabated, will directly increase G by about 0.5 W/m^2 per decade. The cumulative increase in G for the next 50 years will then be about 2.5 W/m^2 . In the same period, the globe is expected to warm by 2 to 4 K (8). Because of the coupling between T_s and g , the global G will actually increase by an additional 6.5 to 13 W/m^2 , leading to a total increase of 9 to 15.5 W/m^2 . Such an increase should be clearly detectable if we continue ERBE type observations into the next century. If the tropical regions exhibit the non-linear increase revealed by the ERBE data (Fig. 2), the effect should be most prominent in the tropics.

Both the cloudy and clear sky greenhouse effect increases rapidly at temperatures characteristic of the tropics. It is at present not known whether this non-linear rise is simply a climatological feature of low latitudes, unrelated to temperature variations. For example, the relatively dry regions found in the subtropics ($290 < T_s < 298 \text{ K}$) may slightly suppress the rate of increase of g with T_s , and hence change the slope.

The non-linearity might, on the other hand, be due to the non-linear dependence of the H_2O continuum opacity (15) or a temperature threshold for deep convection (18). Since deep convection is a source for the ice content of cirrus clouds, the greenhouse effect of cirrus clouds can increase with T_s when T_s exceeds a certain threshold value. If so, the magnitude of the positive feedback is so large that we should look for a compensating negative feedback elsewhere in the system to explain the relative stability of the tropical climate during the last great ice-age.

ACKNOWLEDGMENTS

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INTERPRETATION OF CLOUD-CLIMATE FEEDBACK AS PRODUCED BY
14 ATMOSPHERIC GENERAL CIRCULATION MODELS*

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Understanding the cause of differences among general circulation model projections of carbon dioxide-induced climatic change is a necessary step toward improving the models. An intercomparison of 14 atmospheric general circulation models, for which sea surface temperature perturbations were used as a surrogate climate change, showed that there was a roughly threefold variation in global climate sensitivity. Most of this variation is attributable to differences in the models' depictions of cloud-climate feedback, a result that emphasizes the need for improvements in the treatment of clouds in these models if they are ultimately to be used as climatic predictors.

Observed and projected increases in the concentration of atmospheric CO₂ and other greenhouse gases have stimulated considerable interest in modeling climatic change. The most detailed climate models for this purpose are three-dimensional general circulation models (GCMs). Although most GCMs are of similar design, there are significant differences among GCM projections of climatic warming as induced by increasing levels of atmospheric carbon dioxide (1,2). The reasons for these differences are not fully understood, but variations in how cloud-climate feedback processes are simulated in the various models are thought to be largely responsible (3); cloud feedback is dependent on all aspects of a model and not just on cloud formation parameterizations. Clearly there is a need to isolate and to understand better cloud feedback mechanisms in GCMs, and, more specifically, to determine if they are a significant cause of intermodel differences in recent climate-

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change projections. Consequently we have made an intercomparison of cloud feedback in 14 atmospheric GCMs as part of a larger study directed toward improving GCMs and climatic projections.

Many facets of the climate system are not well understood, and thus the uncertainties in modeling atmospheric, cryospheric, and oceanic interactions are large. In evaluating the differences among models, we have focused first on atmospheric processes, because these uncertainties must be understood before others can be addressed. For simplicity, we have emphasized solely global-average quantities, and we adopted the conventional interpretation of climate change as a two-stage process: forcing and response (4). The concept of global-average forcing and response has proven useful in earlier interpretations of cloud-climate feedback. For example, by performing two GCM simulations for a doubling of atmospheric CO_2 concentration, one with computed clouds and the other with clouds that were invariant to the change in climate, Wetherald and Manabe have suggested (5) that cloud-climate feedback amplifies global warming by the factor 1.3. A somewhat larger amplification (1.8) was estimated by Hansen et al. (6) using a one-dimensional climate model to evaluate climate feedback mechanisms in a different GCM.

The global-mean direct radiative forcing, G , of the surface-atmosphere system is evaluated by holding all other climate parameters fixed. It is this quantity that induces the ensuing climate change, and physically it represents a change in the net (solar plus infrared) radiative flux at the top of the atmosphere (TOA). For an increase in the CO_2 concentration of the atmosphere, G is the reduction in the emitted TOA infrared flux resulting solely from the CO_2 increase, and this reduction results in a heating of the surface-atmosphere system. The response process is the change in climate that is then necessary to restore the TOA radiation balance, such that

$$G = \Delta F - \Delta Q \quad (1)$$

where F and Q respectively denote the global-mean emitted infrared and net downward solar fluxes at the TOA. Thus ΔF and ΔQ represent the climate-change TOA responses to the direct radiative forcing G , and these are the quantities that are impacted by climate feedback mechanisms. Furthermore, it readily follows that the change in surface climate, expressed as the change in global-mean surface temperature ΔT_s , is related to the direct radiative forcing G by

$$\Delta T_s = \lambda G \quad (2)$$

where λ is the climate sensitivity parameter

$$\lambda = \frac{1}{\Delta F / \Delta T_s - \Delta Q / \Delta T_s} \quad (3)$$

An increase in λ thus represents an increased climate change due to a given climate forcing G .

A simple example illustrates the use of λ for evaluating feedback mechanisms. If only the basic temperature-radiation negative feedback exists, then climate change refers solely to temperature change, and there are no related changes in atmospheric composition, lapse rate (vertical temperature gradient), or surface albedo (reflectance). Thus $\Delta Q/\Delta T_s = 0$, and $\Delta F/\Delta T_s$ is evaluated with the assumption that $F = \epsilon \sigma T_s^4$ (7), where σ is the Stefan-Boltzmann constant and ϵ is the emissivity of the surface-atmosphere system, which is constant in this case. It then follows that $\Delta F/\Delta T_s = 4F/T_s = 3.3 \text{ W m}^{-2} \text{ K}^{-1}$ for conditions typical of Earth, so that in the absence of interactive feedback mechanisms, $\lambda = 0.3 \text{ K m}^2 \text{ W}^{-1}$.

A well-known positive feedback mechanism is water-vapor feedback (8), in which a warmer atmosphere contains more water vapor, which as a greenhouse gas amplifies the initial warming. Climate models that contain this positive feedback process typically give $\Delta F/\Delta T_s \approx 2.2 \text{ W m}^{-2} \text{ K}^{-1}$. In addition, the increased water vapor increases the atmospheric absorption of solar radiation, and for a typical model this positive feedback yields $\Delta Q/\Delta T_s \approx 0.2 \text{ W m}^{-2} \text{ K}^{-1}$. Thus, with the inclusion of water-vapor feedback λ is increased from 0.3 to 0.5 $\text{K m}^2 \text{ W}^{-1}$.

Whereas water-vapor feedback is intuitively straightforward to understand, cloud feedback is a far more complex phenomenon. There are several ways that clouds can produce feedback mechanisms. For example, if global cloud amount decreases because of climate warming, as occurred in simulations with the 14 GCMs we employed, then this decrease reduces the infrared greenhouse effect attributed to clouds. Thus as Earth warms it is able to emit infrared radiation more efficiently, moderating the global warming and so acting as a negative climate feedback mechanism. But there is a related positive feedback: the solar radiation absorbed by the surface-atmosphere system increases because the diminished cloud amount causes a reduction of reflected solar radiation by the atmosphere. The situation is further complicated by climate-induced changes in both cloud vertical structure and cloud optical properties, which result in additional infrared and solar feedbacks (2).

In our intercomparison, cloud effects were isolated by separately averaging a model's clear-sky TOA fluxes (2,9), such that in addition to evaluating climate sensitivity for the globe as a whole, we were also able to consider an equivalent "clear-sky" Earth. In other words, a model's clear-sky TOA infrared and solar fluxes were separately stored during integration and then globally averaged by use of conventional latitudinal area weighting. When used in conjunction with Eq. 3, a single model integration thus provided not only the global climate sensitivity parameter but also a second sensitivity parameter that refers to a clear-sky Earth with the same climate as that with clouds present. In effect, we processed GCM output in a manner similar to the way in which data is processed in the Earth Radiation Budget Experiment (10), an experiment that also produces an equivalent clear-sky Earth.

Our choice of a model intercomparison simulation was governed by several factors. Ideally the climate simulation should refer to a relevant situation, such as increasing the atmospheric CO_2 concentration. Only 3 of the

14 models, however, have been employed for this purpose. Furthermore, these three models have, at least, in part, differing climate sensitivities because their control (that is, present-day) climates are different (2,3). If a model produces a control climate that is either too warm or too cold, then it will respectively produce a climate sensitivity parameter that is too small or too large, and clearly the intercomparison simulation had to be designed to eliminate this effect. There was also a practical constraint: the CO₂ simulations require large amounts of computer time for equilibration of the rather primitive ocean models that have been used in these numerical experiments.

As an alternative that eliminated both of the above-mentioned difficulties, we adopted ± 2 K sea surface temperature (SST) perturbations, in conjunction with a perpetual July simulation, as a surrogate climate change for the sole purpose of intercomparing climate sensitivity (2). This procedure is in essence an inverse climate change simulation. Rather than introducing a forcing G into the models and then letting the climate respond to this forcing, we instead prescribed the climate change and let the models in turn produce their respective forcings in accordance with Eq. 1. This procedure eliminated the substantial computer time required for equilibration of the ocean. The second advantage was that because the same SSTs were prescribed (11), all of the models had essentially the same control climate because land temperatures are tightly coupled, through atmospheric transport, to the SSTs. The models then all produced a global-mean ΔT_e between the -2 K and +2 K SST perturbation simulations that was close to 4 K, and different model sensitivities in turn resulted in different values for G .

The perpetual July simulation eliminated another problem. Our study focused solely on atmospheric feedback mechanisms, and inspection of output from all the models showed that climate feedback caused by changes in snow and ice coverage was suppressed through use of a fixed sea ice constraint and because the perpetual July simulations produced little snow cover in the Northern Hemisphere. For this reason we adopted global averages rather than the 60°S to 60°N averages used in an earlier study (2).

Several of the 14 GCMs used in the intercomparison (designated by acronyms in Table 1) have common origins. The GFDL II model, relative to GFDL I, includes parameterization for cloud albedo as a function of cloud water content. The CCM0 and CCM1 are the standard versions (0 and 1) of the NCAR CCM, with version 1 containing a revised radiation code. The CCM/LLNL GCM is CCM1 with a further solar radiation code revision and the incorporation of cloud albedos as a function of cloud water content. The OSU/IAP and OSU/LLNL GCMs are two-level models that contain modifications to the standard Oregon State University GCM. Both the numerical technique and the convective adjustment parameterization were revised in the OSU/IAP model, whereas the solar radiation code was revised in the OSU/LLNL GCM. The ECMWF GCM, relative to ECMWF/UH, has a revised radiation code and a smaller (factor of 2) horizontal resolution.

Table 1. Summary of the GCMs used in the intercomparison (13). There are two GFDL models. NCAR, National Center for Atmospheric Research

Model	Investigators
Canadian Climate Center (CCC)	Boer and Blanchet
Colorado State University (CSU)	Randall
European Center for Medium-Range Weather Forecasts (ECMWF)	Morcrette
European Center for Medium-Range Weather Forecasts - Univ. of Hamburg (ECMWF/UH)	Roeckner and Schese
Geophysical Fluid Dynamics Laboratory (GFDL I and II)	Wetherald
Laboratoire de Meteorologie Dynamique (LMD)	Le Treut and Li
Meteorological Research Institute of Japan (MRI)	Yagai
NCAR Community Climate Model, version 0 (CCMO)	Washington
NCAR Community Climate Model, version 1 (CCM1)	Slingo and Kiehl
NCAR Community Climate Model-Lawrence Livermore National Laboratory (CCM/LLNL)	Ghan and Taylor
Oregon State University-Institute for Atmospheric Physics, Beijing (OSU/IAP)	Liang
Oregon State University-Lawrence Livermore National Laboratory (OSU/LLNL)	Cess and Potter
United Kingdom Meteorological Office (UKMO)	Mitchell

All of the models treat two cloud types: stratiform (large-scale) and convective clouds. Except in the ECMWF and ECMWF/UH models, stratiform clouds are formed in a vertical atmospheric layer when the relative humidity exceeds a prescribed threshold value, which varies among models for 90 to 100%. The models then either prescribe the cloud cover in their respective grid areas, which vary in size from 2.8° by 2.8° to 5° by 7.5° in latitude by longitude, or calculate it as a function of relative humidity. In the ECMWF and ECMWF/UH GCMs, vertical velocity and lapse rate are also used as cloud predictors.

The procedure for convective clouds is far less consistent. The CCC, the two GFDL, and the three CCM GCMs generate convective clouds in the same way as they generate stratiform clouds. However, the fraction of the grid area that is covered by convective cloud varies from 30 to 100% among these models. In the remaining models a parameterization is used that relates the convective cloud fraction to the convective precipitation rate.

In the intercomparison of climate sensitivity parameters, there was a nearly threefold variation in the global sensitivity parameter (Table 2), but excellent agreement in the clear sensitivity parameter. These clear values are also consistent with our conventional interpretation of water-vapor feedback as discussed above. These results suggest that the

Table 2. Summary of climate sensitivity parameters for the perpetual July simulations; λ_c is the clear-sky sensitivity parameter.

Model	λ (K m ² W ⁻¹)	λ_c (K m ² W ⁻¹)	λ/λ_c
CCC	0.39	0.42	0.93
ECMWF	0.40	0.57	0.70
GFDL II	0.45	0.46	0.98
CSU	0.50	0.46	1.09
OSU/LLNL	0.52	0.48	1.08
MRI	0.60	0.47	1.28
GFDL I	0.60	0.48	1.25
UKMO	0.61	0.53	1.15
CCM1	0.70	0.43	1.63
CCM/LLNL	0.76	0.49	1.55
LMD	0.90	0.42	2.14
OSU/IAP	0.90	0.44	2.05
ECMWF/UH	1.11	0.47	2.36
CCMO	1.11	0.45	2.47
Mean	0.68	0.47	
SD	0.24	0.04	

substantial disagreements in global sensitivity can largely be attributed to differences in cloud feedback. Understanding this point requires definitions of cloud feedback and cloud-radiative forcing. Cloud feedback has been discussed for roughly two decades, but there is considerable uncertainty as to its meaning; it has often been confused with cloud-radiative forcing, whereas it is actually related to a change in cloud-radiative forcing.

Cloud-radiative forcing refers to the radiative impact of clouds on the earth's radiation budget as determined at the TOA. Denoting this impact as CRF, and letting the subscript c refer to clear-sky fluxes, then,

$$\text{CRF} = F_c - F + Q - Q_c \quad (4)$$

In this definition CRF is positive when clouds produce a warming of the surface-atmosphere system. Combination of Eqs. 1, 2, 3, and 4 then yields

$$\lambda/\lambda_c = 1 + \Delta\text{CRF}/G \quad (5)$$

where ΔCRF is the change in cloud-radiative forcing as induced by the change in climate and λ_c is the clear-sky climate sensitivity parameter (Table 2).

Conceptually cloud feedback should be related to a change in cloud-radiative forcing, as illustrated in Eq. 5. In the absence of cloud feedback (that is, $\Delta\text{CRF} = 0$), the global sensitivity parameter equals that for clear skies. In turn, a departure of λ/λ_c from unity is a measure of cloud

feedback, and a $\lambda/\lambda_c > 1$ denotes a positive feedback. Cloud-radiative forcing for Earth's present climate is a measurable quantity; the Earth Radiation Budget Experiment (ERBE) is currently producing this information.

Equation 5 provides a convenient means of understanding why cloud feedback is the primary cause of the intermodel variations in global climate sensitivity. A scatter plot of λ versus the cloud feedback parameter $\Delta CRF/G$ for the 14 GCMs (Fig. 1) clearly shows that the intermodel differences in global climate sensitivity are dominated by their corresponding differences in $\Delta CRF/G$: the points scatter about a regression line that is consistent with Eq. 5. The scatter results from the relatively minor intermodel differences in the clear sensitivity parameter. This analysis thus supports the suggestion that cloud-climate feedback is a significant cause of intermodel differences in climate change projections.

The GFDL I and II models provide a direct means of appraising a specific cloud feedback component attributed to cloud optical properties. In GFDL II the cloud albedos are dependent on cloud water content, whereas in GFDL I these albedos are prescribed. Because cloud water content should, on average, increase as the climate warms, producing a related increase in cloud albedos, GFDL II should have, relative to GFDL I, a negative cloud feedback component (12). The global sensitivity parameter for GFDL II is 25% less than that for GFDL I (Table 2), consistent with this expectation.

A similarly straightforward argument does not, however, apply to the CCM1 versus CCM/LLNL models, for which the latter also incorporates cloud albedos that are dependent on cloud water content. An inspection of the output of these two GCMs shows, like the GFDL comparison, that CCM/LLNL contains, relative to CCM1, a negative solar cloud feedback component. But unlike the case for GFDL I and II, this negative feedback is compensated for by a positive cloud-amount feedback. The net result is that the two models produced nearly identical cloud feedback, as shown by their similar λ/λ_c values (Table 2). Nor is it possible to segregate the 14 GCMs into low and high-sensitivity groups on the basis of whether they do or do not incorporate cloud optical properties that depend upon cloud water content. The ECMWF and ECMWF/UH GCMs also incorporate this effect, and they lie at opposite ends of the cloud feedback spectrum (Table 2). Furthermore, even though the CSU and OSU/LLNL GCMs produced nearly identical modest positive cloud feedback (Table 2), this was actually a result of compensation between vastly different cloud feedback components.

In summary, although the 14 atmospheric GCMs produced comparable clear-sky sensitivity parameters, when cloud feedback was included, compatibility vanished and there was a nearly threefold variation in climate sensitivity as produced by the models. The cloud feedback ranged from modest negative to strong positive feedback. Clearly improvements in the treatment of clouds in GCMs are needed. But there are many other facets of a GCM, in addition to cloud optical properties and cloud formation parameterizations, that can influence cloud-climate interactions. The hydrological cycle, to cite one example, will most certainly play a dominant role.

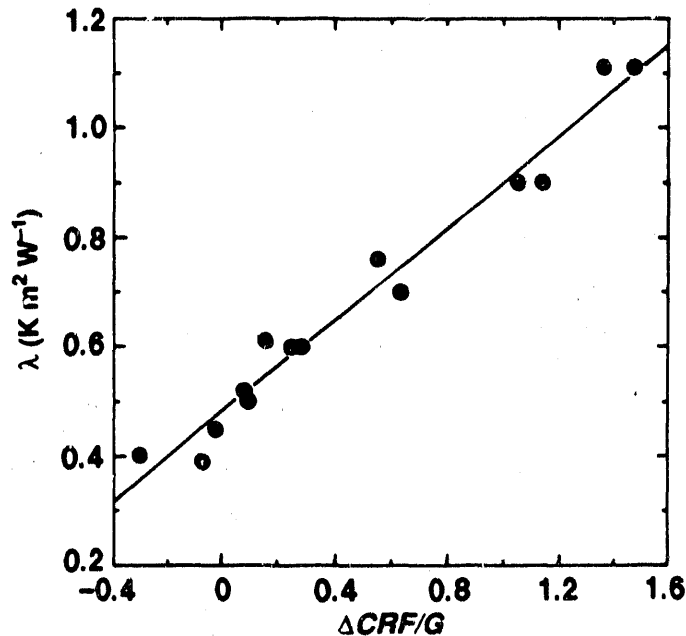


Fig. 1. The global sensitivity parameter λ plotted against the cloud feedback parameter $\Delta CRF/G$ for the 14 GCM simulations. The solid line represents a best-fit linear regression.

Many of these GCMs are in a continual state of evolution. Thus this intercomparison is a snapshot that might no longer represent a specific model. Furthermore, these model-produced cloud feedbacks may not be representative of how the models would behave under realistic climate change conditions when they are coupled with interactive cryosphere and ocean models. Perpetual July simulations cannot be used for this purpose. Nor can the uniform SST perturbations, because they do not account for changes in equator-to-pole temperature gradients associated with actual climate change. For example, it has recently been speculated (10) that this latter effect, by itself, may produce a cloud feedback component resulting from latitudinal shifts in general circulation patterns. But these caveats do not alter our conclusion that 14 different GCMs produced a broad spectrum of cloud-climate feedback.

Climate research benefits from a diversity of climate models. If only a limited number of models were available, we could not confidently conclude that the role of cloud feedback is a key issue for climate studies. Before this study, only two GCMs had been used to provide estimates of cloud feedback (5,6), and these two estimates showed much closer agreement than we have demonstrated.

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13. A brief description of the 14 GCMs will be provided in Cess et al. (in preparation); descriptions of individual models are available from the respective investigators.
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CLOUD PARAMETERIZATION

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INTRODUCTION

The current generation of cloud parameterizations for climate models is successful enough to produce results that can be profitably compared with observations, yet deficient enough to keep several future generations of model developers busy making improvements. The purpose of this paper is to review current parameterizations and their results.

One approach to cloud parameterization, and to parameterization in general, is to identify intuitively plausible relationships between the unknowns, such as cloud amount, and the known variables of the problem, such as relative humidity, static stability, and vertical velocity. This "inductive" approach was pioneered by Smagorinsky (1960), and has recently been pursued by J. M. Slingo (1980; 1987). Its strength is that it can quickly yield parameterizations that are undeniably useful, e.g., for numerical weather prediction. A weakness is that inductive parameterizations lack theoretical underpinnings that could indicate their limits of applicability.

The alternative, "deductive" approach is based on the philosophical view that a parameterization should be a condensed physical representation of the process in question. The limits of applicability of such a parameterization can be inferred, *a priori*, from its physical basis, which should consist of observationally testable assumptions. This discussion emphasizes deductive parameterizations.

THE PHYSICAL BASIS OF CURRENT PARAMETERIZATIONS

Existing climate models typically include three main types of clouds:

- "Large-scale saturation clouds" that occur when the relative humidity exceeds some threshold value. The high relative humidity can be due to many causes, including large-scale lifting and surface evaporation. There are currently no physically based methods for determining the supersaturation cloud amount, so for simplicity some models allow only zeroes and ones, i.e., no cloud or complete overcast. However, cloud fractions are arbitrarily prescribed in many models. When fractional supersaturation cloud amounts occur in more than one layer, random overlap is often assumed for radiative transfer purposes. Precipitation falling from supersaturation clouds often evaporates into subsaturated layers encountered on the way down.
- "Convective clouds" are associated in some way with the cumulus parameterization of the model, perhaps in terms of the convective precipitation rate or the convective mass flux. In some cases, these clouds are supposed to represent the convective towers themselves, while in others they represent the stratiform anvil and/or cirrus clouds that often form near the tops of convective layers. We currently lack a physically based method to determine the fractional cloudiness in convective layers.
- "Boundary-layer clouds" are separately parameterized in some models (Suarez et al., 1983; Randall et al., 1985). These clouds are produced by surface evaporation into a boundary layer whose depth is limited by a strong inversion and large-scale subsidence (Lilly, 1968). Boundary-layer clouds occur in the marine subtropics, in midlatitude cold outbreaks, and in many other situations. Both stratocumulus and shallow cumulus clouds can be included in this category, although existing parameterizations deal primarily with the stratiform types.

SOME RESULTS FROM A CURRENT GCM

In this section, I discuss the cloud parameterizations used in the Colorado State University (CSU) General Circulation Model (GCM). This GCM has been derived from the GCM of the University of California at Los Angeles (UCLA), which was developed at UCLA, over a period of 20 years, by A. Arakawa and collaborators. Many changes have been made since the model left UCLA. The most important of these are revised solar and terrestrial radiation parameterizations (Harshvardhan et al., 1987), and a new parameterization of land-surface processes (Sellers et al., 1986).

The prognostic variables of the model are: potential temperature, the mixing ratios of water vapor and ozone, the horizontal wind components, the surface pressure, the planetary boundary layer's depth and turbulence kinetic energy, the ground temperature, ground wetness, and snow depth at land points, and the ice temperature at land ice and sea ice points. The governing equations are finite-differenced, using highly conservative schemes

(Arakawa and Lamb, 1977; 1981). Fourier filtering of the mass flux and pressure gradient vectors is used to maintain computational stability near the poles (Arakawa and Lamb, 1977). The model has been programmed so that the horizontal and vertical resolutions and the pressure of the model top are easily varied.

A key feature of the GCM is its formulation in terms of a modified sigma coordinate, in which the top of the planetary boundary layer (PBL) is a coordinate surface, and the PBL itself is the lowest model layer (Suarez et al., 1983). As already mentioned, the depth of the PBL is a prognostic variable of the model. The turbulence kinetic energy of the PBL has recently been introduced as a new prognostic variable. This simplifies the entrainment calculation and prepares the way for further improvements to the PBL parameterization. The mass sources and sinks for the PBL consist of large-scale convergence or divergence, turbulent entrainment, and the cumulus mass flux. Turbulent entrainment can be driven by positive buoyancy fluxes, or by shear of the mean wind in the surface layer or at the top of the PBL.

The cumulus mass flux and the warming and drying of the free atmosphere due to cumulus convection are determined through the cumulus parameterization of Arakawa and Schubert (1974) and Lord et al. (1982). The ice phase is taken into account in the cumulus parameterization, but not in the large-scale saturation parameterization. Rain falling through unsaturated grid boxes is allowed to evaporate if it originates through large-scale saturation at an upper level, but not if it originates in cumulus towers. Cumulus friction is included.

The Arakawa-Schubert parameterization applies only to clouds that draw their mass from the PBL. There are moist convective motions that originate above the PBL; these are parameterized through a conventional moist convective adjustment.

The radiation parameterization of the model was developed by Harshvardhan et al. (1987). The terrestrial radiation parameterization includes cooling due to water vapor, carbon dioxide, and ozone. The solar radiation parameterization includes Rayleigh scattering and absorption by water vapor and ozone, and simulates both the diurnal and seasonal cycles. A complete (solar and terrestrial) radiation calculation is done once per simulated hour, to resolve adequately the diurnal cycle and the effects of transient cloudiness (Wilson and Mitchell, 1986). For the tropospheric simulations described in this paper, a zonally uniform ozone distribution was prescribed as a function of latitude and height from the observations of McPeters et al. (1984).

Cloudiness can occur in any GCM layer, and can be associated with large-scale saturation, PBL stratocumulus clouds, or the anvils of deep cumuli. When and where cloudiness occurs, it is assumed to fill an entire grid box, for simplicity; no parameterization of subgrid fractional cloudiness is attempted, although work is under way to develop one (Randall, 1987). The optical properties assigned to the clouds are described by Harshvardhan et al. (1989).

Stratocumulus clouds are present in the PBL whenever the temperature and mixing ratio at the top of the PBL (as determined by a mixed-layer assumption) correspond to supersaturation, provided that cloud-top entrainment stability does not occur. The presence of the stratocumulus clouds is felt through both the radiation and entrainment parameterizations. The latter takes into account the generation of turbulence kinetic energy through increased buoyancy fluxes associated with phase changes and highly concentrated cloud-top radiative cooling (Randall 1980; 1984). As a result of these cloud-enhanced buoyancy fluxes, the presence of a stratocumulus layer in the PBL tends to favor more rapid entrainment and, therefore, a deeper PBL.

When cloud-top entrainment stability occurs, the PBL is assumed to exchange mass with the free atmosphere. The amount of mass exchanged is that required to remove the supersaturation, or to restore stability, whichever is less. The depth of the PBL remains unchanged during this process, which we refer to as "layer cloud instability" (LCI).

The prescribed boundary conditions of the GCM include realistic topography, and the observed climatological global distributions of sea-surface temperature, sea-ice thickness, surface albedo, and surface roughness, which vary seasonally, as well as morphological and physiological parameters for the land-surface vegetation, which also vary seasonally. The surface albedo of the ocean is zenith-angle dependent, following Briegleb et al. (1986), who used the data of Payne (1972).

The albedo of the vegetated land surface is determined according to the method of Sellers et al. (1986), which includes the effects of snow cover. The fraction of the ground covered by snow is not permitted to exceed 0.8. The albedo of sea ice is 0.8 in the visible, and 0.4 in the near infrared, except that when the temperature of the sea ice is within 0.05 K of the melting point, these values are replaced by 0.48 and 0.24, respectively. The albedo of land ice is 0.8 in the visible and 0.5 in the near infrared.

A new land-surface parameterization has recently been introduced, following the ideas of Sellers et al. (1986). The tendency of the terrestrial biosphere to limit the evapotranspiration are explicitly modeled through a simple vegetation parameterization, with 12 types of vegetation. This new parameterization has significantly reduced the simulated precipitation over warm land masses.

Simply by recompiling and setting logical switches, the GCM can be run as a one-dimensional model, or as a limited-area model on an f-plane with periodic boundary conditions, or as a shallow-water model. "Large-scale" pressure gradients, divergences, and advective tendencies can be imposed in the one-dimensional and limited-area models. The one-dimensional model is particularly useful for debugging new parameterizations, and for noninteractive sensitivity tests.

Three types of clouds are generated by the UCLA/GLA GCM: optically thick convective "anvil" clouds, which horizontally fill a grid column from 400

mb to the highest level reached by the convection; supersaturation clouds, which occur when the relative humidity equals or exceeds 100%, are assigned a cloud fraction of 1, and have optical properties that vary with temperature; and planetary boundary layer stratocumulus clouds, which can be arbitrarily thin, and occur only when there is a strong inversion at the top of the PBL. In this paper, I emphasize the effects of the upper-level supersaturation and anvil clouds that are associated with cumulus convection. Randall et al. (1985) analyzed the model's simulation of PBL stratocumulus clouds.

Naturally enough, the processes for parameterized cloud generation are closely coupled to certain aspects of the GCM's hydrologic cycle, especially the distribution of moisture itself and the intensity of moist convection. Surface evaporation is regulated by the model's PBL parameterization. So long as moisture remains inside the PBL, it can only form PBL stratus clouds, which cannot become any deeper than the PBL itself. To form either cumulus anvil clouds or large-scale saturation clouds in the free atmosphere, the moisture must find its way upward, out of the PBL. It cannot escape through vertical advection, because large-scale vertical motions lift the top of the PBL and the moisture together. The only ways that moisture can escape from the PBL into the free atmosphere are cumulus convection, "detrainment" as the PBL turbulence collapses near sunset over land, dry convective adjustment, layer cloud instability, and hole-filling that replaces negative mixing ratios by zeroes by borrowing from humid regions such as the PBL. The latter process is quite negligible in practice; cumulus convection is by far the most important mechanism for injecting moisture from the PBL into the free atmosphere. For example, in the model's tropics, high relative humidities near the tropopause are separated from the surface moisture supply by regions of low relative humidity in the middle troposphere. This can only be explained by penetrative moist convective moisture transport.

Upper level cloudiness in the model is always associated with one or more of three key parameterized processes. The first is penetrative cumulus convection that originates in the PBL. This is parameterized following Arakawa and Schubert (1974), as implemented by Lord et al. (1982). We refer to this parameterization as "CUP," for convenience. CUP represents the effects of cumulus clouds that originate at the top of the PBL. Although we expect this type of cumulus convection to be most important for driving the general circulation, it is also possible for cumulus updrafts to originate in the free atmosphere, above the PBL. For this reason, the GCM includes a supplementary parameterization for moist convection that originates above the PBL. This is parameterized in the GCM using Manabe's moist convective adjustment scheme; we call it "MSTADJ." MSTADJ is applied only for pairs of layers that both lie above the PBL; cumulus convection that originates the PBL is parameterized exclusively through CUP.

Simulated cloudiness in the free atmosphere is always associated with CUP, LSP, and MSTADJ. Cumulus convection that penetrates the 500 mb level is directly associated with anvil clouds, and indirectly tends to produce supersaturation clouds though detrainment moistening. LSP is always associated with supersaturation cloudiness.

In the time-averaged model results, CUP is only active over 9% of the Earth's surface, while MSTADJ is active over 17% of the Earth, and LSP over about 60%. Zonally averaged frequencies of occurrence for a simulated January are given as functions of latitude in Figure 1. The zonally averaged cumulus incidence reaches about 20% in the ITCZ, and has a secondary maximum over the northern continents in July. The cumulus anvil incidence has nearly the same profile as cumulus incidence, but is slightly less frequent at each latitude. Although cumulus incidence and MSTADJ incidence increase and decrease together, MSTADJ is consistently more prevalent, even slightly so in the ITCZ. LSP occurs frequently at all latitudes, but is least active (around 50% of the time) in the subtropics. It has a strong maximum in middle latitudes in winter. All three processes are quite vigorous in the winter storm tracks east of North America and Asia, although most of the precipitation that reaches the Earth's surface there is generated by LSP rather than CUP.

DIRECTIONS FOR THE FUTURE

Following the lead of Sundqvist (1978; 1981), several modeling groups are currently developing cloud parameterizations that employ prognostic variables for cloud species such as cloud liquid water, cloud ice, and cloud rainwater (Smith, 1990).

The utility of condensed water variables is obvious; condensates are the physical link between the latent heat effects and the radiative effects of clouds, so the consistent use of condensate variables helps to ensure physical consistency among the latent heating, radiation, and precipitation parameterizations. Although it is not so obvious that the condensate variables must be prognostic (one could envision neglecting the time-rate-of-change term, and solving a diagnostic or balance equation for the condensates), there are at least two possible motivations for determining the condensate distribution prognostically.

The first possible motivation is that a prognostic cloud water variable is demanded by the physics, in the sense that the current distribution of the condensate is influenced by its past history over the space and time scales of interest. If this is the case, accurate determination of the distribution of the condensate necessarily involves the use of a prognostic equation. However, it is not at all clear that the physics actually does demand a prognostic cloud water variable for all types of clouds. Consider a subtropical boundary-layer stratocumulus cloud, consisting entirely of liquid water droplets. These droplets are formed by condensation that occurs primarily in microscale (i.e., turbulence-scale) convective updrafts a few hundred meters across, with lifetimes of a few minutes. Many of these droplets are subjected to evaporation in nearby microscale downdrafts; therefore, the lifetime of an average cloud droplet in the stratocumulus layer is very brief. This means that the conservation equation for the large-scale average liquid water concentration is dominated by strong condensation and evaporation terms, which very nearly balance each other. For the space and time scales resolved by a climate model, the local time

rate of change and advection terms are negligible, compared to these source and sink terms. A similar line of reasoning applies for some other cloud types, such as fair-weather cumulus clouds or frontal stratus clouds. It should be noted, however, that these arguments do not apply to the mesoscale and microscale models that are used to simulate individual cloud elements. This is why such small-scale models have been using prognostic cloud water variables for many years.

The local time rate of change and advection terms are important for cirrus clouds, even on the relatively large space and time scales resolved by a climate model. Typical cirrus clouds contain many small crystals with terminal velocities less than or on the order of 0.1 m s^{-1} , and also larger crystals with terminal velocities on the order of 0.5 m s^{-1} (e.g., Heymsfield, 1975; Starr and Cox, 1985 a). Such particles take several hours to fall through the depth of the troposphere. The effective fall speed can

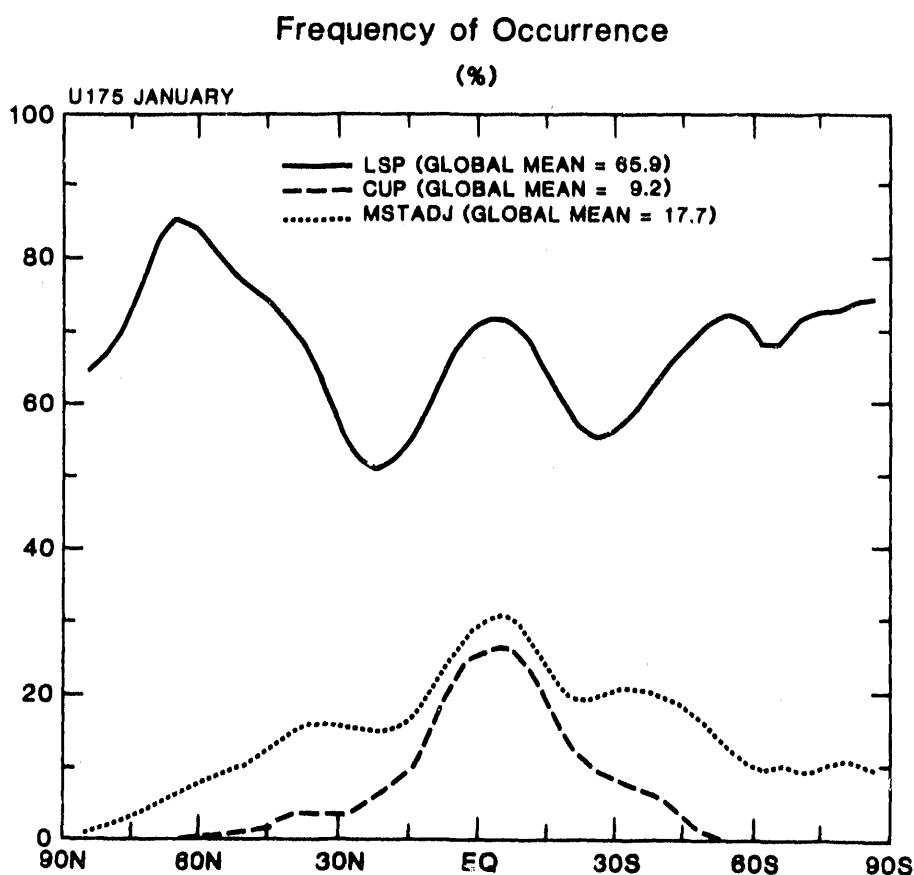


Fig. 1. Zonally averaged incidence of cumulus convection, large-scale saturation, and moist adjustment, in a January simulation with the GCM of Colorado State University.

be further reduced by mesoscale rising motion and/or convective lofting associated with the cirrus clouds themselves. Crystals are capable of surviving for extended periods as they fall through subsaturated air (Hall and Pruppacher, 1976).

Observations (Webster and Stephens, 1980) show that cirrus outflows from tropical convection can extend for many hundreds or even thousands of kilometers downstream from the convective disturbance that generates them. In such cases, the local time rate of change and advection terms of the conservation equation for the large-scale average concentrations of cirrus ice water must be comparable to the source and sink terms, so that a prognostic approach, including advective effects, is necessary to accurately predict cloudiness. In short, prognostic cloud water variables are necessary for accurate simulations of the effects of cirrus clouds on climate.

A second reason for using a prognostic cloud water variable is that a prognostic scheme can actually simplify the computational algorithms of the cloudiness parameterization, by providing a relatively convenient algorithm to determine the cloud water content of GCM grid volumes.

For climate modeling, precipitation processes and the radiative effects of the clouds are of roughly equal importance. It may be useful to separately prognose precipitating and nonprecipitating liquid water drops and ice particles. This suggests that future climate models may include a total of four prognostic condensate variables: large crystals, small crystals, large drops, and small drops.

The use of prognostic cloud water variables to determine the large-scale average distribution of precipitating and nonprecipitating cloud particles can directly replace the large-scale saturation parameterizations. For consistency, the same prognostic cloud water variables should also be incorporated into the cumulus parameterization. The parameterized convective clouds can then act as generators of cloud particles for the parameterized stratiform clouds. This concept points the way to a natural coupling of the convective and stratiform cloud parameterizations.

Methods exist to determine the optical properties of a grid volume containing a known concentration of cloud water (Stephens, 1978). However, this does not imply that a prognostic cloud water variable solves the problem of determining the cloud radiative forcing. As discussed by Harshvardhan and Randall (1985), the distribution of cloud water inside a grid volume strongly determines its optical properties. For example, if all of the liquid water is in one lump, its optical effects will be negligible. On the other hand, if the same mass of water is distributed over the grid volume as a uniform aerosol, its optical effects will be formidable. There is as yet no proven method, based on physical principles, to determine the subgrid-scale distribution of cloud water in a GCM grid volume. This subgrid-scale cloud water distribution is related to but more complex than what is usually called the "fractional cloudiness."

Through the use of a prognostic cloud water variable, we can predict the large-scale average condensate mixing ratio for each GCM grid volume. The cloud-scale average condensate mixing ratio within individual cloud elements has been shown, in both observational and theoretical studies (Feigelson, 1978; Somerville and Remer, 1984; Betts and Harshvardhan, 1987; Platt and Harshvardhan, 1988), to vary systematically with temperature; these cloud-scale average mixing ratios are controlled by microphysical and small-scale cloud-dynamical processes, and so should be amenable to semi-empirical parameterization.

The ratio of the large-scale average mixing ratio to the cloud-scale average mixing ratio is a measure of the fraction of the grid volume that is occupied by cloud. Let this ratio be denoted by w , i.e.:

$$w = \frac{\text{large - scale average condensate mixing ratio}}{\text{cloud - scale average condensate mixing ratio}} \quad (1)$$

A possible approach is to determine the numerator of (1) from the prognostic cloud water variable, and the denominator by an empirical assumption in the spirit of those mentioned above. In case the cloud-scale mean mixing ratio is uniform within the cloud elements, w is exactly equal to the fractional cloudiness (in a volume sense). Even when the cloud-scale mean mixing ratio varies within the cloud elements, w should be a useful measure of the small-scale variability of the cloud field. Evidence for this was presented by Stephens (1988), who found that the satellite observations of the meridional distribution of large-scale average cloud water reported by Prabakhara et al. (1986) do not show the systematic poleward decrease to be expected from the observed dependence of the local average cloud water mixing ratio on temperature. This suggests that the observed variations in the large-scale average cloud water are dominated by variations in "cloud amount" rather than local cloud water concentration. It may be possible to develop a parameterization of the radiative, dynamical, and cloud-physical effects of the small-scale variability of cloudiness, based, in part, on this approach.

However, it may also be necessary to take into account the role of small-scale convective motions in determining the fractional cloudiness. These motions are known to play an important role even in stratiform cloud systems (Lilly, 1968; Starr and Cox, 1985 a,b; Lilly, 1988). Randall (1987) has proposed elements of a parameterization for convectively driven small-scale cloudiness fluctuations. Perhaps the most natural cloud types to parameterize in this way are stratocumulus and shallow cumulus clouds that reside in the planetary boundary layer, since existing climate models already include boundary-layer turbulence parameterizations. Although stratocumulus cloud sheets are sometimes completely unbroken over tens of thousands of square kilometers, both common experience and satellite photos show that they are often riddled with small holes that appear to be associated with the turbulent character of the cloud. Even when the cloud sheet is unbroken, there are significant undulations of cloud top and cloud base, so that at a given level the fractional cloudiness can be between zero and one. The commonly observed break-up of a stratocumulus sheet into shallow cumuli

(Randall, 1980) necessarily reduces the fractional cloudiness; a satisfactory theory of the break-up must allow arbitrary cloudiness.

Betts (1973; 1983), Hanson (1981), Penc and Albrecht (1986), and Wang and Albrecht (1986) have discussed models of partly cloudy PBLs containing a single family of convective circulations. The circulations have both ascending and descending branches, and cloudiness can occur (or not) in either branch. Similar ideas have been used in observational studies (based on conditional sampling and/or joint distribution functions) by Lenschow and Stephens (1980; 1982), Greenhut and Khalsa (1982), and Wilczak and Businger (1983), Mahrt and Paumier (1984), Grossman (1984), Khalsa and Greenhut (1985), and Penc and Albrecht (1986).

These models show that the fractional cloudiness is closely related to the fractional area covered by rising motion, and also to the convective fluctuations of temperature and mixing ratio. To date, no method to determine these quantities has been demonstrated. Even if one can be found, the resulting fractional cloudiness parameterization will be rather complicated, since the fractional cloudiness will be determined, in part, by the turbulence, but at the same time the intensity and character of the turbulence will be determined, in part, by the presence of the clouds and their effects on the radiation field. Therefore, the cloud, turbulence, and radiation parameterizations will give rise to a coupled system of equations that must be solved simultaneously (Figure 2). It will be a challenge to devise an algorithm that is accurate and yet computationally fast enough for use in a GCM.

CONCLUDING REMARKS

Most upper tropospheric clouds are produced by convection. To simulate high clouds realistically, we have to simulate convection realistically. It is a waste of time to introduce a prognostic cloud water variable unless the convective source terms are realistically modeled. This implies that, even though cumulus parameterization has been a high priority for model development since the mid 1960s, it will have to remain so for another quarter century (at least). Now, however, the need to simulate the clouds produced by convection becomes as much of a driver as the need to simulate the convective heating and drying.

We need to expend more effort systematically comparing our model results against observations. Numerous observational studies of large-scale cloudiness have been conducted in recent years (e.g., International Satellite Cloud Climatology Project (ISCCP), Earth Radiation Budget Experiment (ERBE), First ISCCP Regional Experiment (FIRE), International Cirrus Experiment (ICE)), and more are planned (Atmospheric Radiation Measurement Project (ARM), Clouds and the Earth's Radiant Energy System (CERES)). For the reasons given above, it is important that future cloud-radiation experiments sample deep convective regimes such as those of Indonesia, or the Asian summer monsoon, or the midwestern United States in spring and summer.

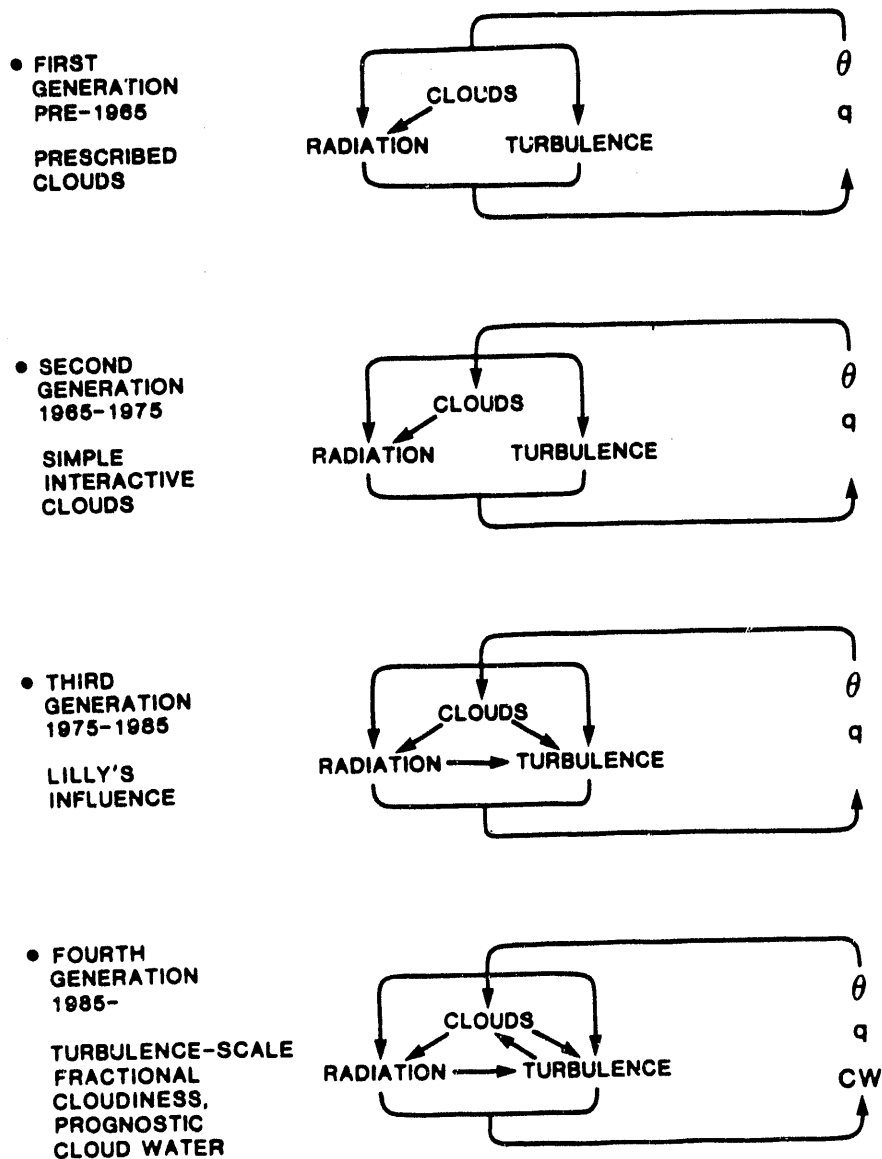


Fig. 2. Diagram illustrating the increasing complexity of cloud parameterizations used in climate models. The earliest parameterizations used prescribed clouds; later models determined the cloudiness as simple functions of temperature and humidity. Lilly (1968) showed that the cloudiness also affects the turbulence parameterization of the model. Emerging parameterizations determine the fractional cloudiness as a function of the turbulence as well as the temperature and humidity.

In view of the emerging importance of prognostic cloud water variables, it would be highly desirable to conduct one or more field experiments in which the various processes contributing to the time rate of change of the liquid water and ice concentrations are quantitatively measured, as functions of height, for a representative GCM grid column.

Finally, there is tremendous potential for the use of high-resolution, cloud-resolving models to develop and test cloud parameterizations for GCMs. Among the few examples of this approach so far are the studies by Moeng (1986) and Xu and Krueger (unpublished). The three-way combination of observations, a cloud-resolving model, and a GCM can lead to rapid progress towards developing improved physically-based cloud parameterizations for climate models.

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COMMENTS ON ATMOSPHERIC FEEDBACKS

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These comments address two aspects of the feedback processes in climate models: (1) the analysis of cloud feedback from the General Circulation Model (GCM) intercomparison, and (2) some new results on ice response to transient radiative forcing from Arctic haze.

One can define a simple analog representation of cloud feedback which permits examination of the reasons for the large dispersion between GCMs in the intercomparisons (Cess et al., 1989). This representation may serve as a guide in determining the variables to consider in future intercomparisons of models. It is also relevant to examine the sensitivities of cloud types to those variables. The objective of this comment is to illustrate one approach currently being carried out. The development is for the purposes of illustration only; in actual analysis, the analog is elaborated to provide a better representation of cloud feedback.

It is noticeable that clear-sky climate sensitivity agrees well within the range 0.4 to $0.6 \text{ K m}^2 \text{ w}^{-1}$ (Fig. 1). Although 68% of the global results (clear + cloudy) of the participant GCMs agree within the same range, some models show a greater climate sensitivity (up to a factor of 3 from the lowest to the highest sensitivity). The objective is to explain the difference between models. Since some data required to do this are not available for most models, a simplified analog approach is used.

In Fig. 2 I have formulated this simple analog in terms of independent cloud layers of a given albedo, emissivity, coverage, and cloud top temperature above a ground of zero albedo (for illustrative purposes only) and emitting like a black body at temperature T_0 . In its simplest form, the linearized cloud feedback may be expressed as a solar plus terrestrial terms. Each term is composed of a change of cloud coverage, plus a change of optical

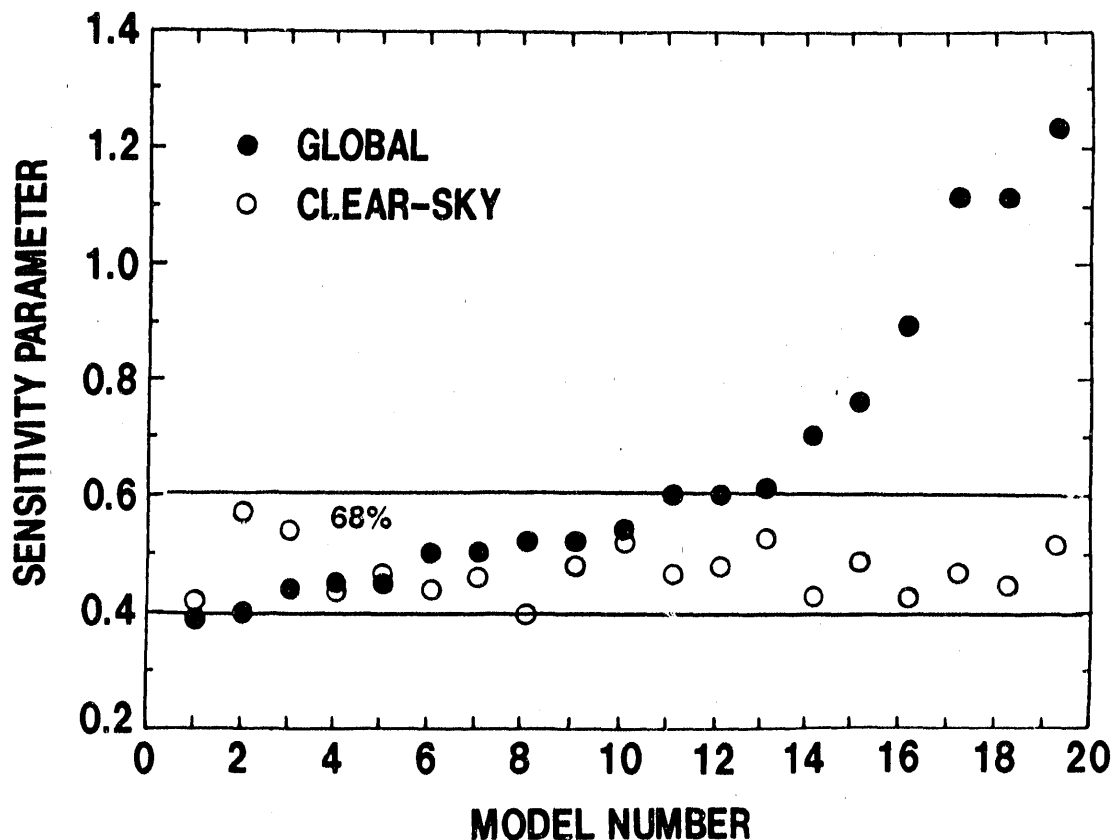


Fig. 1. Clear-sky climate sensitivity agrees well within the range 0.4 to 0.6 $\text{K m}^2 \text{w}^{-1}$. Although 68% of the global results (clear + cloudy) of the participant GCMs agree within the same range, some models show a greater climate sensitivity, (up to a factor of 3 from the lowest to the highest sensitivity).

properties (albedo and emissivity). A lapse-rate term also is included for terrestrial radiation. Of the 10 variables, 7 are identified as determinants for cloud radiative feedback; the variables are cloud cover, albedo, emissivity, and their variations, plus the variation in the lapse rate.

In Fig. 3 representative values have been extracted from the mean fields of the actual GCM run ($\pm 2 \text{ K}$ SST at the bottom of the figure for CCC GCM). At the moment, these values are only approximate numbers with the aim of assessing the individual terms. In application, the top equation is expanded to include a constant atmosphere (fixed transmittance to space, molecular diffusion and fixed surface albedo) and mean values of each variable extracted for the summer hemisphere, winter hemisphere and tropics. We consider one layer of high clouds and one layer of low clouds. The results show that the sign of cloud coverage change is opposite to that of the optical property change; that the lapse-rate term is small; and that net solar and IR nearly compensate each other. Furthermore, the high cloud and low cloud effects are of opposite sign. The net result of this estimation closely agrees with the actual GCM runs done for cases of fixed and variable

Considering a single cloud layer of temperature T , coverage C , albedo α , and emissivity ϵ above a ground surface of temperature T_0 , $\alpha_0 = 0$ and $\epsilon_0 = 1$. The radiative cloud forcing is

$$CF = -Q + \hat{Q} + F - \hat{F}$$

where Q and F are solar and terrestrial net fluxes at the top of the atmosphere, and hats represent clear sky fluxes:

$$Q = Q_0 (1 - \alpha C)$$

$$\hat{Q} = Q_0$$

$$F = -\sigma T_0^4 + \sigma (T_0^4 - T^4) \epsilon C$$

$$\hat{F} = -\sigma T_0^4$$

The radiative cloud feedback is

$$\delta CF = CF(T + \delta T) - CF(T)$$

or

$$\delta CF = -Q_0 (\alpha \cdot \delta C + C \cdot \delta \alpha) + \sigma (T_0^4 - T^4) (\epsilon \cdot \delta C + C \cdot \delta \epsilon) + 4\sigma (T_0^3 \delta T_0 - T^3 \delta T) \epsilon C$$

This model does not account for land-sea albedo contrast versus regional cloud changes, nor diurnal changes.

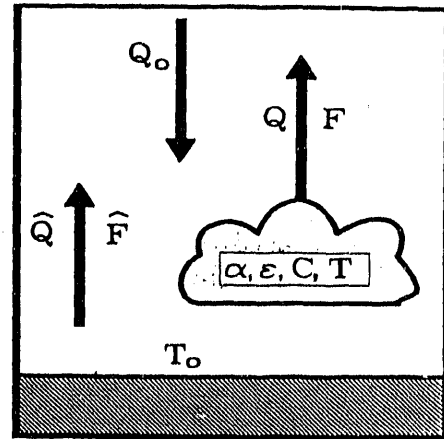


Fig. 2. A conceptual model of cloud feedback. This simple analog was formulated in terms of independent cloud layers of a given albedo, emissivity, coverage, and cloud top temperature above a ground of zero albedo and emitting like a black body at temperature T_0 . See the text for details.

SOLAR TERRESTRIAL LAPSE RATE

$$\delta CF = -Q_o(\alpha.\delta C + C.\delta \alpha) + F_o(\epsilon.\delta C + C.\delta \epsilon) + 4\sigma(T_o^3.\delta T_o - T^3.\delta T).\epsilon.C$$

High clouds (200 mb)					
-1.1	4.6	-13.0	-11.1	18.3	6.6
-1.1	-8.4		7.2		0.1
Low clouds (900 mb)					
1.1	1.8	-0.4	-0.6	0	3.4
1.1	1.4		-0.6		0.3
					(actual GCM runs)

- net cloud feedback with variable optical properties ≈ 0 W.m⁻² (-0.7)
- net cloud feedback with fixed optical properties = -5.1 W.m⁻² (-5.3)
- net cloud feedback with fixed clouds and optical properties = 0.4 W.m⁻²

Assumed values extracted from GCM:

T_o = 300K, Q_o = 350 W.m⁻², F_o = σ (T_o⁴ - T⁴) = 326 W.m⁻²
High clouds (200 mb):
T = 220K, δT = 10K, α = 0.32, δα = 0.12, ε = 0.86, δε = 0.19, C = 31%, δC = -4%
Low clouds (900 mb):
T = 290K, δT = 4K, α = 0.54, δα = 0.017, ε = 1.0, δε = 0., C = 14%, δC = -1%

Fig. 3. Evaluation of cloud feedback. Representative values have been extracted from the mean fields of the actual GCM run (± 2 K SST at the bottom of the figure for CCC GCM). These values are approximate; see the text for explanation.

optical properties. This agreement encourages us to pursue the exercise and inter-compare different participant GCMs.

One difficult area seems to be the highest clouds near the tropopause (Fig. 4), that typically are sub-visible thin cirrus, except in heavy convective cells. In most GCMs these clouds systematically increase in amount, partly due to lifting of the tropopause. At very low temperatures and low optical depths these clouds have a weak contribution to the total cloud feedback. However, if we assume black-body emissivities (or half-black), their contribution to cloud feedback becomes strongly positive. This fact may account for some of the dispersion in Fig.1.

Fig. 5 represents the partial derivative of cloud forcing for each of the seven selected variables discussed above. They are linearized about the state point (marked in the figure). In the CCC case, the cloud feedback due to low-level clouds is only sensitive to change in cloud amount. It is striking that thicker cirrus (near 200-300 mb) are very sensitive to most variables, with considerable changes in sign of the feedback, illustrating the importance of cirrus clouds in GCMs. Sub-visible cirrus are sensitive only to absolute albedo and emissivity, while, if assumed half black-body, these high cirrus become sensitive also to cloud coverage, resulting in a net positive cloud feedback.

The preliminary results for four other GCMs using this kind of simple analysis (in brackets) are shown in Fig. 6 in comparison to the result of their actual full simulation runs. Except for case 4 (due to crude estimates of cloud emissivity and albedo), the results show good consistency, lending support to this approach.

The second comment concerns the response of sea-ice to a transient radiative forcing from Arctic haze. Experiments using a moist aerosol model in the Canadian Climate Centre GCM (T32 version 2) results in the same feedback processes as in the double CO₂ experiment, which are responsible for the polar amplification of the climate warming. One particular result of the seasonal and regional Arctic haze forcing is the formation of a cool anomaly at lower Arctic latitudes. A direct consequence is an extension of the sea-ice in marginal areas. Fig. 7 shows the periodical increase in Arctic haze that leads to a seasonal radiative forcing (solar + IR).

The sequence of events in a 19-month simulation of Arctic haze is represented in Fig. 8. The control run is from years 15 and 16 of the basic climate experiment. The shaded boxes represent periods of anthropogenic haze when radiative forcing occurs. Snow and sea-ice melt during late spring and summer. During summer the oceans store extra heat which is released during fall and winter. The maximum ground temperature anomaly occurs in January, due to the steep surface inversion. During the second spring the area of sea-ice expanded at lower latitudes.

Fig. 9 shows the result of a change in ice mass (0.01 kg/m²) in the CCC GCM due to Arctic haze (5x actual concentration). The effects of spring warming in the central Arctic and cooling at lower latitudes are translated

SOLAR TERRESTRIAL LAPSE RATE

$$\delta CF = -Q_o(\alpha \delta C + C \delta \alpha) + F_o(\epsilon \delta C + C \delta \epsilon) + 4\sigma(T_o^3 \delta T_o - T^3 \delta T) \cdot \epsilon \cdot C$$

Sub-Visible Thin Cirrus (100 mb)				
-0.1	-1.5	-0.2	1.1	0.4
-0.1	-1.7		1.5	0.1
Over-Estimated Thin Cirrus (100 mb)				
9.0	-3.2	-2.1	8.8	4.4
9.0	-5.3		13.2	1.1

- net cloud feedback sub-visible thin cirrus $\approx 0 \text{ W.m}^{-2}$
- net cloud feedback for over-estimated thin cirrus = $+9 \text{ W.m}^{-2}$
- net cloud feedback with fixed optical properties = $+6.7 \text{ W.m}^{-2}$

Assumed values:

$T_o = 300\text{K}$, $Q_o = 350 \text{ W.m}^{-2}$, $F_o = \sigma(T_o^4 - T^4) = 369 \text{ W.m}^{-2}$
 Sub-Visible Thin Cirrus (100 mb):
 $T = 200\text{K}$, $\delta T = 6\text{K}$, $\alpha = 0.1$, $\delta \alpha = 0.006$, $\epsilon = 0.1$, $\delta \epsilon = 0.012$, $C = 10\%$, $\delta C = 3\%$
 Over-Estimated Thin Cirrus (100 mb):
 $T = 200\text{K}$, $\delta T = 6\text{K}$, $\alpha = 0.3$, $\delta \alpha = 0.06$, $\epsilon = 0.8$, $\delta \epsilon = 0.12$, $C = 10\%$, $\delta C = 3\%$

Fig. 4. Effect of thin-cirrus clouds on cloud feedback. In most GCMs these clouds systematically increase, partly from lifting of the tropopause. At very low temperatures and low optical depths they have a weak contribution to the total cloud feedback. However, if we assume black-body emissivities (or half-black), their contribution is positive.

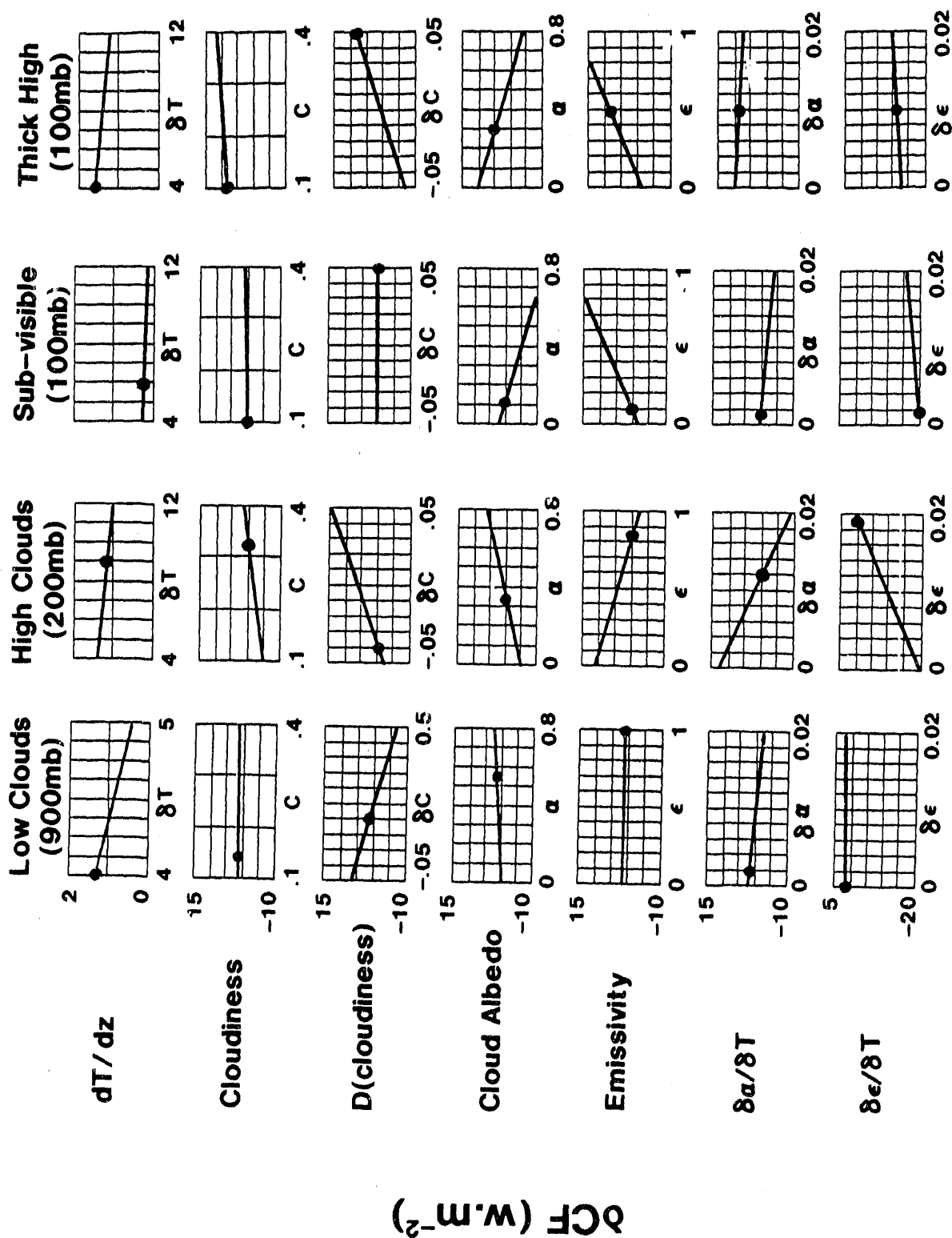


Fig. 5. The partial derivative of cloud forcing for each of the seven selected variables.

Models	$\Delta CF [w.m^{-2}]$		
	solar	terrestrial	net
1	-0.3 (-0.2)	-1.3 (-0.1)	-1.6 (-0.3)
2	1.5 (2.7)	0.7 (1.5)	2.1 (4.1)
3	1.5 (2.5)	0.3 (0.9)	1.8 (3.3)
4?	2.1 (0.1)	0.8 (6.5)	2.9 (6.5)

Fig. 6. GCM simulations vs. simple model. The preliminary results for four other GCMs using this simple analog approach (in brackets) are compared to the result of their actual full simulation runs. Except for case 4, because of crude estimates of cloud emissivity and albedo, the results show good consistency.

into a reduction and expansion of sea-ice, respectively. This cooling anomaly is due to a reduction of the mean meridional heat flux from mid-latitudes associated with the warm temperature anomaly near the pole during winter.

Results similar to the above are shown in Fig. 10 for four different runs with aerosol concentrations of 1, 5, 10, and 100 times current concentration. The cooling effects in the southern edge of the Arctic and the formation of a secondary circulation, apparent in the isotherms, are notable in the first three cases.

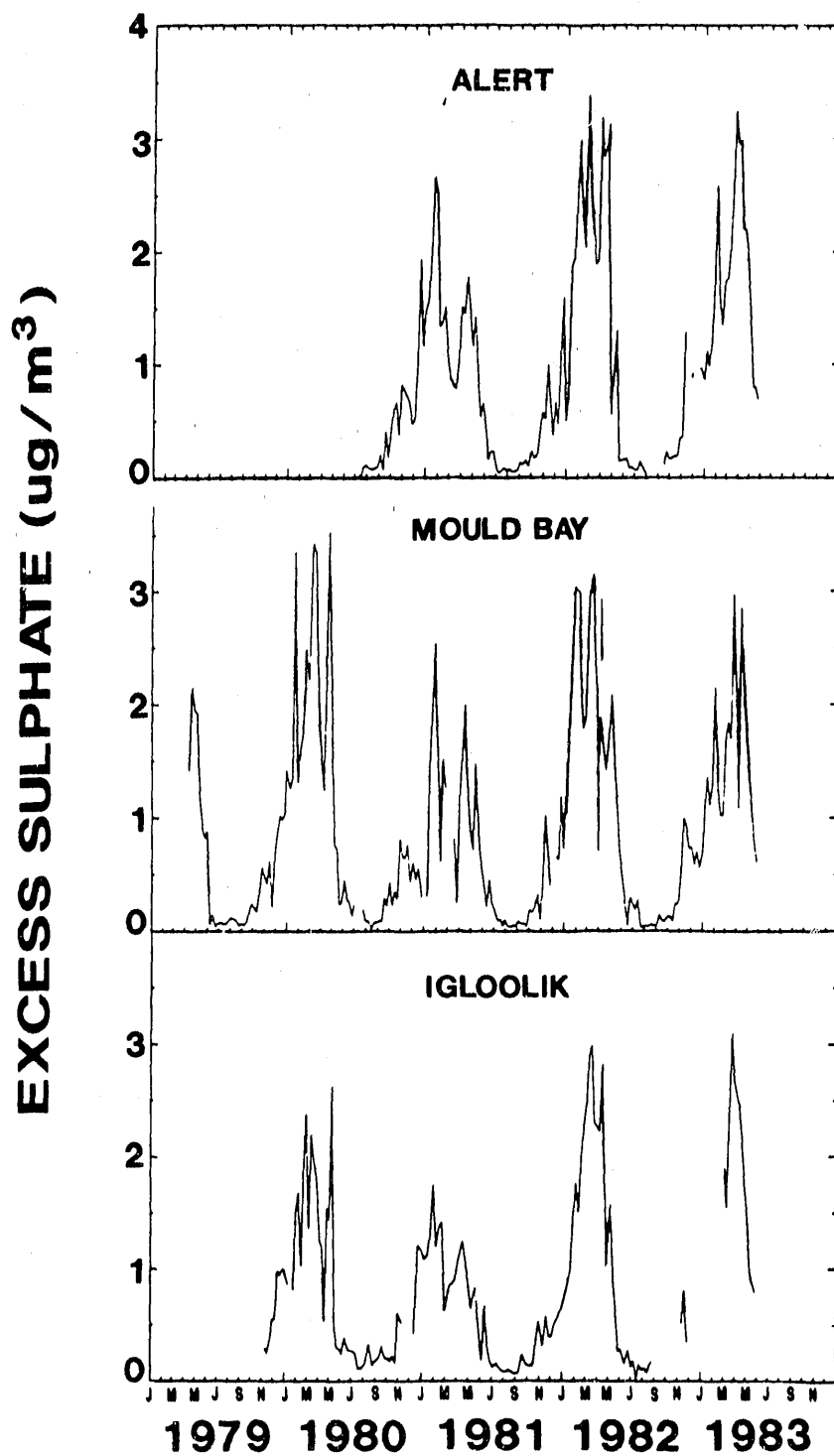


Fig. 7. The periodical increase in Arctic haze that leads to a seasonal radiative forcing (solar + IR).

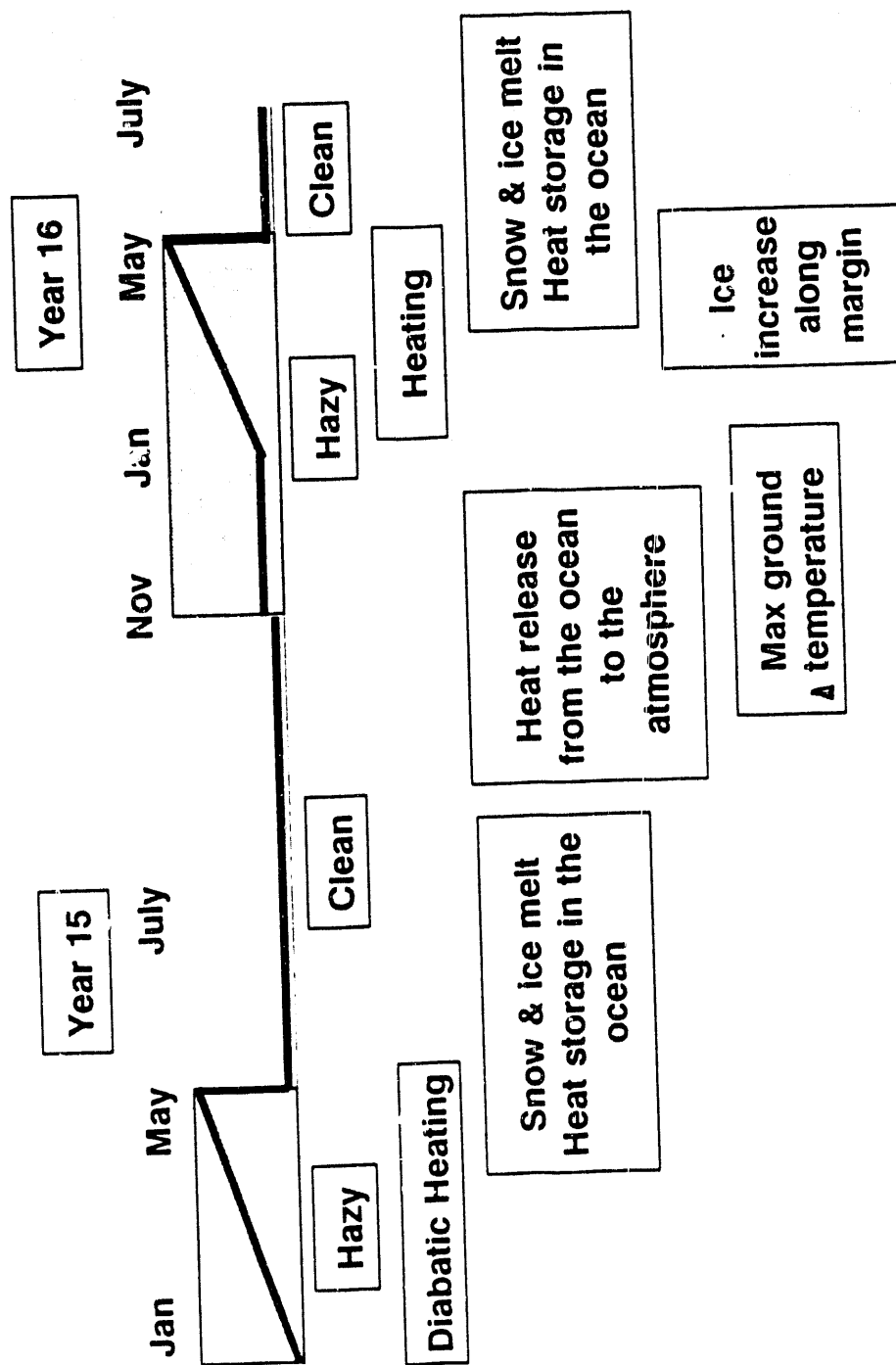


Fig. 8. Arctic Haze Experiment. The sequence of events in a 19-month simulation of Arctic haze. See the text for details.

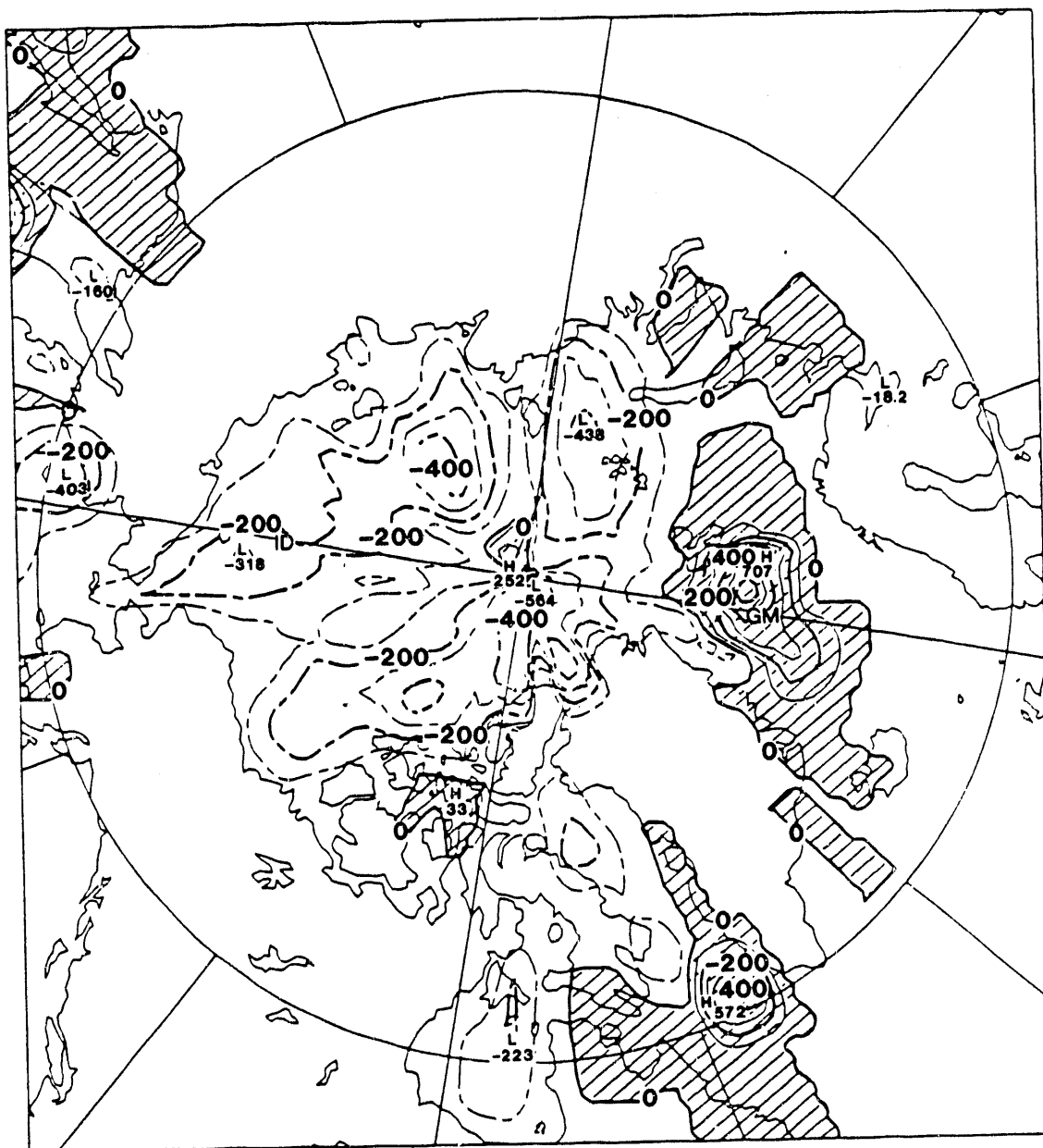


Fig. 9. The result of a change in ice mass (0.01 Kg/m^2) in the CCC GCM due to Arctic haze ($5\times$ actual concentration). See the text for details.

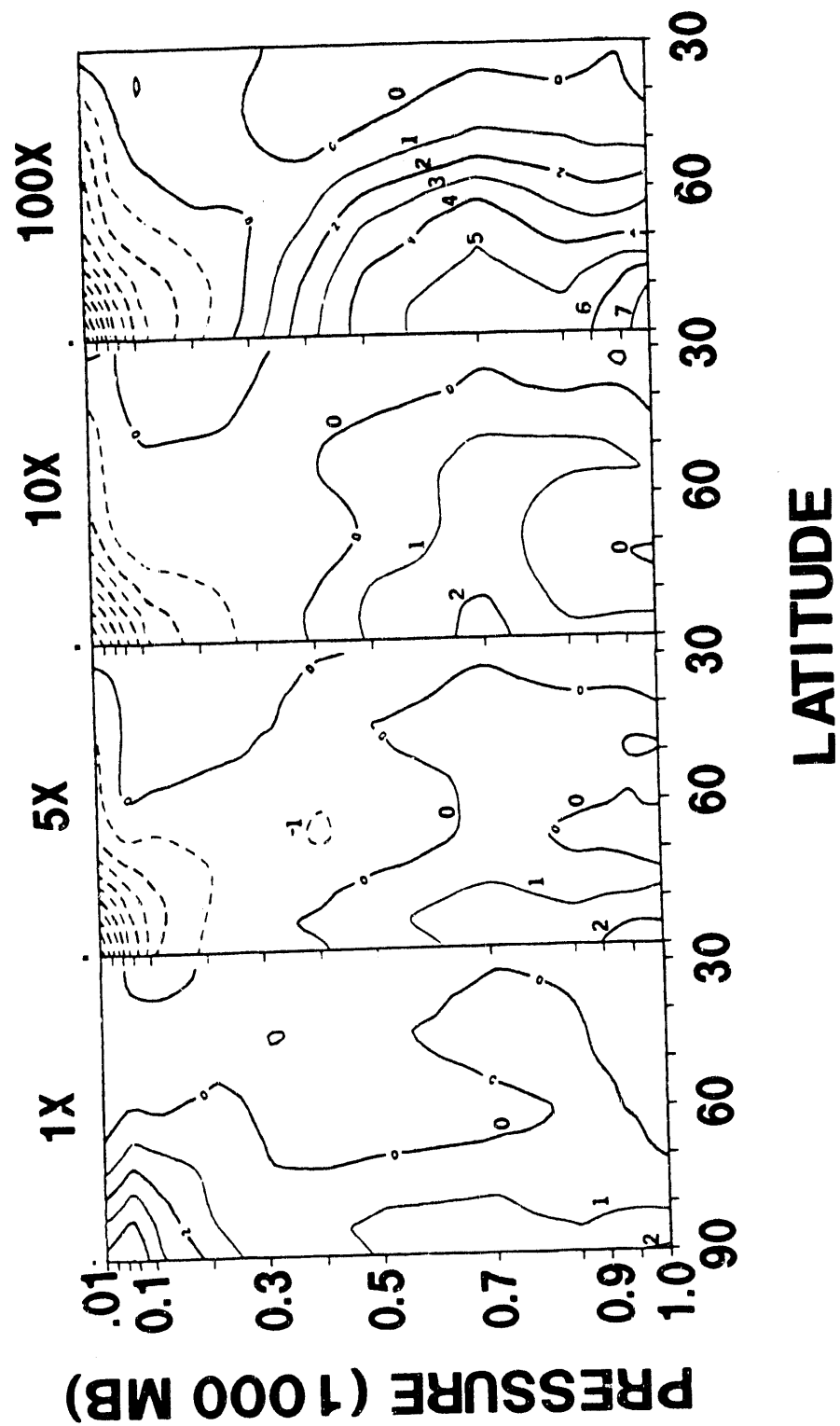


Fig. 10. Changes of the zonal mean temperature (C) with increasing concentration of aerosols. These results are similar to those in Fig. 9

COMMENTS ON ATMOSPHERIC FEEDBACKS

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The first point that I address deals with underscoring the importance of Ramanathan's comment that shortwave warming in cirrus may play an important role in limiting oceanic warming through stabilization of the thermodynamic structure of the upper troposphere.

Radiative effects are thought to play a key role in the evolution of stratocumulus, altocumulus, cirrus, and thunderstorm anvils, among other types of cloud. Destabilization of a cloud layer resulting from radiative warming at cloud basetop can lead to overturning, convection, and cloud-top entrainment. The latent heat, which is released (or absorbed) as a result of condensation (or evaporation) contributed by the radiative cooling (or warming), acts to partially offset pure radiation effects on cloud temperature and water mass evolution.

At low temperatures characteristic of tropical anvils almost all of the radiative effects are realized, i.e., condensation (or evaporation) effects on air temperature are minimal. This effect is very significant at warmer temperatures, for example at +20°C, where only about 30% of the radiative heating is realized. Therefore, anvils may have a significant effect on stabilizing the upper troposphere in tropical areas and acting as a brake on convective processes. I support the concept of a tropical cirrus experiment to quantify the effect of cirrus on local and global climate. Experiments such as Tropical Ocean Global Atmosphere/Coupled Ocean-Atmosphere Response Experiment (TOGA/COARE), and First ISCCP Regional Experiment (FIRE) are not likely to organize a tropical/cirrus component of their experiment since it does not fit in with either program objectives or site locations.

In connection with a tropical cirrus experiment, I point out that current aircraft instrumentation and aircraft platforms are inadequate to

study tropical cirrus anvils. Aircraft hydrometeor spectra cannot be measured accurately down to sizes typical of tropical cirrus, tens of microns or less. Aircraft platforms cannot reach altitudes topping the cirrus (above 50 Kft) to obtain both hydrometeor information and to measure radiative properties at these altitudes. New instrumentation and aircraft platforms are needed for such an experiment. I note the potential importance of sub-visual cirrus on climate in tropical areas, possibly implying an effect similar to that proposed by Ramanathan.

Finally, I believe that a parameterization for cirrus clouds recently developed by myself and Leo Donner of the University of Chicago will aid in evaluating the effects of cirrus on climate (Heymsfield, 1990).

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ATMOSPHERIC FEEDBACKS: COMMENTS

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Perhaps the most surprising aspect of General Circulation Models (GCMs) is that they work as well as they do. It is daunting to think of starting with pure silicon (i.e., a computer), knowledge of the basic laws of physics, data for the physical properties of air and water, and some geophysical properties of the planet, and setting out to model the earth climate system. Once that daunting challenge is faced, then modeling becomes a series of choices, constrained by the finite capacity of the computer at hand, guided by a sense of what is important to model actively versus what can be built in or "hard wired" in the model.

Why is there an interest in and emphasis on models? It is because we expect to learn something we do not already know. If we build models that we have confidence in, then we can address "What if" questions, such as: "How would the climate system respond to a particular perturbation in forcing?" and have some degree of confidence in the answer the model gives us. However, a concern is that features which are "hard wired" in the models may fix the values of certain quantities that should be allowed to vary as the models attempt to respond to forcings imposed while examining a given "what if" scenario. If this situation occurs, then a feedback is overlooked and the model will give a wrong answer.

Cess has discussed (this volume) two classes of diagnostics of GCMs: comparison with observation, such as the Earth Radiation Budget Experiment (ERBE) data, and model-to-model intercomparisons. Clearly, these comparisons and intercomparisons have revealed major areas of departure, which point to the areas that should be addressed as we try to improve the models. To me, more remarkable is the extent of agreement, both among the several models and between models and observations. For example, from a policy perspective, much emphasis is placed on the uncertainty associated with a factor of three

in λ , the climate sensitivity coefficient, that is obtained in model inter-comparisons. I think it borders on astounding that the range in λ is so narrow.

A concern that I raise here is whether the extent of this agreement is a consequence of conscious or unconscious calibration of the models to agree with climate properties characteristic of the present earth. Intellectually satisfying as it might be, it is impossible to model everything in a GCM from first principles; therefore, many features of the modeled phenomena are chosen on the basis of common sense or common experience. If these choices were made in all of the GCMs being compared, and if they were based on the present climate, they would not be discerned either in these intercomparisons or in comparisons with the present climate.

Cloud parameterization is one such area where experience has heavily influenced the models. A major reason for this is the great range of length scales involved, from tenths of micrometers for cloud condensation nuclei, to tens of microns for cloud droplets, to hundreds of meters for cloud depths, to hundreds of kilometers for the horizontal dimension of a grid cell in a GCM that may or may not have a cloud in it or may be partly cloudy by some algorithm; see Fig. 1.

Why is there such emphasis on and concern with clouds? There are two reasons. First, clouds exert a great influence on short- and long-wave radiative forcing: $\sim -45 \text{ W m}^{-2}$ for shortwave, and $\sim +30 \text{ W m}^{-2}$ for long-wave forcings, as inferred from ERBE. Second, these forcings depend entirely on radiative properties of clouds, which, in turn, depend entirely on the cloud's microphysical properties, which are not dealt with explicitly in GCMs. The persistence and even the existence of clouds are sensitive to their microphysical properties: clouds are inherently unstable thermodynamically (by virtue of their large surface area, compared to a single, large drop) and gravitationally (compared to the liquid water sitting on the earth surface). The existence of clouds, as we know them, and their radiative properties important in GCMs depend sensitively on their microphysical properties, and, in turn, on the presence, concentration, and character of the aerosol particles that serve as cloud condensation nuclei (CCN) on which water condenses when the cloud forms. These properties include:

- Number of drops per cm^3 .
- Radius of a drop.
- Short-wave reflective properties.
- Short-wave heating.
- Blackness/grayness in the IR.
- Precipitation rate.

If parameterization of these features of the clouds is fixed in models, whereas in fact they change in response to a change in forcing, such as enhanced CO_2 , then the models will give the wrong answer, yet the problem would not be discerned in comparisons among models nor in comparisons with observations.

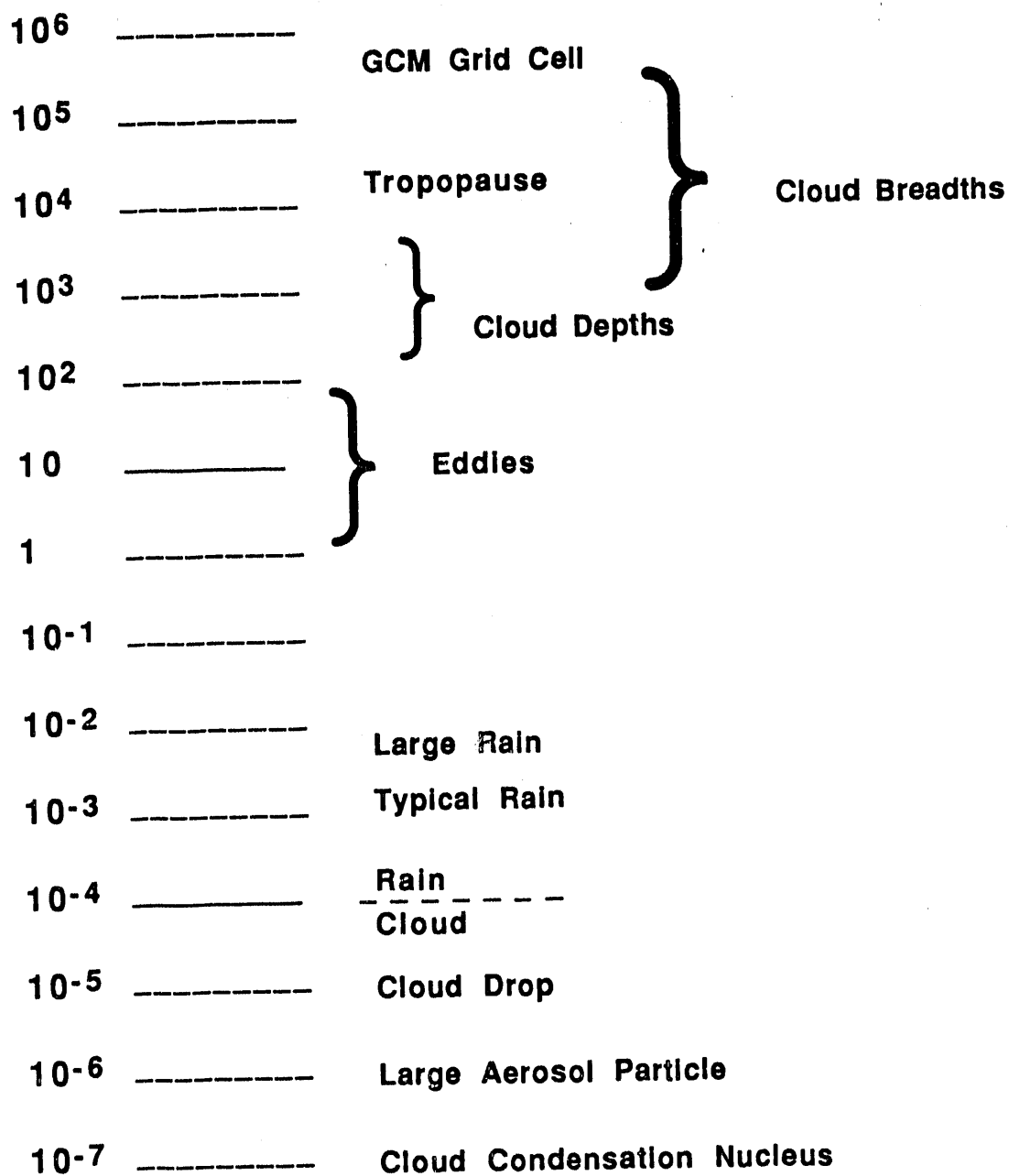


Fig. 1. Length scales (meters) pertinent to clouds.

A related concern is the question of whether industrial activity occurring on the same time-scale as the perturbation in CO_2 forcing might have appreciably perturbed the microphysical and radiative properties of clouds. This suggestion, which appears to have been first made by Twomey in 1972, recently received considerable attention. There is firm reason to believe that if the number density of cloud droplets (CD) is increased while other cloud properties such as liquid water content and path are held constant, then the top-of-cloud albedo will increase (Twomey, 1977). Such an increase would be expected if the number density of CCN were increased as would be expected as a consequence of industrial pollution, particularly the emission

of sulfur dioxide which forms sulfate aerosol. Fig. 2 shows the sensitivity of top-of-cloud albedo to an increase in CD number density, other things being held equal. Fig. 3 shows the expected sensitivity to change in cloud droplet number density of top-of-cloud albedo, top-of-atmosphere albedo, hemispheric-mean albedo, and hemispheric-mean short-wave cloud radiative forcing (CRF). The latter two values are based on Warren's statistics for mean planetary coverage by marine stratus and stratocumulus clouds. The figure shows that these quantities are highly sensitive to relatively small perturbations in CD number density. A doubling in CD concentration leads to a decrease in short-wave CRF by 5 W m^{-2} . A similar sensitivity was recently reported by Slingo (1990), based on GCM calculations. Increases in CCN and CD concentration by an order of magnitude or more associated with industrial pollution are common, and it is argued that even for the entire Northern Hemisphere the average increase in CD concentration due to industrial emissions might be as much as 30 or 50 percent (Twomey et al., 1984; Schwartz, 1988).

In view of these apparent perturbations in forcing, without any apparent hemispheric or regional response (though this is far from established), it seems imperative to examine whether there are unsuspected feedback processes that are stabilizing the climate against changes caused by this perturbation.

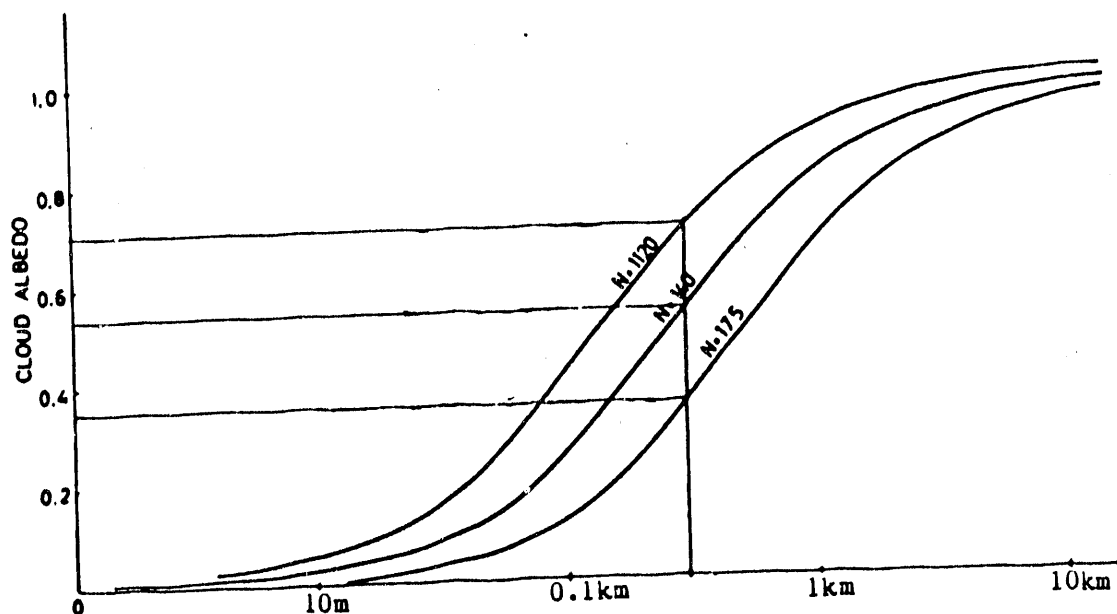
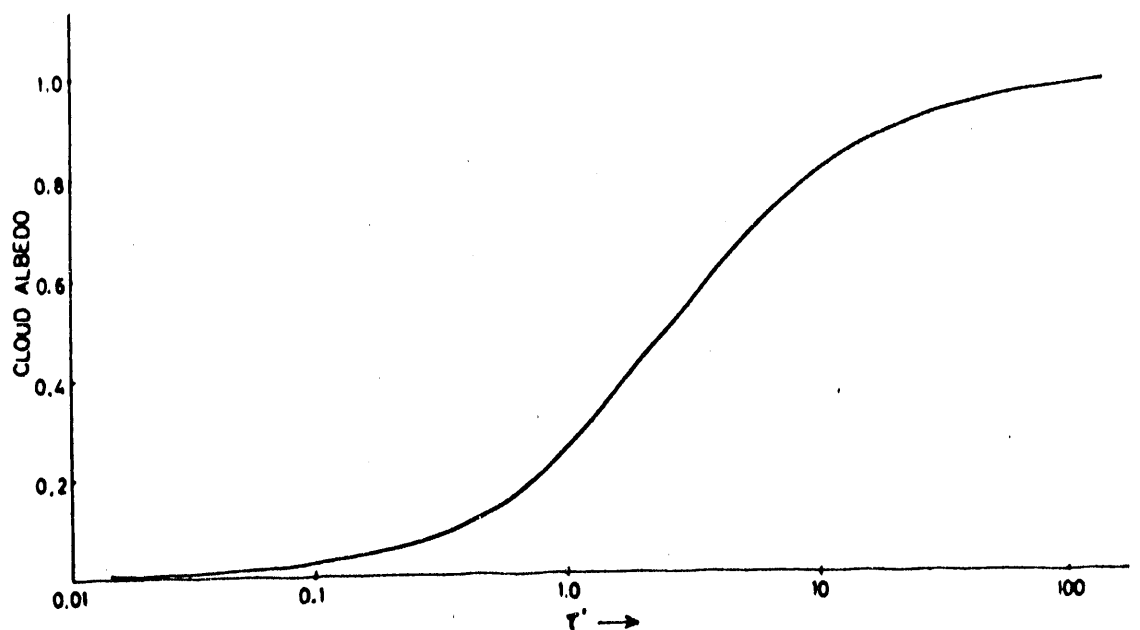


Fig. 2. a, Dependence of cloud albedo on scaled optical thickness of cloud. b, The data of Fig. 2a converted to geometric thickness of cloud. The central curve relates to fairly clean maritime conditions in which the cloud nucleus concentration and ensuing cloud drop concentration are 140 cm^{-3} . The effect of an eight-fold increase in the concentration is the upper curve, while the lower curve applies to case of a concentration of one-eighth of the first value. Adapted from Twomey (1977).

Calculated Sensitivities to Cloud Droplet Number Density

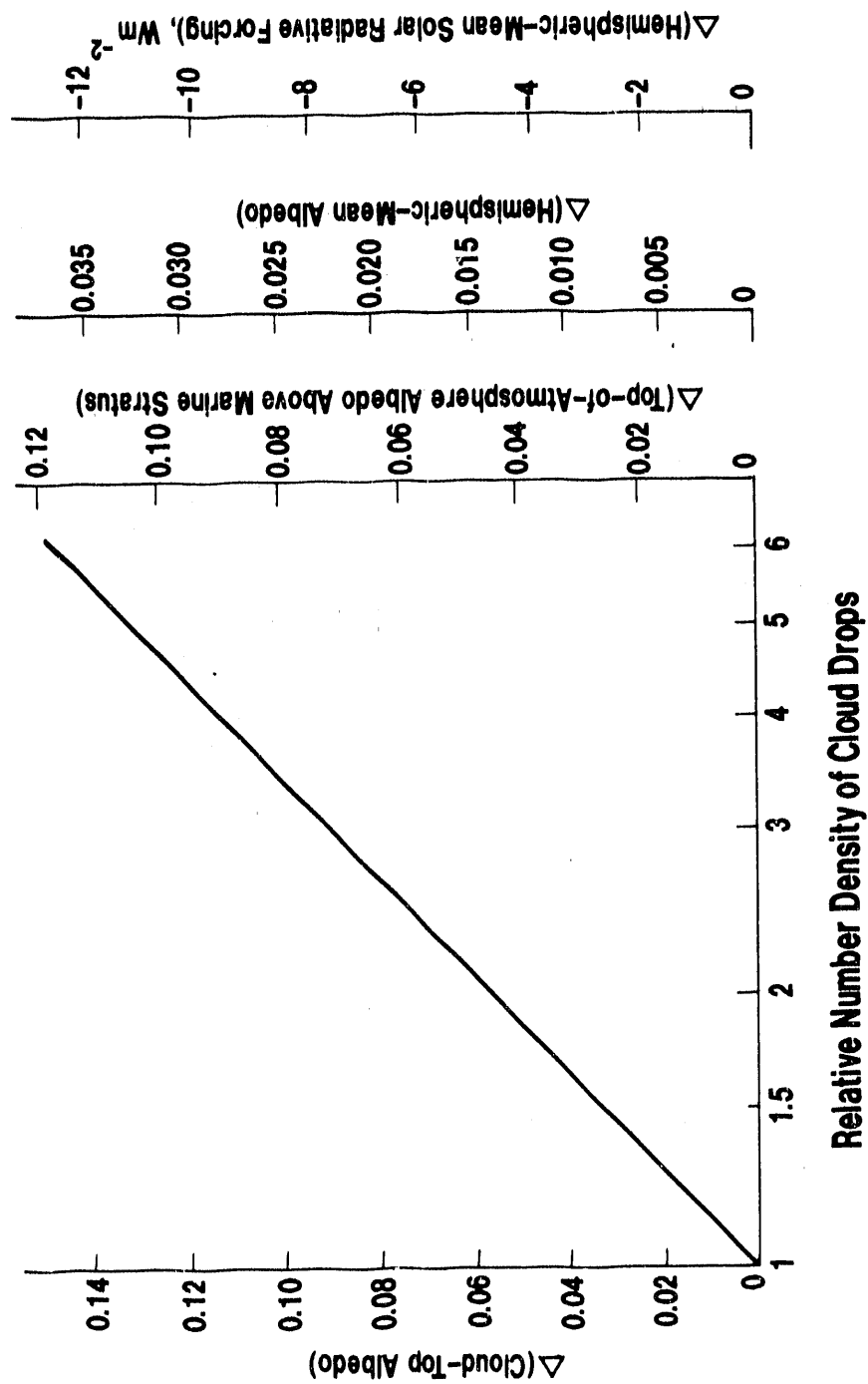


Fig. 3.

Dependence of cloud top albedo, top of atmosphere albedo, hemispheric mean albedo and hemispheric mean solar radiative forcing on relative number density of cloud drops. From Schwartz (1989) based on Charlson et al. (1987).

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OCEAN THERMAL TRANSIENTS: A PROGRAM OF DATA ANALYSIS, MODELING, AND
MONITORING OF THE ATLANTIC OCEAN THERMOHALINE CIRCULATION*

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INTRODUCTION

CLIMATIC VARIATIONS AND THE ATLANTIC SST

The rate at which greenhouse warming will occur is a critical unknown in our ability to react to this climate change. Predictions based on models show global warming ranging from 1.5°C to 4°C in response to a doubling in greenhouse gases. However, whether this warming occurs in 50 years or 100 years still is not certain. Part of this uncertainty is related to an inability to predict the natural cycle of atmospheric temperatures. Changes in hemispheric temperature approaching 1°C over a span of 50 years have been observed in the past century. Since the natural and greenhouse effects are additive, the phase of the former cycle (whether background temperatures are rising or falling), in part, will determine how rapidly global warming proceeds. To predict the effects of natural climate variability at periods ranging from interannual to decadal and longer requires an understanding of the role of the ocean in driving climate change. Studies in the Pacific have shown that the variability of sea surface temperature (SST) in this basin significantly impacts global climate. The Atlantic Ocean also seems to be a major player in establishing change in global climate at various time-scales. However, the present understanding of the Atlantic's role in climate change is much less developed than in the case of the Pacific.

*Much of the text comes directly from the ACCP planning document, Atlantic Climate Change Program Science Plan, NOAA Climate and Global Change Program Special Report No. 2.

The Atlantic sector has several advantages for understanding climate change. The wealth of historical data (Fig. 1), the strength of the thermohaline signal in terms of water mass formation, and the magnitude of surface temperature anomalies make this an obvious region of focus. The atmospheric response and its role in the air-sea coupled problem demand a global analysis.

From a viewpoint on climate, the direction and intensity of the thermohaline circulation are highly significant, because of the large north-south transfer of heat from the South to North Atlantic mentioned previously, which extends to very high latitudes in the North Atlantic. Theoretical work by Stommel (1961) and Rooth (1982) and modelling experiments by Bryan (1986) suggest that the thermohaline circulation of the ocean may have more than one equilibrium state. This idea has been illustrated in the context of a detailed coupled ocean-atmosphere model by Manabe and Stouffer (1988). Their global coupled model has two equilibrium climates. In one case, an active thermohaline circulation exists in the North Atlantic, corresponding to today's climate. In another case, fresh, relatively cold surface waters dominate the northern North Atlantic and the thermohaline circulation is almost nonexistent. These studies raise interesting questions about the stability of the Atlantic thermohaline circulation, its possible variations in the recent past, and how the thermohaline circulation might be modified by greenhouse warming.

ATMOSPHERIC RESPONSE OF ATLANTIC SST: MODEL EXPERIMENTS

High speed computers have made it possible to organize large masses of data in convenient data sets, and numerical experiments with models permit the quantitative exploration of a wide range of ideas suggested by the data and by climate theory. As shown in the Tropical Ocean/Global Atmosphere (TOGA) program, these new tools go a long way to removing the most formidable roadblocks to understanding large-scale air-sea interaction. Wallace and Jiang (1987) show that the "dipole" type of sea-surface temperature in the western North Atlantic in winter is associated with a displacement of the normal storm track and a wave train propagating eastward over Europe. Numerical experiments carried out by Palmer and Sun (1985), Dymnikov and Filin (1985) and, more recently, by Lau and Nath (in press), demonstrate on a seasonal time-scale that specified sea-surface temperature anomalies can excite very similar responses in atmospheric models. The study by Lau and Nath considers an integration carried out from 1950-1980, which provides a much higher level of statistical confidence than in any previous study.

In a separate study with the climate model of the British Meteorological Office, Folland et al. (1986) demonstrated that sea-surface temperature anomalies in the Northwest Atlantic could excite changes in monsoonal rainfall in the Sahel region of Africa, very much like the observed changes between the 1950s and 1970s. Warm sea-surface temperature anomalies in the Northwest intensify the African monsoon and cause greater summer rainfall. The implication of the model experiments of Folland et al. (1986) is that

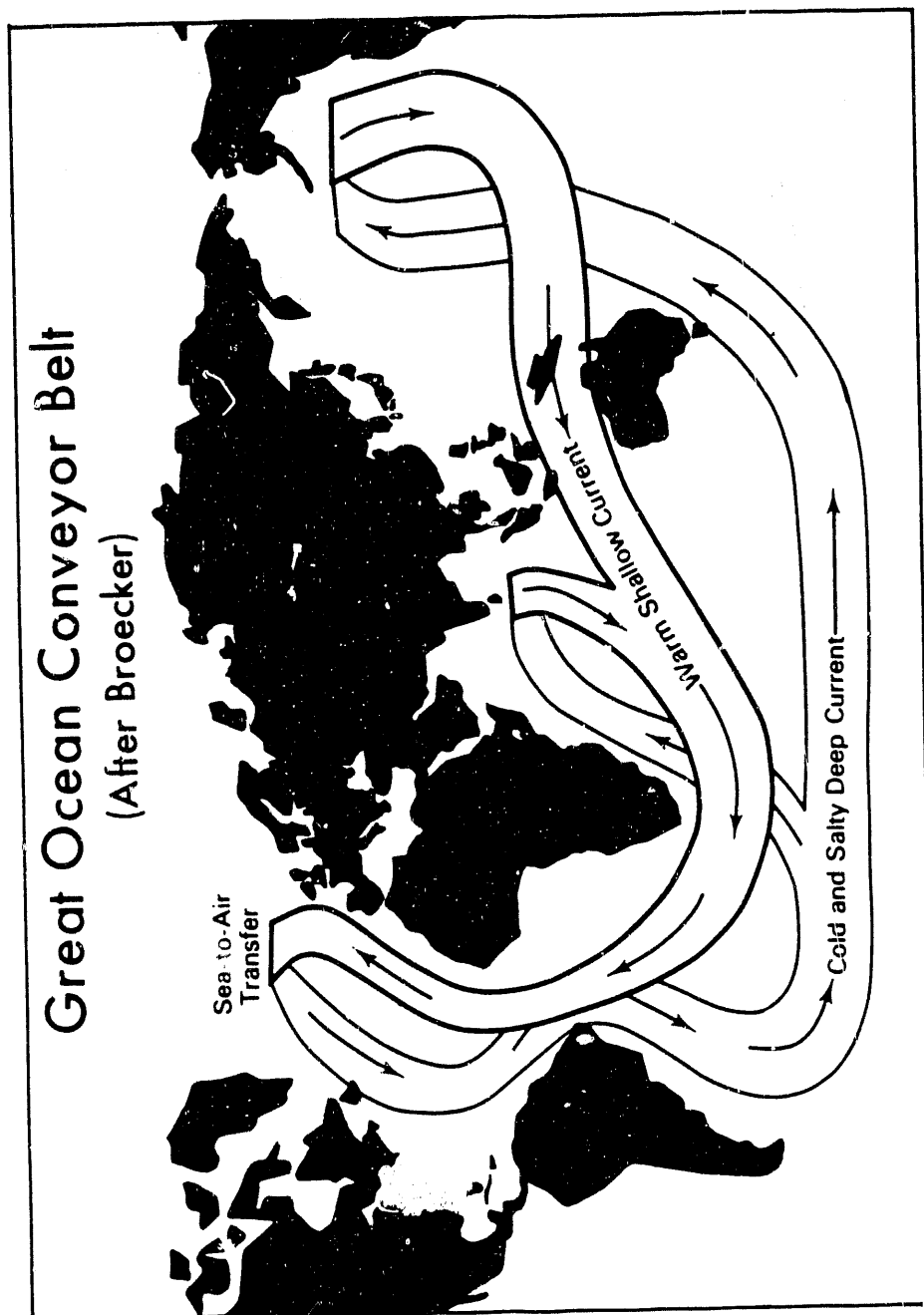


Fig. 1. Schematic diagram of the thermohaline conveyor belt circulation (Broecker, 1985), which illustrates the role of water mass formation in the overturning circulation of the world ocean. Only parts of this circulation have been directly measured.

the persistent dry conditions in recent decades in the Sahel region have been largely caused by the cooling of the North Atlantic, which began in the late 1960s.

Increasing attention is being drawn to the role in climate of salinity anomalies. Under present conditions, the deep World Ocean is relatively fresh because of the injection of low-salinity waters in the high-latitude sinking regions.

A variety of evidence points to major changes in the Atlantic thermohaline circulation and in atmospheric conditions as recently as the last deglaciation, probably associated with large injections of freshwater from the melting North American ice sheet. Certainly, recent years have seen a temporary shutdown of convection in the Labrador and the Iceland Seas because of a large but transient salinity anomaly. This event represents one of the most persistent and extreme variations in global ocean climate observed in this century.

Numerical experiments carried out recently provide some new insight on how north-south displacements of the polar halocline can produce bi-stable climatic states (Bryan, 1986; Manabe and Stauffer, 1988). F. Bryan coined the term "halocline catastrophe" for an equatorial displacement of the halocline which shuts off the normal convective overturning of the thermohaline circulation.

The entire heat capacity of the atmosphere is equivalent to less than 3 m of water. For this reason, climate variability extending over several years must involve the ocean in an important way. The question is whether the role of the ocean is a purely passive one through heat storage in the upper ocean, or whether non-local processes in the ocean are important. We know that changes in oceanic circulation are an integral part of the El Nino-Southern Oscillation phenomena. While results from models and empirical evidence support an active role of circulation climate variations of the Atlantic Ocean, the question is still open. We propose a program to analyze historical data, modelling, and ocean monitoring on seasonal to decadal time-scales to answer this basic scientific question.

Uncertainties in Climate Change Associated with Thermal Transients

As pointed out, decadal variations of climate associated with very large-scale interaction are a major difficulty in identifying climate change due to the build up of greenhouse gases in the atmosphere. Climate variations on a decadal time scale tend to be polar amplified, in the same way as predicted greenhouse warming, making the detection particularly difficult.

The relative importance of local versus non-local processes in the ocean in decadal climate variations is not known. It is very important to answer this question in order to devise a good monitoring system for the post-World Ocean Circulation Experiment (WOCE) period.

Coupled models are one of the best tools for gaining insight on the very limited data sets available for studying decadal climate variations. A major difficulty of the present generation of coupled models is that they cannot simulate sea-surface temperature and sea-surface salinity distributions without artificial corrections to the surface flux fields (Manabe and Stouffer, 1988). Until these corrections can be significantly reduced, this will remain a significant source of uncertainty in the use of these models for projecting greenhouse warming, and gaining insight on decadal climate variations.

Recommendations for an Air-Sea Interaction Program

Recent geological findings and results of ocean-atmosphere coupled models have provided new insights into the sensitivity of global climate to Atlantic thermohaline processes. Preliminary experiments with atmospheric models have shown significant response in the global atmosphere to realistic sea-surface temperature anomalies in the Atlantic. An initial emphasis of the program should be to confirm these atmospheric results and provide more detail. Observations show that sea-surface temperature anomalies in the Northwest Atlantic are modulated by wind and irregular seasonal thermohaline convection. A particularly important question is why these anomalies persist over many years. The low heat capacity of the atmosphere implies that these anomalies are intimately tied to changes in the thermocline structure of the ocean and its feedbacks to the atmosphere. Therefore, the focus here is on the thermohaline circulation and its interaction with the atmosphere. The focus on thermohaline circulation in the Atlantic is deliberate. The signal is strong and the historical data base is superior to that in any other region of thermohaline convection.

The scope of the atmospheric component of the program must be global, while the oceanographic monitoring and measurements would be restricted to the Atlantic because of indications that this region plays a major role in long-term climate variability.

GOALS OF THE ATLANTIC CLIMATE CHANGE PROGRAM

The goals of the Atlantic Climate Change Program (ACCP) are:

1. To monitor, describe, and model the space-time variability of the large-scale meridional circulation of the Atlantic and its role in producing variability of SST, sea ice, and salinity over the Atlantic Ocean seasonally, annually and interannually.
2. To determine the response of the global atmosphere to persistent SST and sea ice anomalies in the Atlantic Ocean and to develop coupled ocean-atmosphere models to simulate and predict seasonal to decadal changes over and around the Atlantic Basin.
3. To design a program to monitor the changes in the conveyor belt circulation (Fig. 2) and to develop a suitable modelling program to

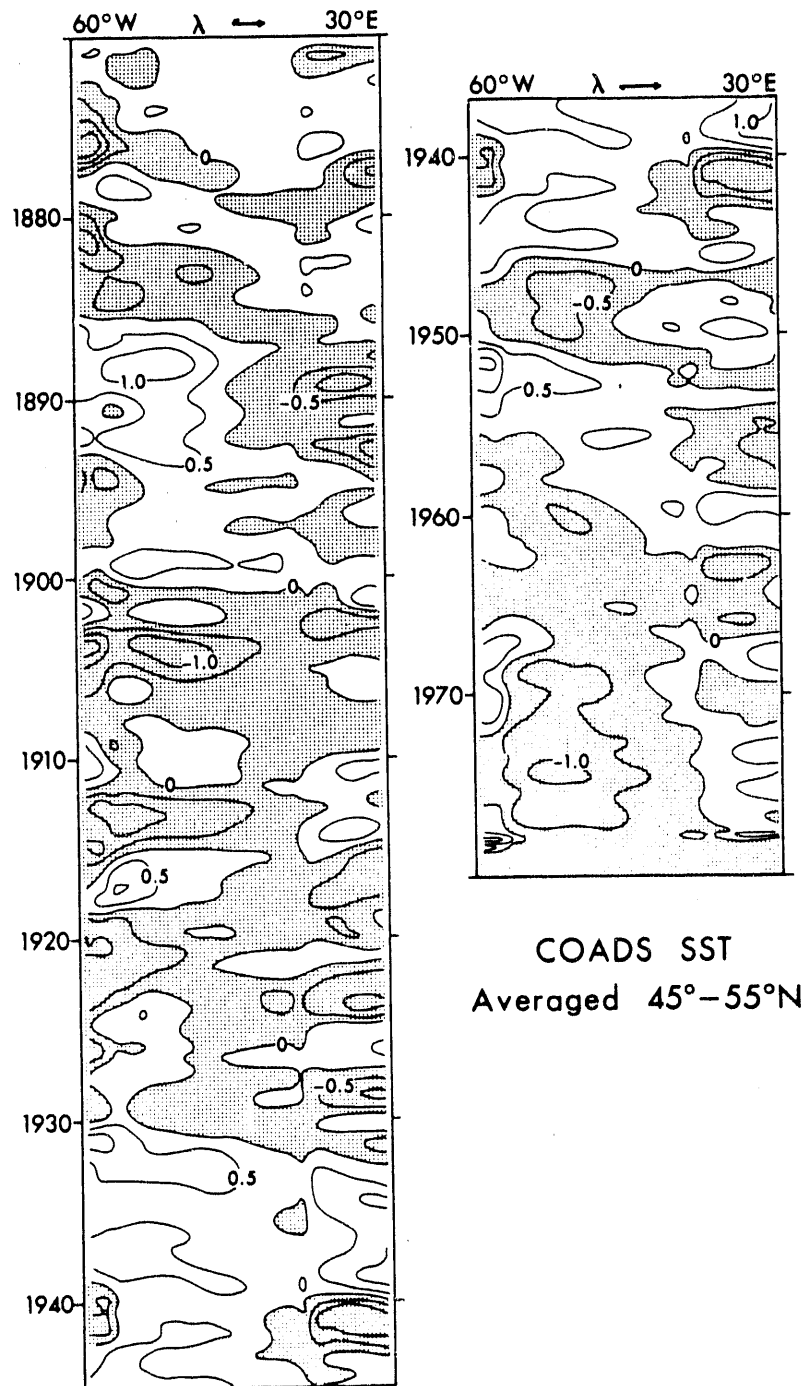


Fig. 2. Patterns of Atlantic sea-surface temperature based on COADS (Comprehensive Ocean-Atmosphere Data Set) data averaged from 45° to 55° N and detrended with respect to time for 1880 to 1970.

assimilate observations to help understand the mechanisms that determine the fluctuation of the conveyor belt circulation.

Program Elements

The Atlantic Climate Change Program is a study of climate variability related to thermohaline processes in the Atlantic Ocean. It is a joint undertaking by atmospheric scientists and oceanographers, based on modelling, examination of data sets, and long-term measurements of thermohaline processes in the Atlantic Ocean. The ACCP is envisioned as a balanced effort in each of these three areas.

The Program is envisioned as having three major elements to be carried out at Universities and in National Oceanic and Atmospheric Administration (NOAA) laboratories. An important premise of the ACCP is that there be a balanced effort in each of these elements.

Element I. Much of the data on climate variability in the Atlantic area is not organized in a way in which it can be easily used to study air-sea interaction. This element consists of the organization and analysis of data sets for both the ocean and atmosphere.

Element II. The application of models to understand the data and to design a monitoring system is a major focus of ACCP. A major effort will be made with atmospheric models to determine the sensitivity of the global atmosphere to Atlantic SST anomalies. The core of the monitoring system is envisioned as an operational ocean circulation model of the Atlantic, continuously updated by real-time data provided by satellites, ships of opportunity, and continuously reporting *in situ* instruments. We anticipate that developing a climate data assimilation scheme will lead to very significant improvements of atmospheric and oceanic models. In connection with operational ocean modelling for the TOGA program, it was found that the ocean mixed layer is a sensitive indicator of low frequency errors in atmospheric model heat fluxes. Data assimilation shows where the first assumptions of the models are seriously incorrect, and shows where the models have serious deficiencies.

Element III. The observational component of ACCP will concentrate on time-series measurements, to provide insight on the link between the large-scale thermohaline circulation and SST anomalies. Special attention will be paid to monitoring variations in surface salinity and to monitoring transport in the western boundary of the Atlantic. Monitoring the strength of the thermohaline circulation requires a coordination with the WOCE program in making repeated hydrographic sections of 25°N and at a higher latitude nearer the source areas of the water mass.

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PALEO DEEP WATER VARIABILITY: MAGNITUDE AND RAPIDITY

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CARBON DIOXIDE, OCEAN CIRCULATION, AND CLIMATE: THE PROBLEM

The record of climate change in ice cores and deep sea sediments (and a few other lines of evidence) provides the only information on what happens to the earth during large-scale climate changes -- changes that are comparable in magnitude to model predictions of future climate change.

Ice core data show a surprisingly close coupling between past variations in CO_2 and climate, but we still cannot resolve the "chicken-egg" problem of cause and effect. However, we do know that atmospheric CO_2 fluctuations are driven by the ocean: by physical circulation, and biogeochemical cycling.

There are several deep ocean paleoceanographic tracers that we can use to study ocean climate changes:

1. $\delta^{18}\text{O}$, a function of $\delta^{18}\text{O}$ of water and temperature, that can reveal temperature gradients,
2. $\delta^{13}\text{C}$, in part, a tracer of dissolved ΣCO_2 , showing organic remineralization, and hence, deepwater flow patterns,
3. Cd/Ca , a phosphorus-analogue, again revealing organic remineralization and deepwater flow patterns,
4. Ba/Ca , a "deep regeneration" tracer, similar to Si or Alk that shows inorganic remineralization and deepwater flow patterns, and

5. The ^{14}C age difference between planktonic and benthic foraminifera, which reflects the ^{14}C "age" of the deepwater (deepwater ventilation rates).

Other tracers, such as advective tracers, e.g., Antarctic diatoms, are valuable adjuncts; other tracers are under development.

These data can be used in three types of models, namely "conceptual" box models, "inverse" box models, and 3-D ocean circulation models. In a later section, I discuss the advantages and disadvantages of each type.

BASIC CHANGES IN OCEAN CIRCULATION

The basic changes in ocean circulation and chemistry that have been documented during extreme climate change were those that took place during the last glacial maximum. Several basic facts about the circulation patterns and variability of the late Pleistocene ocean have been established by these data.

First, the North Atlantic Deep Water (NADW) flow was reduced significantly during glacial times (Fig. 1), but most glacial maxima have a continued source of nutrient-depleted water into the deep North Atlantic (on average, for the 2-4 thousand years represented by the typical bioturbated sediment sample): $\delta^{13}\text{C}$ and Cd in the Atlantic is always lower than in the Pacific (Figs. 2 and 3) (Boyle and Keigwin, 1982; Boyle and Keigwin, 1985/6; Duplessy et al., 1988; Curry et al., 1988). However, there are a few glacial extrema when deep NADW does appear to have been entirely terminated (Raymo et al., 1990).

As for many other climate characteristics, the temporal variability of NADW chemistry shows the hand of the Milankovitch orbital cycles, with statistically significant and consistent phase relationships (Boyle and Keigwin, 1985/6; Oppo et al., 1989; Raymo et al., 1990; Imbrie et al., (unpublished data)). But NADW chemical variability does not look exactly like any of the other climate indicators, such as ice volume. For example, during oxygen isotope stage 3, when the ice volume was 1/2-2/3 that of the glacial maximum, NADW flow appears to have been as high as it is today (Boyle and Keigwin, 1982, 1985/6; Mix and Fairbanks, 1985).

The radiocarbon age of the Atlantic at the last glacial maximum appears to have been slightly higher than today (about 650 years compared to 350 years today), but was always significantly lower than that of the Pacific (Broecker et al., in press).

The second basic fact to emerge is that the upper waters (<2500m) of the North Atlantic were more nutrient-depleted than they are now; the cause of this depletion is controversial. It was first suggested that this nutrient depletion was due to a greatly enhanced North Atlantic Intermediate Water flow (Boyle and Keigwin, 1987; Duplessy et al., 1988); others have suggested that an enhanced proportion of Mediterranean Outflow Water was

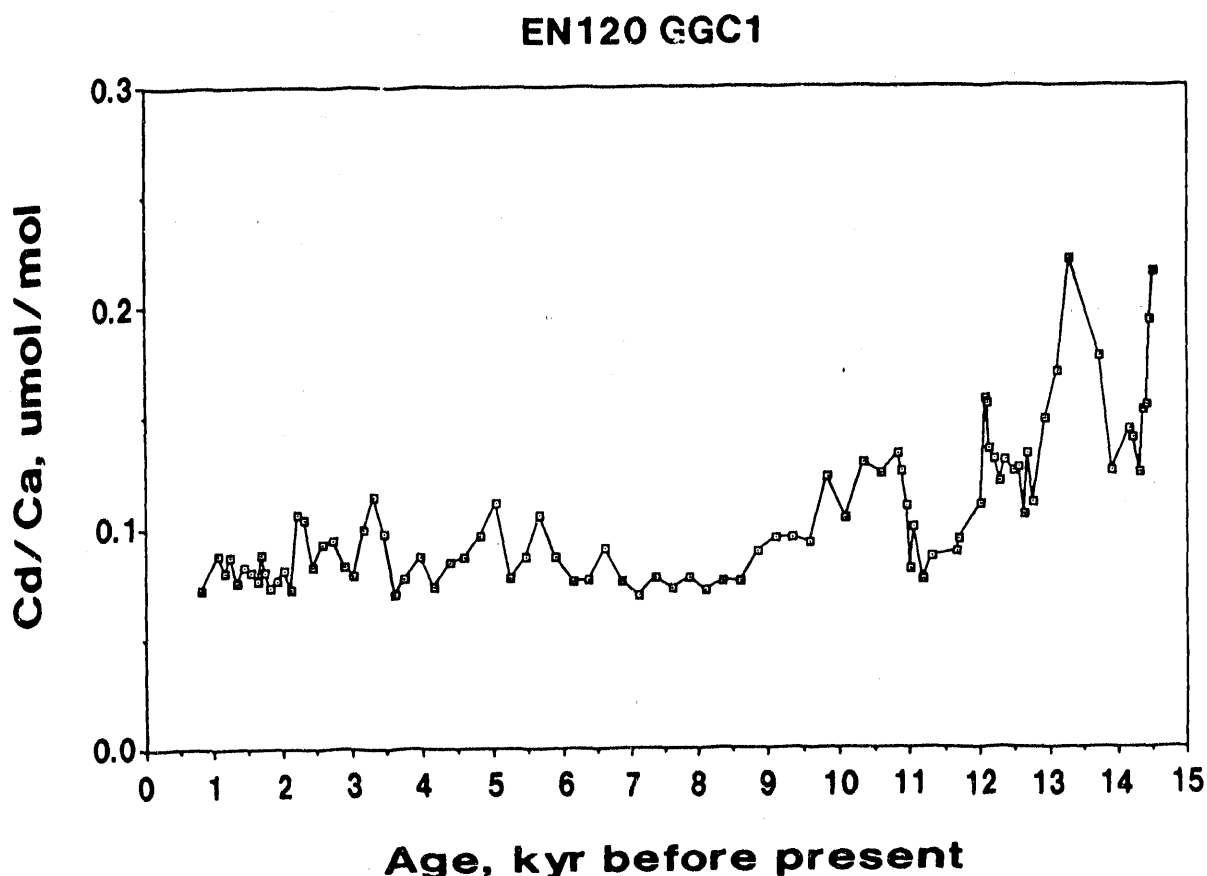


Fig. 1. Changes in the phosphorus-analogue Cd during the last 15,000 years in the deep (4400m) western North Atlantic. Note the higher Cd (i.e., less NADW) during the Younger Dryas (11,500-10,500) and glacial times.

responsible (Zahn et al., 1988; Oppo and Fairbanks, 1987). It is also possible that a decreased supply (or more nutrient-depleted input) of Antarctic Intermediate Waters (AAIW) contributes (Figs. 4 and 5).

Third, it appears that the upper waters of other ocean basins also were more nutrient-depleted than they are today. This was first suggested on the basis of $\delta^{13}\text{C}$ data (Duplessy et al., 1988; Kallel et al., 1988). Our new Cd data tend to support this interpretation (Fig. 6).

Fourth, the deep ocean was cooler (by 1.5-3°C) during most of the late Pleistocene, warming to its present values during the warmest interglacial periods (Shackleton and Chappel, 1986; Shackleton, 1987; Labeyrie et al., 1987).

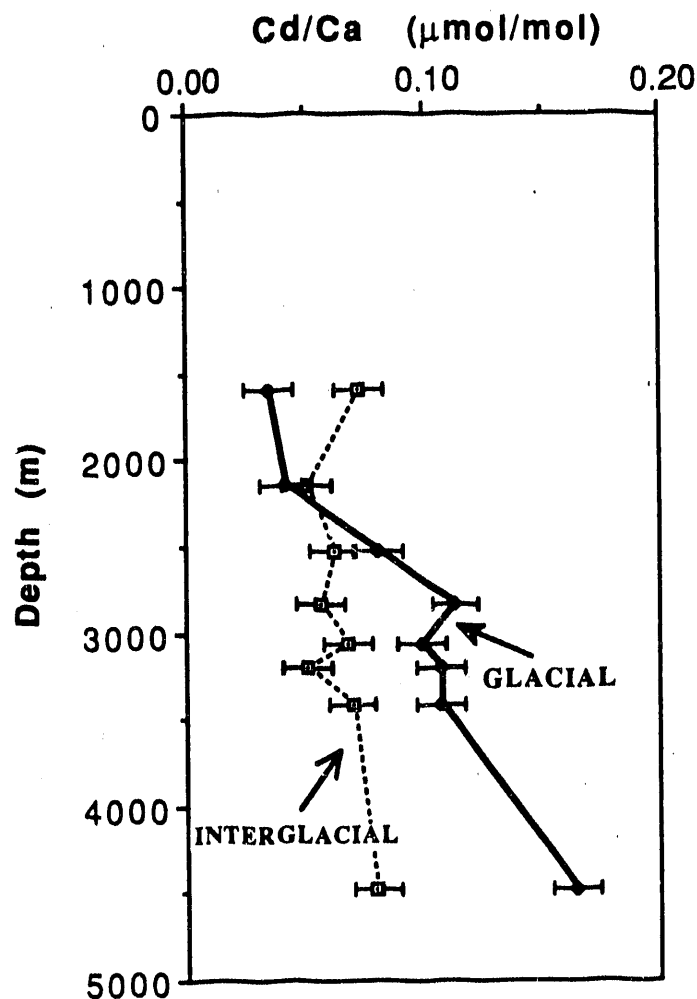


Fig. 2. Vertical profile of phosphorus-analogue Cd in modern and glacial western North Atlantic. Note the major change in vertical distribution.

The final fact that was established is that the radiocarbon age of the deep Pacific was about the same or slightly older during glacial times than it is today (1600 yrs.) (Broecker et al., in press; Shackleton et al., 1988). In view of the dramatic climate changes associated with glacial climate (in particular, major changes in NADW circulation), this subtle whole-ocean change is a surprising but important constraint on models of glacial deep ocean circulation.

HOW RAPIDLY CAN OCEAN CLIMATE CHANGE?

Deep sea sediment and ice cores have shown that major climatic changes can occur on a century time scale (Fig. 7). Thus, the Greenland Ice core

showed rapid changes in variability during glacial times, especially during deglaciations. In the northern North Atlantic there have been large-scale temperature fluctuations in the time span of a few centuries or less (Fig. 8). Such changes appear to be linked to equally rapid changes in the NADW.

Rapid variations in the rate of erosion and sediment transport have been found in the western North Atlantic. In the Sargasso Sea, there were rapid changes in salinity during deglaciation; it has been suggested that these changes may reflect meltwater events and fluctuations in deepwater.

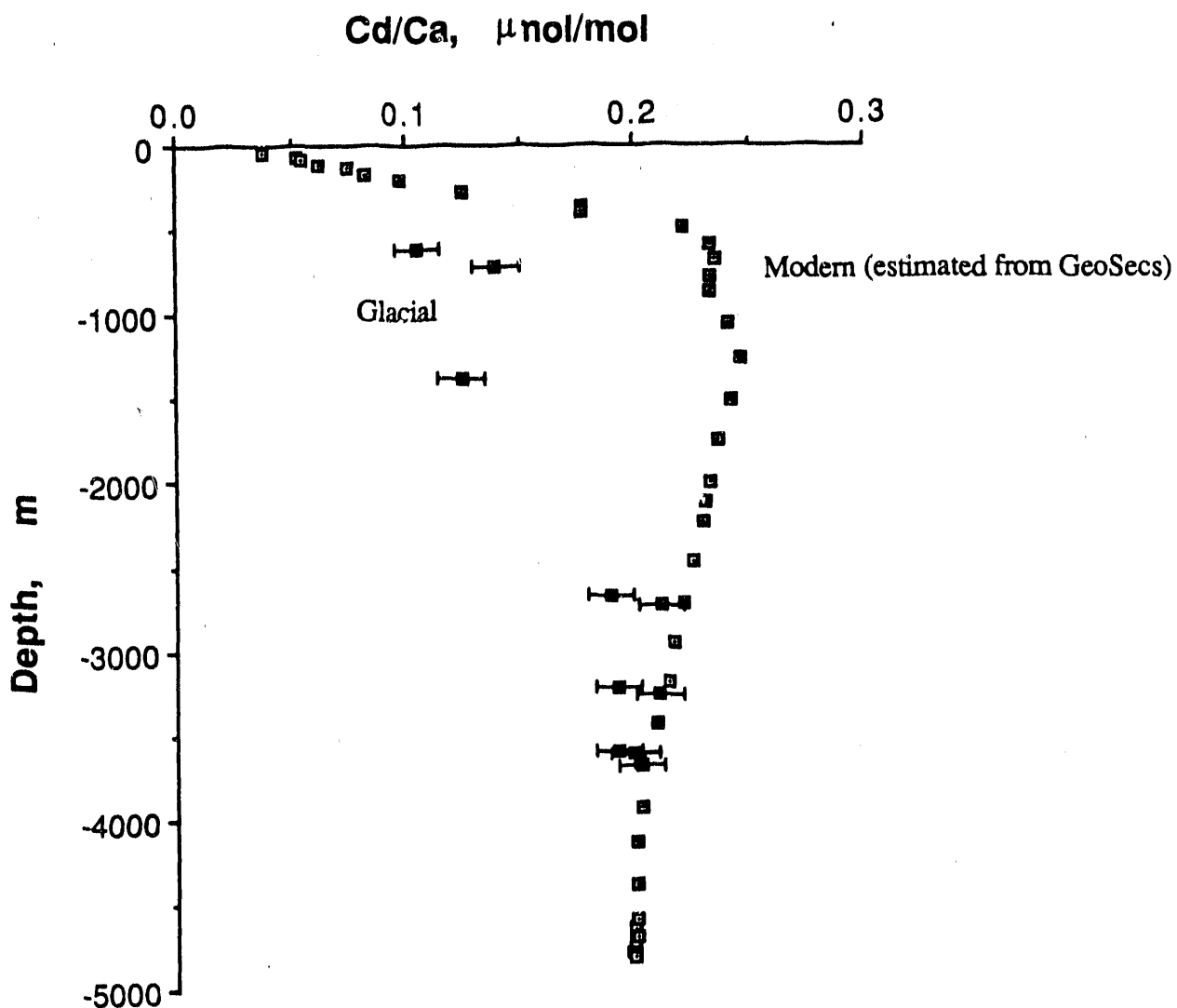


Fig. 3. Vertical profile of phosphorus-analogue Cd in the glacial eastern tropical Pacific. Note the major change in vertical distribution.

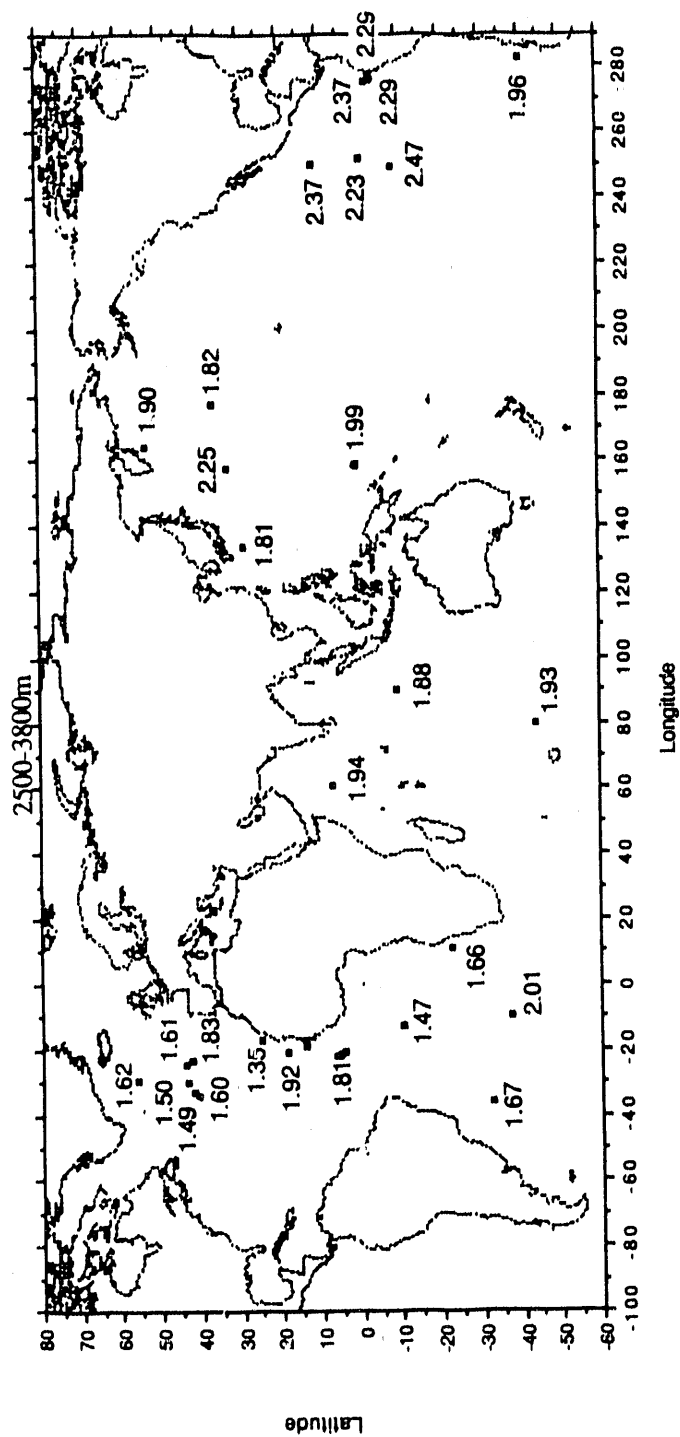


Fig. 4. Estimated phosphorus distribution 18,000 years ago at mid-depths. Note that glacial phosphorus is higher in the Atlantic than it is now due to NADW reduction.

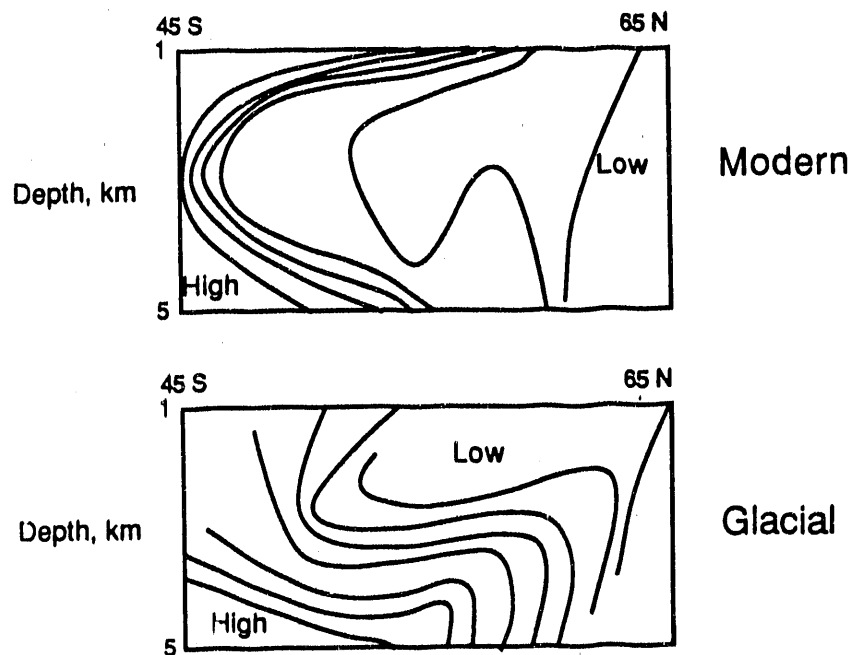


Fig. 5. Schematic section showing phosphorus distribution in the modern and glacial Atlantic Ocean. Note the large changes in intermediate and deep phosphorus.

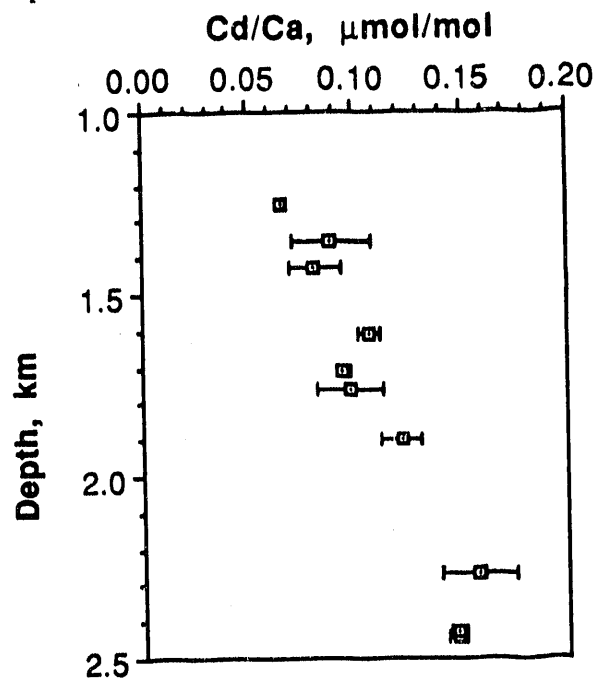


Fig. 6. Vertical distribution of phosphorus-analogue Cd in the northern Indian Ocean. The modern ocean would have the highest Cd in shallower waters.

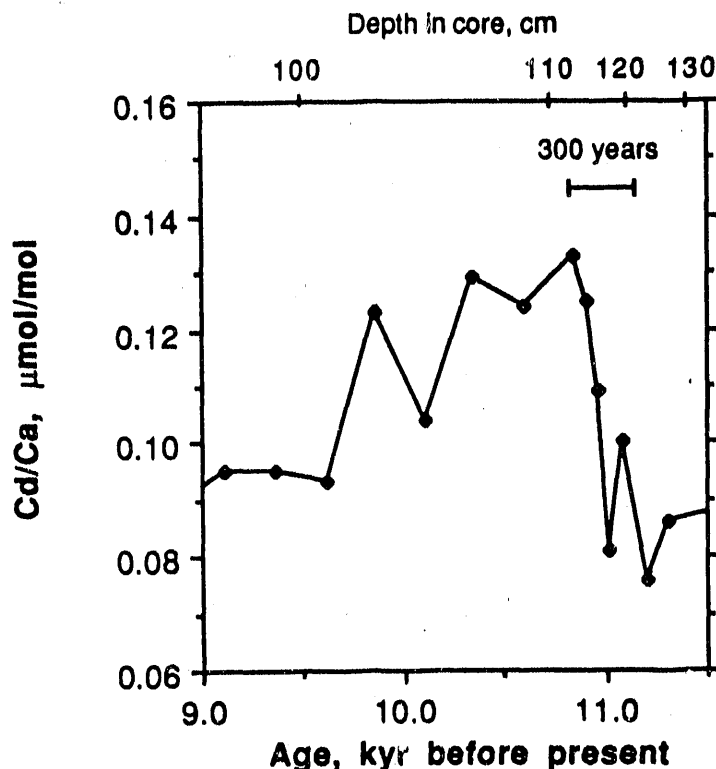


Fig. 7. Expansion of Younger Dryas section of Fig. 1. Note rapid (<300 yr) increase of Cd at initiation of Younger Dryas.

MODELS FOR OCEAN CLIMATE CHANGE AND THE CARBON SYSTEM

"Conceptual box models" are usually very simple (a few boxes), not very realistic, but adequate for illustrating processes and their sensitivity. These types of models provide for the chemical climate insight similar to that which generalizations such as geostrophy and conservation of potential vorticity provide for physical processes.

Attempts to explain glacial CO₂ cycles through such models have been extremely productive in providing insight into the key parameters of the carbon systems that might be important. Surprisingly, these simple models are continuing to provide new insight; new ideas have emerged in the last couple of years, and it would be naive to think that the insights have been exhausted. However, one problem with these models is that they often focus on the smallest number of parameters necessary to illustrate the effect on CO₂ and do not directly incorporate all of the relevant constraints. Another problem is that they function on a trial-and-error, forward-calculation basis; hence, people manipulate the models until they seem to agree with the facts; it is difficult then to be sure that the full range of potential solutions has been found.

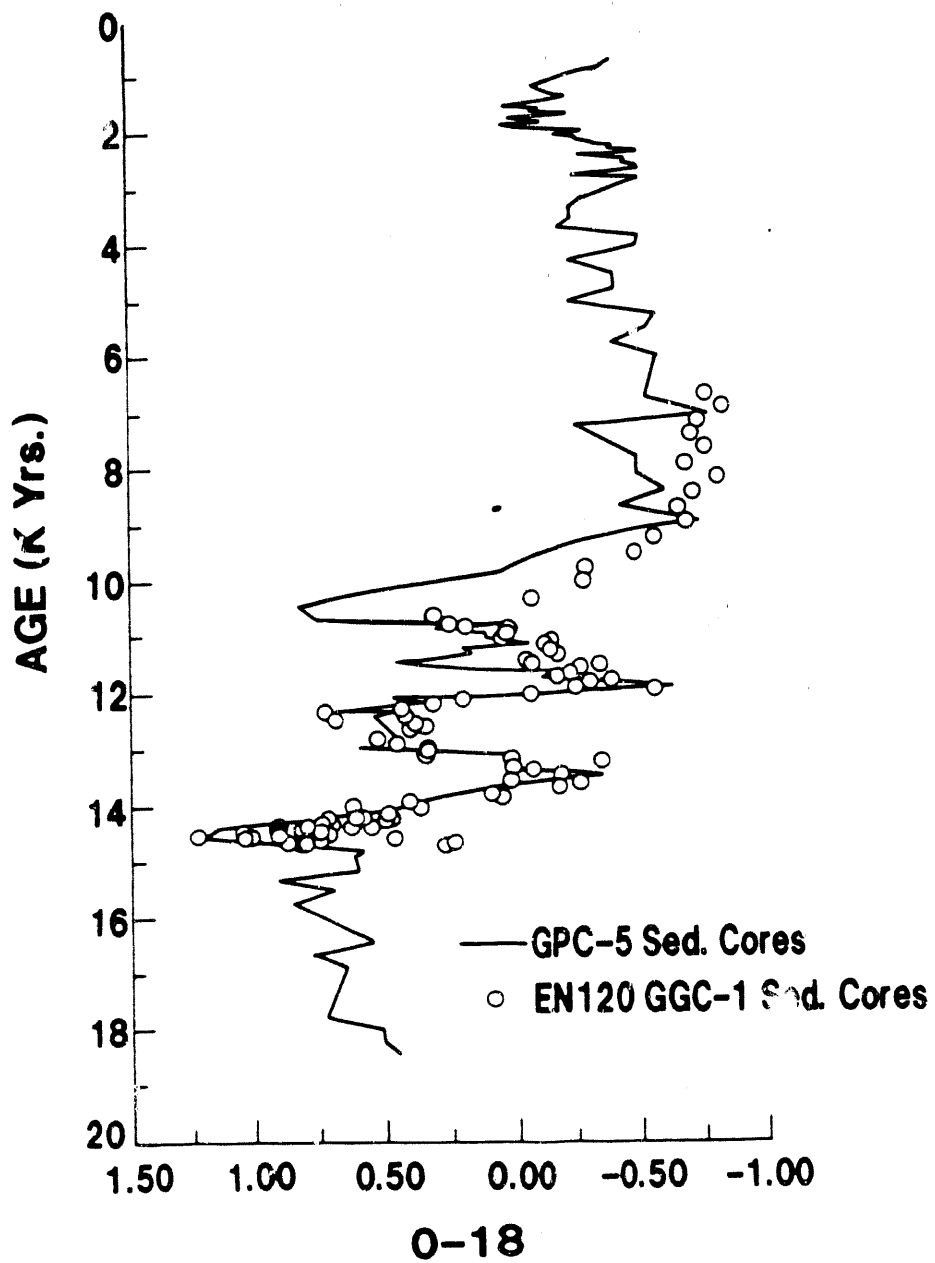


Fig. 8. Rapid variations in surface water properties (T,S) in the western North Atlantic during the past 18,000 years. Note rapid variability. Data of Lloyd Kelgin, WHO.

"Inverse" models have mainly been used so far for clarifying circulation problems in the modern oceans. The basic math and other requirements are still being worked out.

So far, inverse models have not been used for paleo-inverse modeling. Before this can be done, significant problems need to be worked out to determine how to cope with the great reduction in constraints (e.g., geostrophy is a very weak constraint at the glacial maximum). However, a great benefit is that inverse models can be made to incorporate all of the diverse quantitative and qualitative types of information that are available.

In addition, inverse models allow some basic questions to be asked without any manipulation of the model. Such important questions that we may pose are the following ones.

1. What is the minimal change in ocean circulation and chemistry that is necessary to account for the observations?
2. What is the maximum change that might be consistent with the data?
3. What should new-data be obtained to provide the maximum level of benefit in constraining a solution?

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ROLE OF THE OCEAN CARBON CYCLE IN DETERMINING ATMOSPHERIC $p\text{CO}_2$

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OCEAN UPTAKE OF FOSSIL CO_2

Ocean biological interactions affect climate indirectly through the removal of CO_2 as organic carbon from the upper ocean and by its regeneration at depth. The effect of this "biological pump" on atmospheric $p\text{CO}_2$ is considerable: if the pump were to cease operating, the atmospheric $p\text{CO}_2$ would increase from its pre-industrial value of ~280 ppm to ~530 ppm (Baes and Killough, 1986) with an enfolding time-scale of about 160 years (Siegenthaler and Wenk, 1984). On geologic time scales, the burial of carbon in ocean sediments would be affected, with even greater consequences for atmospheric $p\text{CO}_2$. However, until recently, simple ocean models had led to the conclusion that changes in the natural cycling of carbon over short enough time scales to affect anthropogenic perturbation were very unlikely. Thus, all estimates from models of fossil CO_2 uptake by the oceans are based on the assumption that the natural carbon cycle is in steady state. The observations of ice cores by Siegenthaler et al. (1988), showing that atmospheric $p\text{CO}_2$ was remarkably steady within about 10 ppm for about 1000 years before the beginning of the industrial revolution, suggest that this viewpoint is not without basis.

Table 1 gives the global CO_2 budget for 1980-1989 as estimated by the Intergovernmental Panel on Climate Change (1990). All steady-state models of the ocean give oceanic uptake rates which are consistent with a fossil CO_2 uptake of the order of 2 GtC/yr for 1980-1989. Estimates of the oceanic uptake based on the sparse ocean data of Takahashi as reported on in Tans et al. (1990), give an uptake of the order of 1.6 GtC/yr that is consistent with the model-based estimates. These estimates of oceanic uptake of fossil CO_2 recently received a major challenge from an analysis based on atmospheric

observations and a model of atmospheric CO₂ transport (Tans et al., 1990). Tans's study suggests that the inter-hemispheric transport of CO₂ from its primarily Northern Hemisphere source to the Southern Hemisphere is not large enough to support a Southern Hemisphere sink of fossil CO₂ in the ocean. Tans et al. conclude that the oceanic uptake for 1981-1987 must have been primarily in the Northern Hemisphere. An estimate of less than 1 GtC/yr for the oceanic uptake in the Northern Hemisphere is provided by the data of Takahashi reported in this paper. Takahashi's estimates of Southern Hemisphere uptake, which are very poorly constrained, are ignored, bringing the IPCC (1990) budget even further out of balance. Tans et al. (1990) propose that there is a large Northern Hemisphere terrestrial sink to close the balance.

Keeling et al.'s (1989b) study of the atmospheric CO₂ budget, which uses the same atmospheric transport model as Tans et al. (1990), comes to the same conclusion on the inter-hemispheric transport of CO₂. However, they further use the ¹³C data to show that it is unlikely that the terrestrial sink can be as large as was suggested by Tans et al. Keeling et al. conclude that the oceanic sink is of the same order of magnitude as proposed in the IPCC (1990) summary, but that it must be primarily in the Northern Hemisphere. The inconsistency between their conclusion and the observations of Takahashi, as summarized by Tans et al. (1990), remains to be resolved.

Together, these results suggest that our knowledge of the oceanic carbon cycle may not be as good as was suggested by the IPCC (1990) summary (Table 1). In the remainder of this section, I suggest several alternative explanations for the results of Keeling et al. (1989b) and Tans et al. (1989), and recommend measurements and model studies that would help to test my ideas.

One possibility is that the Goddard Institute for Space Studies (GISS) atmosphere model used by Keeling et al. (1989b) and Tans et al. (1990) does not properly represent the inter-hemispheric transport of CO₂. It is well known that the inter-hemispheric exchange in the GISS model is far too low, probably because of having too low a horizontal resolution to properly represent equatorial processes. This deficiency has been remedied by increased diffusivity, which was calibrated with krypton-85 and chlorofluorocarbon observations. However, the behavior of CO₂ is likely to be more complex than can be represented by these tracers because of its much larger seasonal excursions. Clearly, further studies of the models are warranted. Among other things, one of the important aspects of the models for which there is little data is the vertical attenuation of the CO₂ mixing ratio in the atmosphere.

If we accept that the Southern Hemisphere ocean is not taking up CO₂, an alternative explanation offered by Keeling et al. (1989b) is that the entire 2 GtC/yr uptake predicted by the ocean models is being taken up in the Northern Hemisphere. This explanation violates Takahashi's estimates that were based on field measurements; however, the data base of measurements of the air-sea CO₂ gradient and our knowledge of the gas exchange rate are problematic enough to harbor considerable uncertainty. We need to develop strategies for measurements that would help to resolve the air-sea CO₂

Table 1. Budget for CO₂ Perturbation
1980 to 1989 (from IPCC, 1990)

Perturbation	Average (1 Gt C = 10 ¹² kg C)
SOURCES	
(1) Fossil	5.4 ± 0.5
(2) Deforestation	1.6 ± 1.0
TOTAL	7.0 ± 1.2
SINKS	
(1) Atmosphere	3.4 ± 0.2
(2) Oceans (Steady-State Models)	2.0 ± 0.8
TOTAL	5.4 ± 0.8
SINKS UNACCOUNTED FOR	1.6 ± 1.4

gradient in the Southern and Northern Hemisphere. Above all, we need to ascertain the sensitivity of gas exchange to wind speed, for which, at present, there is an uncertainty of more than a factor of two between wind tunnel and many tracer-based estimates at the low end, and radiocarbon-based estimates at the high end (Thomas et al., 1988). Keeling et al. (1989b) point out that we are particularly ignorant of the extent of air-sea CO₂ exchange in the North Pacific, where it is likely to be low and, thus, more difficult to measure.

Three-dimensional primitive-equation ocean models of the perturbation CO₂ suggest that the Southern Hemisphere oceans should be taking up of the order of 1.2 GtC/yr of the perturbation (Sarmiento et al., unpublished). If this uptake is not occurring through the air-sea interface, it must take place by oceanic transport from the Northern to the Southern Hemisphere, most likely in the North Atlantic Deep Water, and possibly by a flux of biogenic material from the surface into the southward-flowing Pacific Deep Water, and possibly by a flux of biogenic material from the surface into the southward-flowing Pacific Deep Water. Ocean models of the pre-industrial ocean carbon cycle, such as those of Maier-Reimer and Hasselman (1987) and Bacastow and Maier-Reimer (1990), should be able to show such an inter-hemispheric oceanic transport. Direct estimates of oceanic transport using oceanic hydrographic sections such as those undertaken by Brewer et al. (1989), should also be able to show this transfer. The measurement of parameters of the carbon system on World Ocean Circulation Experiment (WOCE) sections will make an important contribution here. The communication between hemispheres in the

deep ocean is much slower than the rate of CO₂ increase. Thus, it is likely that the southward ocean transport of CO₂, if it exists, was functioning before the anthropogenic perturbation, and, in the pre-industrial era southward transport must have been balanced by a northward transport in the atmosphere. Observations of the Mauna Loa/South Pole CO₂ gradient by Keeling et al. (1989a) support such a view by showing that the pCO₂, extrapolated back to pre-industrial times, was 0.82 ppm higher at the South Pole than at Mauna Loa.

The WOCE sections will also provide a data base to compare with older and future data bases to directly estimate the change in the ocean inventory of carbon. An alternative to assessing oceanic CO₂ uptake by comparing measurements made at different times is to estimate the total uptake by subtracting an estimate for the pre-industrial CO₂ content, as was suggested by Brewer (1978) and Chen and Millero (1979). However, this technique requires precise estimates of the pre-industrial, pre-formed total carbon, and the addition to this total from regeneration of organic carbon. The uncertainties in such estimates are too great to do anything more than provide a broad brush overview (Shiller, 1981; Broecker et al., 1985).

IMPACT OF PERTURBATIONS OF THE OCEAN CARBON CYCLE

Another point of view on the ocean carbon cycle arises from the discovery, from measurements of trapped air bubbles in glacial ice, that atmospheric pCO₂ levels during the last ice age were as low as ~200 ppm (Neftel et al., 1982). Efforts to explain the ice age pCO₂ phenomenon have led to a greatly improved understanding of the dynamics of the ocean carbon cycle and its relevance to "global change" (Sarmiento et al., 1988). There are mechanisms by which changes in the natural cycling of carbon in the oceans may lead to rapid changes in the level of atmospheric CO₂, which are of the same order of magnitude as the anthropogenically induced atmospheric transient; these mechanisms might also serve to explain the natural changes observed in ice cores.

The major focus of such studies has been on the effect of changes in regions of deep-water formation, particularly around the Antarctic. These regions are the windows through which the deep ocean "sees" the atmosphere, unencumbered by the insulating thermocline. Thus, they are the places through which the excess SO₂ stored in the deep ocean can escape to the atmosphere. Also, the vast amount of excess nutrients found at the surface in regions of deep-water formation at present, and the rapid replacement of these nutrients by exchange with waters below, provides a potential for a significant increase in biological uptake of CO₂ that can lead to a decrease in atmospheric pCO₂. Box-model studies suggest that changes in the balance between vertical exchange, which supplies nutrients and CO₂ to the surface from below, and biological uptake, which removes nutrients and CO₂, may lead to pCO₂ levels ranging from approximately 165 ppm for the total depletion of nutrients along with an amount of CO₂ determined by the average composition of organic matter, to 425 ppm when there is no removal. These values can be compared to the pre-industrial value of 265 ppm used in the early

study of Sarmiento and Toggweiler (1984; see also the simultaneous and independent studies of Knox and McElroy, 1984, and Siegenthaler and Wenk, 1984). These same box models show that the response to a perturbation in the biological uptake can initially be very large, of the order of 1 ppm/year, which is of the same order as the present increase in atmospheric CO_2 (Siegenthaler, personal communication).

Our understanding of the relevant physical and biological processes is abysmal. Regarding the physical processes, the box models depict the deep-water formation regions in high latitudes by a single, well-mixed surface box sitting on top of a single, well-mixed box representing the entire deep ocean (the low-latitude surface is represented by another single box). A first attempt to represent these high-latitude regions by a diffusive region sitting on top of a well-mixed, deep-ocean box results in a remarkable decrease in the sensitivity of the CO_2 to changes in the biological uptake, from about 100 ppm for the box models of Sarmiento and Toggweiler (1984) and others, to a few 10s of ppm (W.S. Broecker and T.-H. Peng, personal communication). The diffusive region reduces the rapid communication between the high-latitude surface with the deep ocean in the box models of Sarmiento and Toggweiler (1984). Such a reduction is suggested by the low radiocarbon concentrations from atomic bombs that were found in the Southern Ocean deep waters by GEOSECS. The new, diffuse model appears to trap much of the nutrients and CO_2 , removed from the high-latitude surface in the diffusive zone immediately below. The potential impact of a change in high-latitude processes thus appears to be extremely sensitive to how the physical processes are characterized. Clearly, we need to learn more about the physical processes by direct measurement as well as by more studies of sophisticated models. More realistic models of primitive equation ocean general circulation already exist in which the above sensitivity studies need to be carried out with great urgency. However, even in these models the simulation of the processes of deep water formation is problematic. We cannot escape the conclusion that we need to learn more about processes occurring in regions of deep-water formation by making, among other things, measurements of tracers that will tell us how rapid the exchange is and show us which waters are involved in that exchange.

The other problem in predicting the potential impact of changes in the ocean carbon cycle is our extremely poor understanding of what controls the biological uptake of nutrients and CO_2 . The aforementioned sensitivity studies blithely change the efficiency of biological uptake with no regard as to whether such changes are realistic. There is strong evidence from measurements of pCO_2 in crapped air bubbles in ice cores that atmospheric pCO_2 did not vary by more than 10 ppm around its pre-industrial value of ~280 ppm for at least a millennium before the industrial revolution (Siegenthaler et al., 1988). Why has this steady state prevailed over such a long period? In the vast regions of the oceans where nitrate and phosphate are depleted it is usually assumed that nutrient concentration limits photosynthesis, and, therefore, that biological uptake will adjust to any changes in supply rate so as to keep the nutrients at very low concentrations at all times. Since the biological removal of nutrients from the surface and regeneration at depth are linked to removal and regeneration of carbon by the Redfield ratio,

the CO₂ concentration should also remain essentially constant, assuming the Redfield ratio of organic matter formation and regeneration are unchanged.

However, there are large regions of the ocean where nutrients are not depleted, the most notable being the northeastern Pacific, eastern Equatorial Pacific, and Antarctic. It is apparent in these regions that something other than nutrient supply is controlling new production. It has often been postulated that the ecosystem structure plays a role (Evans and Parlsow, 1985; Frost, 1987), but the recent series of studies by Martin and his colleagues (Martin and Fitzwater, 1988; Martin and Gordon, 1988; Martin et al., 1990) provide evidence that iron deficiency may limit phytoplankton growth in the northeastern Pacific and Antarctic. Over the last millennium, the supply of iron to the surface ocean, or the ecosystem structure, or whatever it is that controls the surface nutrient and carbon content of these regions, appears to have kept in step with the upward supply of nutrients and carbon, so that the average CO₂ content of the surface ocean has remained constant.

It seems likely that this balance will not continue into the future. The changes in atmospheric pCO₂ that occurred during the last ice age show that the ocean carbon cycle has not always been in steady state. These changes have been ascribed to variations in ocean circulation or ocean biology in the high-latitude regions of deep-water formation such as those discussed above; other possible mechanisms have been suggested such as a reorganization of the nutrient distribution due to changes in productivity in low latitudes (Boyle, 1988), or to an increase in the total amount of dissolved organic carbon (DOC) due to a reduction of the rate at which DOC is metabolized as the water temperature cooled (Brewer, personal communication). Greenhouse warming will almost certainly result in changes in ocean circulation (Bryan and Spelman, 1985; Manabe and Stouffer, 1988; Stouffer et al., 1989) which will not only affect the direct uptake of anthropogenic CO₂, but will also very likely disturb the balance of the natural carbon cycle, thus impacting atmospheric pCO₂ indirectly. Clearly, we need to know more about what factors limit the biological uptake of nutrients and carbon in the upper ocean, particularly in regions of deep-water formation, as well as what processes control the cycling of dissolved organic matter in the oceans.

The possibility that iron is limiting the full utilization of nutrients led to the suggestion that iron might be used to fertilize the oceans as a way of mitigating the atmospheric increase of CO₂ from anthropogenic sources (J.H. Martin, S.E. Fitzwater and R.M. Gordon, personal communication). The possibility of phosphate fertilization in phosphate-poor areas of the subtropical gyres also has been suggested. The impact of such fertilization on ocean biology and atmospheric pCO₂ has not been determined, although initial studies suggest that an upper limit to the possible effect is of the same order of magnitude as the present pCO₂ increase in the atmosphere. Model studies are underway to improve these estimates, but much more effort is needed to understand the relevant physical and biological processes.

A discussion of global climate feedbacks from ocean biological interactions cannot be considered complete without mentioning several other possible feedbacks that have received attention in the literature. One is the potential for a dimethyl sulfide (DMS)/cloud feedback. Coccolithophores and many dinoflagellates as well as other chromophyte algae release significant DMS (Keller et al., 1989) which may then escape to the atmosphere and form condensation nuclei for cloud formation (e.g., Bates et al., 1987). Might changes in this process lead to significant climate feedback effects, as has been suggested by Charlson et al. (1987)? Schwartz (1988) argued that the lack of any clear connection between recent increases in manmade SO_2 emissions and climate changes gives evidence to the contrary, but this point of view is disputed by the recent work of Savijoe and Prospero (1989) as well as Wigley (1989).

Another possible feedback is through the effect of ultraviolet radiation on marine life (e.g., Smith and Baker, 1989; and Hardy and Gucinski, 1989). UV-B (280-320 nm) radiation causes significant damage to marine life. A 16% reduction of ozone increases UV-B by 47% at the ocean surface. However, the effect of UV-B is very shallow. The vertical attenuation coefficient weighted by DNA damage is about 0.4 m^{-1} (Smith and Baker, 1979), i.e., a factor of 10 reduction in 10 m. Would this have a significant impact on oceanic uptake of nutrients and carbon, or does most of the photosynthesis occur deep enough that the impact of increased UV-B would be minimal? It should be noted that the impact of decreased ozone will be greatest in the high-latitude regions where deep water formation occurs.

Finally, there has been considerable interest in the possible role of organic carbon export from the shelves to the deep waters of the slope as a sink for fossil CO_2 (e.g., Walsh, 1989). The suggestion made by Walsh is that measurement-based estimates of CO_2 exchange at the air-sea interface such as those of Takahashi reported on in Tans et al. (1990) are missing a large net sink of the order of 1 GtC/yr in continental shelves. This sink is due to carbon which is taken up by organisms and then exported off the shelf before being regenerated. Falkowski (1988) reports that recent measurements off the coast of Long Island, New York, do not support such an interpretation. These measurements show a large in situ sink due to zooplankton grazing that was missed in previous studies, as well as suggesting a large, additional microbial sink based on observations of the oxygen budget.

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COMMENTS ON THE PROBLEM OF THE OCEAN CIRCULATION

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In general terms, one must be aware that no matter how sophisticated the meteorological models become over the next decade, the present comparatively primitive state of understanding of the oceans must be overcome. Otherwise, any future forecast of climate change, whether owing to greenhouse effects or other causes, will be vulnerable to the sweeping and devastating criticism that the oceanic component is not believable, and hence, that no climate forecast has any credibility. The geological record clearly shows that the ocean circulation was radically different in the past, when the climate was radically different. The problem is that we lack the understanding to model the circulation of the present ocean adequately - much less to forecast it in any convincing way.

The present difficulty has many root causes. Many of them are traceable to the sheer difficulty of observing the system: the ocean is opaque to electromagnetic radiation; it is very large, as measured by the speed with which ships can cross it; it is turbulent (i.e., highly time-dependent with strong inter-scale linkages); and governments have never seen a need to construct a proper observational system. The difficulties are compounded by the widespread presence of certain evolving oceanic myths - myths which have arisen, I suspect, largely through wishful thinking. Among those myths I would include the supposition that one only has to worry about seasurface temperature. This myth gradually was replaced by the argument that maybe only the mixed layer is of importance. Now we see the WCRP arguing that only the top 1000m is involved. One wonders how long it will be before the recognition, that manmade tracers appear in the abyssal ocean on a timescale of decades, will penetrate the collective consciousness with the conclusion that in a three-dimensional turbulent fluid there are no defensible arbitrary interior boundaries.

Existing pictures of the ocean circulation are based upon treating the system as though it were in a steady state, with all observations lumped together in a geological-like view. The literature is replete with papers expressing astonishment that some element of the system seems to be changing. But, to the contrary, the discovery that some element of a turbulent fluid system were rock steady would be an enormously powerful simplifying (but astonishing) discovery. Corresponding dynamical studies of the system with models are usually based upon driving by atmospheric climatologies, which are hardly representative of actual over-ocean conditions.

Anyone who is serious about realistic climate forecasting must face up to these problems if the forecasts are ever to be taken seriously. How might these problems be overcome? There are several elements to a solution, but I believe they lie mainly in being clever about employing the technologies that are now available. Indeed, the past decade has seen the emergence of technical solutions for observing the ocean on climatic scales which could be at the heart of a comprehensive oceanographic data base. These techniques include (but are not limited to) the emergence of unmanned profiling/submersible devices - which remove the severely limiting reliance on expensive, labor-intensive ships. Altimeters and scatterometers provide truly global measurements both of the oceanic dynamics, and of the major coupling (the windfield) between the ocean and the atmosphere. We have potentially cheap unmanned drifters (the ALACE floats), climate-scale integral capability (tomography), and the potential for great improvement in modeling both of the ocean, and of the atmospheric boundary layer.

The question really is whether we are politically astute enough to put such a system into place. The present situation in which the major global oceanographic observing programs, which have rational strategies for achieving the necessary understanding of the ocean, are being financially strangled even while governments speak of wondrous intentions, leads one to a pessimistic conclusion.

ELEMENTS OF OCEANIC CO₂ FEEDBACK IN RESPONSE TO CLIMATIC CHANGES

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INTRODUCTION

The upper mixed layer (75 meters thick on the global average) of the oceans contains an amount of CO₂ nearly equal to that in the atmosphere (6.2×10^{16} moles of CO₂), and the subsurface waters below it (about 3900 meters thick on the global average) contain about 50 times as much CO₂ as the atmosphere. Fig. 1 shows the concentration of CO₂ in seawater at various depths in the major ocean basins (Takahashi et al., 1981). The subsurface waters are highly supersaturated with respect to atmospheric CO₂: they would exhibit a pCO₂ value about three times as high as the atmospheric value if they were warmed to the mean surface water temperature of about 18°C. This suggests that the storage capacity of CO₂ in the oceans and the transfer of CO₂ between the atmosphere and deep ocean waters are regulated by complex physical and biogeochemical processes occurring in the upper layers of the oceans. Climatic changes, including not only temperature changes, but also changes in precipitation/evaporation and wind regimes, would undoubtedly affect these processes in the upper ocean layers, and hence, affect the dynamics of air-sea CO₂ exchange processes. However, the regulatory mechanisms involved in the air-sea CO₂ exchange are complex, and our knowledge is so incomplete that even the direction of oceanic CO₂ feedback in response to the anticipated climatic warming and associated changes cannot be estimated with certainty. In this paper, some of the major variables and their possible effects on the dynamics of air-sea CO₂ transfer will be briefly discussed.

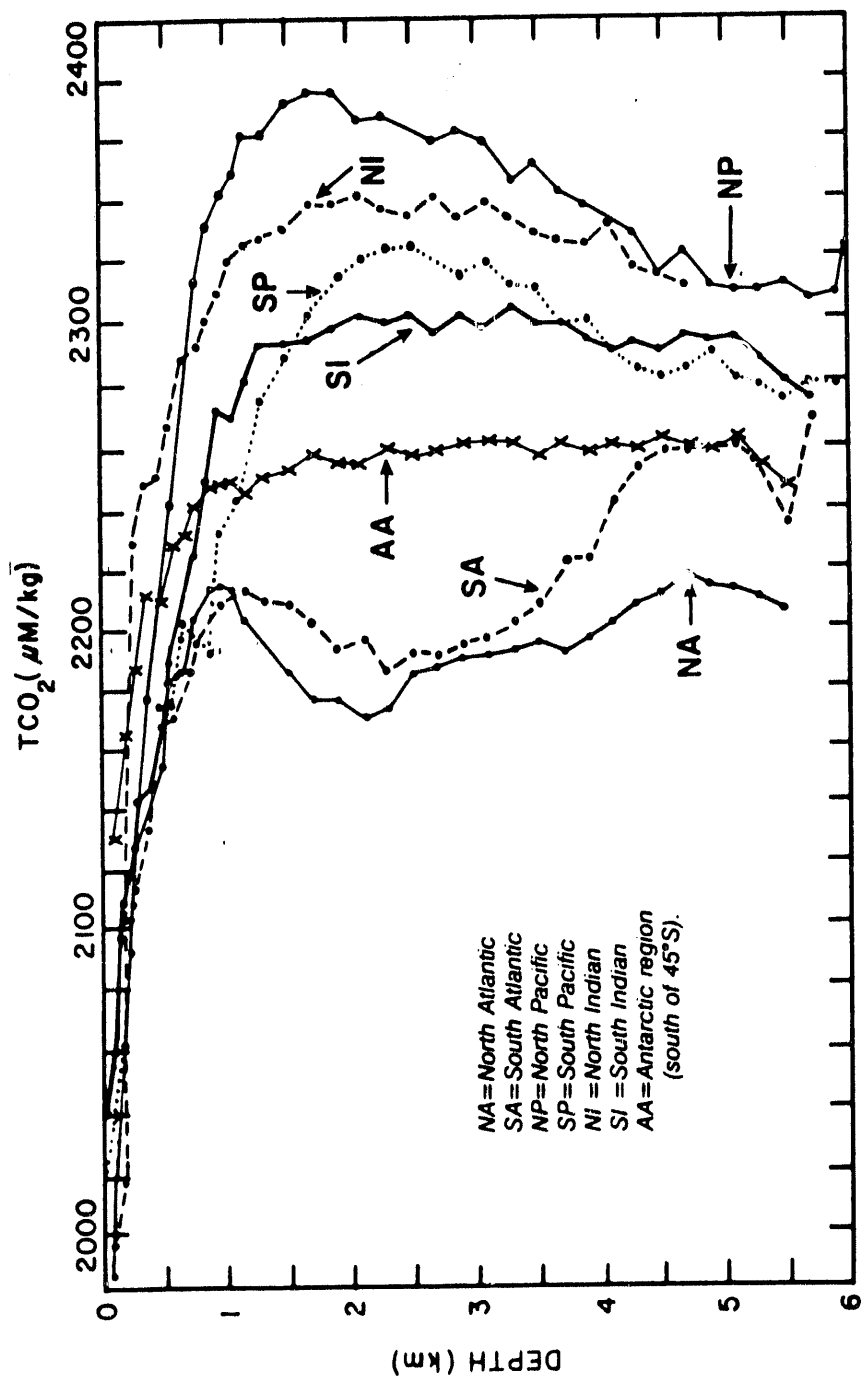


Fig. 1. The mean vertical distribution of the total concentration of CO_2 in the seven regions of the world oceans.

THERMOCHEMICAL PROPERTIES OF CO₂ IN SEAWATER

The partial pressure of CO₂ (pCO₂) of seawater increases at a rate of 4.3% per °C (pCO₂ doubles every 16° warming) with increasing temperature. Hence, warming of surface ocean water gives a positive feedback. The seawater pCO₂ increases nearly logarithmically as CO₂ is added: in a normal surface ocean water, a 1% increase in the total concentration of CO₂ in seawater causes a 10% increase in pCO₂. This ratio ($d \ln pCO_2 / d \ln TCO_2$), which is called the Revelle factor, increases rapidly with increasing concentrations of CO₂ (within a range of realistic oceanic concentration) as shown in Fig. 2 (Takahashi et al., 1980). This means that, as the ocean water takes up more CO₂ from the atmosphere, it takes up less CO₂ for a unit increase in atmospheric pCO₂. Hence, a lesser proportion of CO₂ released into the atmosphere would be dissolved in ocean water, and therefore, the airborne fraction of industrial CO₂ emission would increase. This increase would cause an additional positive feedback to the CO₂-induced warming.

OCEAN CIRCULATION

The vertical stability of the upper water column will be reduced when the density in the surface layer is increased by cooling or increased salinity (via evaporation or ice formation). This reduction will result in an increase in the transport into the surface water regime of deep waters rich in CO₂ and nutrient salts, thus causing an increase in the concentration of atmospheric CO₂. An increase in wind speeds also will enhance the vertical mixing in the upper layers, and the air-sea CO₂ gas exchange rate. On the other hand, warming or freshening of the surface water and calm winds would reduce the vertical mixing rate, thus decreasing the transfer rate of deep water CO₂ into the atmosphere. Consequently, the direction of the ocean CO₂ feedback in response to the anticipated global warming also depends upon other climatic forcing factors, such as precipitation and winds, associated with climatic changes. Furthermore, since these climatic variables are likely to change at differing rates and by differing magnitudes in various regions of the globe, it is not possible to predict even the direction of feedbacks of the oceanic CO₂ reservoir.

MARINE BIOLOGY

The total amount of new biological production is supported by the amount of nutrient salts supplied to the euphotic zone by the vertical mixing of deep waters rich in CO₂ and nutrient salts. In a steady state ocean, the amount of CO₂ supplied to the surface water regime by the vertical mixing is balanced by the amount of carbon transported down into the deep water regime via the gravitational settling of biogenic particulates and the advective transport of dissolved organic carbon (DOC). Thus, although an increase in the deep water upwelling rate in response to climatic changes would increase pCO₂ in the surface layer, the biological production, and hence, the downward transport rate of carbon by the biological pump also would increase to counteract the pCO₂ increase in the surface water.

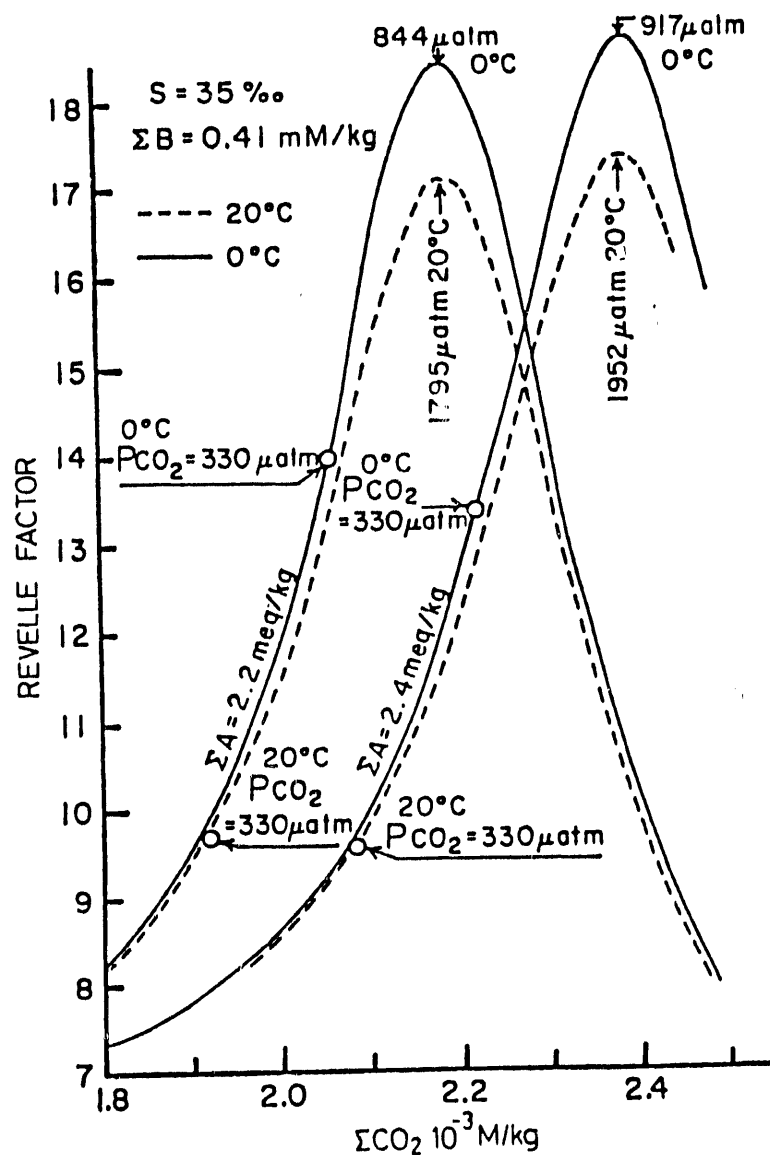


Fig. 2. Variation of the Revelle factor ($d \text{ pCO}_2 / \text{pCO}_2$) / ($d \text{ TCO}_2 / \text{TCO}_2$), calculated as a function of the total CO₂ concentration in seawater (35‰ salinity and 0.41 mM/kg total borate concentration) using the first and second apparent dissociation constants for carbonic acid determined by Mehrback et al. (1973), the first apparent dissociation constant for boric acid determined by Lyman (1956), and the solubility of CO₂ in seawater determined by Weiss (1974).

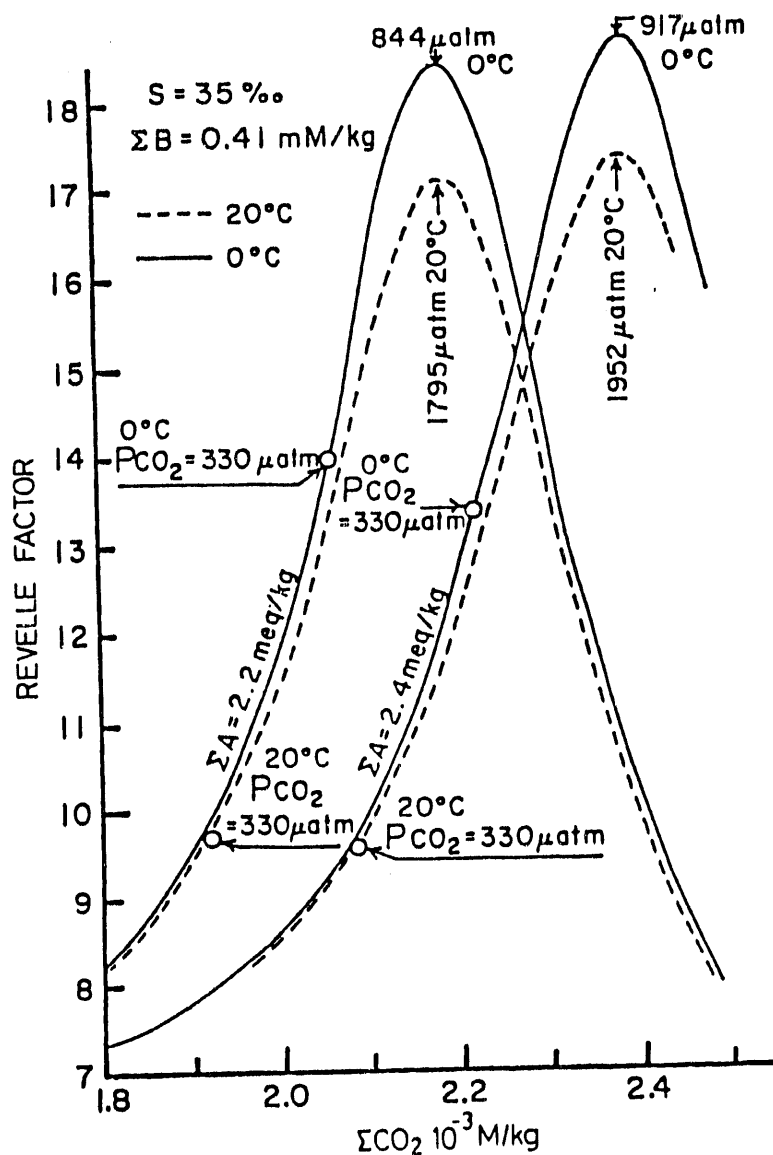


Fig. 2. Variation of the Revelle factor ($d p_{\text{CO}_2}/p_{\text{CO}_2})/(d \text{TCO}_2/\text{TCO}_2)$, calculated as a function of the total CO_2 concentration in seawater (35‰ salinity and 0.41 mM/kg total borate concentration) using the first and second apparent dissociation constants for carbonic acid determined by Mehrback et al. (1973), the first apparent dissociation constant for boric acid determined by Lyman (1956), and the solubility of CO_2 in seawater determined by Weiss (1974).

Therefore, the net effect of climatic changes on the CO₂ budget in the ocean-atmosphere system depends on the complex interactions between changes in the biological production and climate forcings.

Climatic changes also may affect the ecological assemblage of calcareous (CaCO₃) and non-calcareous marine plankton. If the growth of calcareous organisms is favored, it would cause an increase in the alkalinity (by the removal of Ca⁺⁺ ions), and hence, an increase in surface water pCO₂. Furthermore, various marine organisms use CO₂ and nutrient salts in different proportions, although the average C/N/P ratios are known to be about 106/16/1. If the growth of the organisms with higher C/P ratios is favored, more CO₂ would be removed from the surface water; hence the oceans would tend to take up an increasing amount of CO₂ from the atmosphere. However, no information is available to predict the impact of climatic changes on the global marine ecosystem.

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TERRESTRIAL ECOSYSTEMS AND CLIMATIC CHANGE

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INTRODUCTION

Climatic change caused by increasing concentrations of greenhouse gas may alter the earth's vegetation and the turnover of dead organic matter on land. The dependence of plant cell processes and microbial activity on light, temperature, moisture, and nutrients is the basis for this change, but with time, the consequences extend to variations in ecosystem characteristics, including species composition, biomass, and leaf profiles, and to continental distributions of vegetation zones or biomes (Larcher, 1975; Walter, 1979; Woodward, 1987).

Land systems partially shape climatic change as they respond to it. Terrestrial ecosystems influence climate by exchanging moisture and heat with the atmosphere, and they form a land boundary in terms of albedo and roughness. Furthermore, climatic perturbations can alter terrestrial carbon and nitrogen cycles, leading to further shifts in concentrations of atmospheric greenhouse gas.

The human habitat will vary tremendously if these tightly coupled transients in the global biogeochemical cycles and in climate unfold as most analyses indicate they will. In many regions, people will manage ecosystems to meet human objectives in the face of global change. But a large part of the earth's landscape will respond without intervention, and a clear

understanding of natural responses forms the basis for successful ecosystem management where humans do choose to control terrestrial ecosystems.

Terrestrial ecosystems and the global systems they interact with are intricate and tightly coupled. Observational studies are often impractical because of the time scales involved (decades to centuries), the vast extent and spatial heterogeneity of earth systems, and the small but critical perturbations to natural levels that are causing global change. Furthermore, many mechanistically important variables cannot be measured directly. Mathematical models are an important means of contending with these limitations of direct studies -- they are unsurpassed as tools for synthesizing and integrating diverse concepts and data.

Atmospheric general circulation models and models of the global carbon cycle assume that the major features of vegetation and soils are fixed -- maps of biome or vegetation and distributions of soil type prescribe these characteristics. But several studies suggest that terrestrial ecosystem responses to climatic change can violate static vegetation and soil assumptions (Emanuel et al., 1985a, b; Solomon, 1986; Pastor and Post, 1988; Prentice and Fung, 1990). While major shifts in the locations of biomes, such as grasslands and forests may eventually occur, more subtle variations in the biomass or leaf area of ecosystems may already be influencing global climate and element cycling.

A comprehensive, global model of terrestrial ecosystems is needed in climate and element cycling studies. This need has been apparent for some time, but only recently have the necessary concepts, data, and formalisms for such a model become available. In what follows, we describe some approaches to developing a global ecosystem model with particular attention to the difficulties associated with the range of scales involved and data limitations that require empirical substitutes for more formal relationships.

TERRESTRIAL ECOSYSTEM RESPONSES TO CLIMATIC CHANGE

Terrestrial ecosystems participate in climate and in the biogeochemical cycles at several scales. The metabolic processes that are responsible for the growth and maintenance of plants and soil microbes move carbon and water through rapid circuits in plants and soil and exchange heat and moisture with the atmosphere. However, plants establish a parallel circuit by incorporating some of the carbon they fix by photosynthesis into their tissue, delaying its return to the atmosphere until it is oxidized by decomposition or fire. This slower carbon loop through terrestrial ecosystems, which is matched by cycles of the nutrients required by plants and decomposers, affects the increasing trends in concentrations of atmospheric CO₂ and CH₄ and imposes a seasonal cycle on those trends. Furthermore, the structure of terrestrial ecosystems, built by this slower responding carbon machinery, supports the surface roughness and albedo boundary conditions of the climate system.

As plants respond to environmental change, competition for light and resources influences reproduction, growth, and mortality to different degrees

depending on species characteristics. As a result, the dynamics of species composition, biomass, leaf area, and similar community variables are more complex than the collected responses of individuals that do not interact. Climatic change affects plant productivity immediately, but the immediate responses are modified later by changes in composition and community structure. This adjustment process is similar to secondary succession after disturbances such as fire or harvest.

Natural areas are mosaics of patches that were disturbed at different times by fire, wind, flood, or disease. Large areas of vegetation reflect disturbance regimes that are part of their normal households -- such systems maintain steady state under ordinary local disturbances, but are perturbed by changes in disturbance frequency or intensity. Natural disturbance can quicken the response of vegetated areas to climatic change (Davis and Botkin, 1985), and presumably, the harvest of wood or crops can have the same effect. Climatic change can alter disturbance regimes; for example, warmer or drier conditions can increase the frequency of fire, causing forests to be replaced by prairie (Grimm, 1984).

The abundance of taxa surrounding an area determines the availability of propagules for local recruitment. At this level, climatic change can alter the regional distributions of taxa. Such changes are well documented for the past 20,000 years (Davis, 1981; Huntley and Birks, 1983; Webb, 1986; 1987). Trees migrated with remarkable speed in response to quaternary changes; however, the time needed for large-scale spread and changes in regional population may be 1,000 years or more (Davis et al., 1986; Bennett, 1986; Davis, 1987) and can presumably limit the rate of vegetation response to rapid, large climatic changes.

Vegetation and soil are tightly coupled systems. The holding capacity of soil-water and nutrient availability in the soil-water solution influence primary production. Although some aspects of soil formation are slow enough to be considered constant through plant community changes, nutrient turnover depends on litter characteristics so that variations in species composition can affect plant growth and further alter composition (Pastor and Post, 1986).

Thus, in the face of climatic change, terrestrial ecosystems have a certain inertia (Smith, 1965) comprised of several components: (1) patches of vegetation adjust to climatic change when the available flora is held constant (Type A Response of Webb, 1986), (2) nutrient cycles adjust to changing environmental conditions and vegetation characteristics, (3) the spatial mosaic of vegetation adjusts to climatically induced changes in the frequencies of fire and other natural hazards, and (4) succession further alters the structure and function of ecosystems as species establish and compete (Type B Response of Webb 1986).

HIERARCHICAL GLOBAL ECOSYSTEM MODELS

Intuitively, we might develop a global model of terrestrial ecosystem dynamics by combining descriptions of each of the physical, chemical, and biological processes involved. In such a scheme, longer term changes are derived by integrating rapid responses. However, we cannot simply integrate models that describe the rapid processes of CO_2 diffusion, photosynthesis, fluid transport, respiration, and transpiration in cells and leaves to estimate productivity of whole plants and ecosystems. The nature of the spatial averaging implied and the selection of parameters and processes to consider is difficult because of nonlinearities and heterogeneity. For example, photosynthesis models assume uniform biochemical concentrations over an entire plant, a condition rarely found outside of the laboratory. Furthermore, computational error accumulates when small deviations are multiplied by scaling factors or accumulated over long time periods.

To progress in the development of terrestrial ecological models, we choose processes to treat in different models based on the phenomenological scales involved. As is common in the more physically-based sciences, terms in fundamental equations are included or ignored depending on the temporal and spatial scales of interest. We expect the best global ecosystem analysis from a carefully organized suite of models, each describing processes that operate at different rates and with differing degrees of spatial variation. It is useful to organize models at three time scales and relate these in various ways for different kinds of analyses.

The most rapid responses, from minutes to days, involve gas and water vapor exchanges, the biochemical reactions of photosynthesis, and the microbial activity that decomposes dead organic matter. These processes, which are responsible for the rapid circuit of carbon through terrestrial ecosystems, quickly equilibrate to climatic variations and to changes in the concentration of atmospheric CO_2 .

At the slow extreme of response times, the life cycles of individual plants and interactions of plant communities due to shading and competing for nutrients and water determine the dynamics of species composition, biomass, leaf area, and similar community variables over decades and centuries. This long-term aspect of vegetation dynamics can lead to responses that resemble secondary succession.

On intermediate time scales, biotic activity, allocation, system structure, and leaf display change during a year in response to seasonal climate. These intermediate variations are associated with carbon allocation and other processes that are closely correlated with climatic variables.

Models of the rapidly responding components of terrestrial ecosystems assume that the major characteristics of the system remain constant and can be associated with large grid cells, as in atmospheric general circulation models. Inputs at this level include the full suite of climatic drivers and summary, static information about the state of vegetation and dead organic

matter. Outputs include the heat and moisture balance over the land surface; some models also calculate a net carbon flux into plants.

Seasonal, phenological changes in vegetation are described by models at a second level that are driven by climatic variables summed over time scales of days to weeks (e.g., degree-day sums, photosynthetically active radiation, total precipitation) as well as a productive potential or target structure of the system toward which the vegetation moves in response to climatic drivers. Outputs include seasonal changes in leaf area, leaf nutrient content, and soil-water storage. Changes in the chemistry of soil solution and the physical environment of the soil at this time scale are appropriate for driving models of trace gas fluxes.

A third class of models represents annual changes in total net primary production and ecosystem structure. Inputs include indices that summarize the effects of climatic conditions on annual plant growth and decomposition (e.g., estimated actual evapotranspiration, annual growing degree-days, and drought stress indices). Outputs include total net primary production and its allocation, including total leaf area. Traditionally, total annual nutrient release through decomposition is calculated with an annual time step.

Interactions between time scales occur in two ways: (1) short-term climatic data provide the fundamental forcing functions for all three levels, and (2) the longer-term models determine the surface conditions and vegetation structure used as input to the next, more rapidly responding representation. Thus, all three levels are driven by climatic data and constrained by calculated ecosystem structure. A level-three model converts annualized indices of climatic conditions and the current ecosystem state into total leaf area and structure for the next year. Given these total values, the second level calculates the phenology of leaf production and loss, and hence, the seasonal pattern of ecosystem structure. Using these seasonal patterns, the first level converts climatic data into energy and water balances over very short time steps.

Decomposition and nutrient cycling are simulated with a time step of approximately one month. They depend on litter quality, and thus require input from level-three vegetation models that describe slower dynamics. Rates in the decomposition model also are influenced by temperature and moisture, and thus require monthly aggregated climatic data. Growth and leaf nitrogen concentration are constrained by nutrient uptake.

A general structure can be used to represent respiration and carbon storage in both forested and grassland portions of the world. Soil organic matter in forests is dominated by the intermediate fractions, often accumulating in surface horizons. This intermediate material may be derived from lignin, which is not microbially reworked to form stable humus. Its accumulation is simply a function of its slow decomposition rate.

PERTINENT TERRESTRIAL MODELS AND ANALYSES

While comprehensive global ecosystem models are just becoming practical, several modeling approaches that comprise their basis have already given important insight into the potential responses of terrestrial ecosystems to global environmental change.

The natural, equilibrium distribution of vegetation is similar to climate. Plant geographers use this correlation to relate vegetation classes to climate (e.g., Holdridge, 1947; Box, 1981). Woodward (1987) developed an equilibrium model based on the physiological limits of plants, and Bartlein and co-workers (Bartlein et al., 1986) related regional abundances of particular taxa to climatic variables by empirical response surfaces.

The results of sensitivity tests of such equilibrium models suggest bounds on climatic change impacts. Emanuel et al. (1985a, b) compared world maps of Holdridge life zones derived from the climate recorded by 7000 meteorological stations and with simulated temperature increases for the concentration of atmospheric CO₂ twice the reference concentration. Lashof (1987) did similar exercises with a vegetation-climate relationship derived from Olson's world map of ecosystem complexes (Olson et al., 1983) and meteorological records. Although the transient responses would be complex (Shugart et al., 1986; Solomon, 1986), such tests indicate the sensitivity of the asymptotic distributions toward which transients eventually converge in the absence of disturbances not included in these equilibrium relationships.

The experiment described by Emanuel and co-workers shows greatest changes at high latitudes where simulated increases in temperature are largest and where narrow temperature intervals define Holdridge life zones. Changes are along boundaries and are more extensive than the uncertainty in determining these boundaries within the Holdridge scheme. Such speculations indicate some of the challenges agriculture and forestry may face and bound the broad vegetation changes that might feed back on the global element cycles and climate.

Past vegetation patterns reflected the climates of their time, so static vegetation models can be used to describe past vegetation in terms of reconstructed paleoclimates (Anderson et al., 1988). This approach assumes dynamic equilibrium (Webb, 1986), which is appropriate on time scales much longer than the vegetation's response time (Prentice, 1986a, b). The time scale of interest in studies of human-induced global changes is a few hundred years, comparable with the time needed for replacement of forest types; static models are not sufficient for prediction in this context.

For example, Emanuel and co-workers show conversion of much of today's boreal forest to temperate deciduous forest, but this result does not mean that a gradual, straightforward transition between the two forest types is likely. Higher summer temperatures may at first stimulate the growth of boreal conifers, but higher winter temperatures may be unfavorable for the natural regeneration of some of these taxa at their oceanic limits. Beyond

a certain point, increased summer temperature will reduce growth rates again; such a warm climate would probably be suitable for temperate trees, but their recruitment will take time, so production could fall before rising again toward the high level characteristic of temperate forests.

Various vegetation models describe transient responses from different viewpoints. Forest stand growth and succession models (Botkin et al., 1972; Ek and Monserud, 1981; Shugart, 1984) simulate changes based on species attributes and tree interaction assumptions that emphasize competition for light. These relatively new models borrow many ideas from forestry models (Munro, 1974). Nonwoody vegetation dynamics can probably be treated within similar descriptions, and although these models describe small landscape patches, they simulate community phenomena with considerable generality and use factors to describe environmental influences that can be evaluated along gradients, or to sample variability within larger landscape units.

The growth equations in these models are based on the following observations (Ek and Monserud, 1981):

1. Through time tree dimensions change according to sigmoid growth patterns; change in any one dimension (e.g., height) can be modeled as a sigmoid function (e.g., Richards 1959).
2. Other dimensions (e.g., stem diameter or crown width) are allometrically related to height.
3. Changes in stand density reflect thinning because of competition.
4. The heights of dominant trees of a given age and species are fairly insensitive to changes in stand density and are more closely related to potential site productivity than to other indices (Spurr and Barnes, 1980).

Stand models treat a plot of a definite size that is constrained by the assumption that light availability is independent of horizontal position (Shugart and West, 1979). Stand size can be constrained further to achieve realistic simulations of variations, such as gap-phase dynamics, the species replacement sequences initiated by the death of a large tree (Watt, 1947; White, 1979; Shugart, 1984; Smith and Urban, 1988). Shugart and West (1977) simulated gap dynamics in an eastern Tennessee temperate forest on a 833 m² plot. However, in boreal forests, tree crowns are narrow, sun angles are low, and the gap created by the death of a single tree is insignificant. Leemans and Prentice (1987) found that to simulate gap processes, models of these forests must be able to treat 1000 m² gaps for establishment of light-demanding species (Whitmore, 1982).

Conventionally, leaf area is directly related to stem diameter and considered to be concentrated at the top of the stem, so that the total leaf area of each tree shades all trees below. But some more recent models determine leaf area from sapwood area, allowing for sapwood turnover, and vertically distribute the leaf area of each tree (Leemans and Prentice,

1987). A vertical light profile is computed from the Lambert-Beer law, again assuming that shading is horizontally homogeneous.

Trees respond according to their species-specific shade tolerance. Environmental effects enter through factors that multiply growth increment to reflect deviations from best conditions. Usually below-ground competition is modeled only in terms of crowding, but some models now include an explicit nitrogen cycle, treating litter production, decomposition, immobilization, mineralization, uptake, and growth response (Pastor and Post, 1986).

Solomon (1986) showed that a stand simulation model treating 76 viable species can generate qualitatively correct forest types across eastern North America. Bonan et al. (1990) were able to simulate successional patterns in boreal forest sites from central Alaska to Newfoundland. These results show that if appropriate species are treated, a forest stand model can be applied at different locations by specifying values of its lowest level environmental variables: temperature, rainfall, light intensity, and soil characteristics. These same low-level variables are modified to analyze environmental change.

Solomon (1986) went on to illustrate responses to climatic changes derived from general circulation model solutions with increased atmospheric CO₂ concentration. He found that changes in composition brought about substantial changes in biomass that could be positive or negative, depending on location. This experiment considered only the effects of increased CO₂ through climatic change. Using reasonable assumptions about the direct effects of CO₂ increase on tree growth, Solomon and West (1986) showed that climatic effects are likely to predominate. Pastor and Post (1988) report model experiments similar to Solomon's, but with explicit treatment of nitrogen cycling. They find that effects of climatic change through nitrogen dynamics can be as important as direct effects on plant growth.

These results show what can be accomplished using stand simulation models to analyze the consequences of environmental change. By developing solutions that sample landscape units or environmental gradients, we can use a collection of these models to analyze the impacts of environmental change on world forests where human intervention can be ignored, or perhaps introduced into simulations as are natural disturbances. Other types of vegetation can be analyzed when nonwoody plants are incorporated in these or similar models.

Data requirements of plant growth for evaluating parameters of stand models are not demanding compared to other approaches; however, even these data are not available for some regions, particularly tropical moist forests of South America and Africa. Huston and Smith (1987) demonstrate that reasonable growth and succession dynamics can be simulated by a stand model that treats functional plant types rather than species -- these functional types and their parameter values can be derived from theoretical considerations (Smith and Huston, in press). Where necessary, vegetation can be modeled at this prototypic level until more data are available. As this

theory develops, functional plant types may prove to be more practical on the whole than are species.

Forest models based on the Botkin formulation are uniquely successful in describing stand dynamics on time scales ranging from decades to centuries. But a model that describes population changes for each species or even plant types, as opposed to treating individuals, may be satisfactory for most continental- to global-scale applications. The computational requirements of population-based models can be substantially less than those of models that track individuals.

One possibility is to describe the state of plant communities by the number of individuals of each species or plant type in each of a set of arbitrary width layers dividing space above the landscape unit supporting the vegetation. Again, the patch is sufficiently small so that light extinction can be assumed to be horizontally homogeneous.

Advancement through height layers is a stochastic process. The probability of transfer to higher layers is derived from a mean height increment for the population of each plant type, assumed to be uniformly distributed through each layer. The increment depends on current leaf area and size, and environmental conditions including available light, air temperature, and soil moisture. The structural emphasis on height and the assumption of homogeneity within height layers for each plant type or species are conducive to incorporating plant types other than trees for which a growth description based on individuals is unnatural.

Parton et al. (1987) describe a general model of organic matter turnover in temperate grasslands that can serve as a companion to forest stand models. Competition for light is less important in lower stature ecosystems; however, nutrient cycling, moisture demands, and other below-ground processes require careful description in grassland models. The Parton model considers several classes of dead organic matter that differ according to their resistance to further decomposition. Immobilization and mineralization of nitrogen determine its availability for plants. The model simulates broad-scale variations in grasslands with the specification of only four external variables: annual precipitation, temperature, soil texture, and plant lignin content. The nutrient cycling scheme used in the Parton model is consistent with that used by Pastor and Post (1986) in their forest stand model. Thus, a uniform description of below-ground processes can be worked out for use in grassland and forest models.

LARGE-SCALE APPLICATIONS

Terrestrial ecosystem models need to be organized for use at several levels of geographic detail and with the rapid and slow modules running in concert or separately. At the most detailed geographic level, underlying data are organized on a grid of land cells, and models are solved for each grid cell or with a sampling strategy. Both data and model solutions are mapped and managed by a geographic information system. The data requirements

and implementation logistics are very demanding at this level. In regional studies, data and model results are tabulated against biome or ecosystem extents; while in some applications, it is useful to average or lump data and model results to global scale.

Continental-scale simulations of vegetation dynamics can be generated by deriving very large sets of patch model solutions. The region of interest is subdivided; a uniform grid in spherical coordinates is convenient, but arbitrary polygons can be used. Resolutions of about 0.5° are reasonable (Olson et al., 1983; Emanuel et al., 1985a); such cells are approximately 50 km on a side at the equator.

A set of patch model solutions is generated for each of these smaller units with appropriately distributed random environmental variables. Disturbance frequency and intensity also are specified, which may depend on environmental characteristics such as temperature or soil moisture as well as the status of vegetation (Kercher and Axelrod, 1984). The solution set reflects specified variances in environmental variables, including disturbance frequency and intensity, but is complicated by the stochastic processes simulated by patch models.

For land units small enough to support reasonable resolution, the distributions of basic environmental variables, such as temperature, rainfall, soil type, and soil texture, cannot be derived from available observations. Rather, the distributions of these variables must be based on geomorphologic features; topography is most important.

There are models for calculating solar intensity at a specified location, elevation, slope, aspect, and time (Swift, 1976; Bonan, 1988; Kutzbach and Gallimore, 1988). Although basic principles are understood for other relations, for example temperature change as a function of elevation, these have not yet been expressed systematically in schemes suitable for global terrestrial ecosystem models. It is noteworthy that absolute relationships are not required, rather we need to derive the distributions of environmental variables from the underlying variability in topography and other structural characteristics.

Useful simulations of vegetation responses can be derived without considering connections between basic land units. However, at least three connections are clearly important: hydrologic transport, the dependence of disturbance frequencies on events in adjacent cells as in the spread of fire, and the dependence of seed propagation on the abundances of species in an area and in adjacent areas (Rudis and Ek, 1981). The latter two connections can be expressed without adding complex submodels; however, the large-scale hydrologic modeling needed to include the first is only beginning to be addressed (Eagleson, 1986).

Global terrestrial ecosystem models require the same geographic data management and analysis capabilities that are being demanded by virtually all fields of environmental science. However, some unique features are needed:

1. Values of environmental variables, such as temperature and rainfall, must be estimated for each land unit by interpolating observations that are usually of coarser spatial resolution. In continental-scale applications, interpolation on a sphere is required, and smoothing may be necessary (Renka, 1982).
2. Statistical analysis of spatial data on arbitrary landscape units of different resolutions is required to organize solutions by assigning joint distributions to environmental variables, and to summarize and interpret model solutions.
3. Collections of solutions must be analyzed to find geographic patterns in variables, such as biomass, and to map features that reveal the distributions of major vegetation types.

CONCLUSION

Those wishing to study interactions between climate and ecosystems must reconcile the different spatial and temporal scales of atmospheric and ecological processes. The design of atmospheric general circulation models reflects the fast horizontal transport and short memory of the atmosphere -- integrations are performed at intervals of less than an hour and at grid points 200 km or more apart. The slowest responding terrestrial ecosystem models are solved on yearly intervals, but on patches about 30 m across. These differences reflect the inertia of vegetation and the fine spatial scales at which ecosystem processes act compared to those determining weather.

There are general circulation models that allow a prescribed vegetation, in terms of height, structure, phenology, rooting depth to interact dynamically with the atmosphere (Dickinson, 1984; Sellers et al., 1986). Physical vegetation characteristics, such as albedo, leaf-area index, and stomatal aperture, vary diurnally and seasonally in response to atmospheric variables, and in turn, affect exchanges of energy, water, and momentum. The task of a global ecosystem model is to simulate the slower processes by which primary ecosystem characteristics -- taken as constant in climate models -- are transformed through time by changes in climate, including the processes of lateral spread.

We described the components of a hierarchical scheme for simulating global transient responses of natural terrestrial ecosystems to climatic change. The approach is based on several core descriptions associated with different temporal and spatial scales that in our experience simulate the processes that are responsible for terrestrial responses to environmental changes reasonably well. Substantial data are required to estimate model parameters for world-scale studies, but these needs are realistic.

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THE RESPONSE OF ECOSYSTEMS TO GLOBAL CHANGE: RESEARCH AGENDA

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THE BIOSPHERE - ATMOSPHERE INTERACTION

The biosphere plays a very important role in climatic change, principally by CO_2 exchange, water circulation, and energy budgets. The total contribution of the biosphere to the global carbon cycle is a source of controversy (Hobbie et al., 1984). Estimates differ in suggesting that the biosphere serves as either a sink for carbon as high as 4.7 Gt y^{-1} (Houghton et al., 1983), or acts as a source of carbon as high as 4.7 Gt y^{-1} (Houghton et al., 1983). Recent evidence from observed concentrations of CO_2 in the atmosphere and partial pressures in the surface waters of the ocean implies that the north-south CO_2 atmospheric gradients can be maintained only if sinks for CO_2 are greater in the northern hemisphere than in the southern hemisphere. These major CO_2 sinks in northern temperate latitudes, whose magnitude is as large as $2.0\text{-}3.4 \text{ Gt}$ of carbon per year, are provided by terrestrial ecosystems (Tans et al., 1990). However, forests in the north-eastern United States may be declining in productivity; the extent and cause of this decline are uncertain (Pitelka and Raynal, 1989; Stout, 1989).

Thus, to understand the contribution of the biota to the global carbon cycle and its subsequent influence on global climatic change, we must study how the productivity of terrestrial ecosystems, especially forests, is affected by high CO_2 atmospheres. Elevated global CO_2 levels and their associated increases in temperature may initially enhance plant photosynthesis, thereby absorbing additional CO_2 from the atmosphere. Yet, the increase in primary production may not continue indefinitely because of nutrient limitation caused by the predicted decrease in the decomposition rate of litter. A different scenario would suggest that because of large amounts of fixed carbon, trees could allocate more resources to root growth

and root exudates, which would increase nutrient availability to increased microbial activities. Whatever the scenario considered, the little doubt that temperate forests play a major role in the global carbon budget and are an important feedback to climate. Evidence to discern validity of these scenarios is essential if we hope to understand the role of temperate forests in modifying the global carbon budget and subsequent climatic changes. Therefore, we need to examine how plant physiological growth responses are altered by enriched CO₂ environments.

THE INFLUENCE OF ELEVATED CO₂ ATMOSPHERES ON PLANT PRODUCTIVITY

Predictions about the changes in the productivity of ecosystems are based on the observed increase in plant growth under high conditions of CO₂. Using data on single leaves, Gates (1985) calculated B' factors (the fractional increase in net primary productivity (NPP) with a fractional increase in the concentration of CO₂) for several species of deciduous forest trees and showed that they range from 0.33-0.53. Using high B' values, Gifford (1980) estimated high carbon storage in the biosphere (1.65 Gt y⁻¹ for B'=0.60). However, several authors (e.g., Botkin, 1977; Kramer, 1981; Hobbie, 1984; Lugo, 1983; Gates, 1985) have pointed out that because of the limits on plant growth already set by water and nutrient deficiency, and temperatures at the northern limits of distribution, primary productivity in natural ecosystems may not be substantially enhanced by the higher global CO₂ concentrations. Furthermore, even in systems that have the potential for an increase in production (e.g., arctic tundra), Oechel and Strain (1985) show that negative feedbacks may minimize any enhancement in growth induced by rising levels of CO₂. In contrast, Luxmoore (1980, 1981) suggests that increased photosynthesis in a high CO₂ environment would increase the amount of carbon allocated to roots, root exudation, mycorrhizal proliferation, and N-fixation. There also is evidence for increased nitrogenase activity at high levels of CO₂ (Norby, 1987). These factors, in turn, can lead to increased water and nutrient supplies to the plants and increased phytomass, even in somewhat infertile habitats (Strain and Bazzaz, 1983). The very limited evidence from field studies shows that there are either increases in productivity, especially during the first year or two of exposure to elevated CO₂ (e.g., Overdieck et al., 1984; Curtis et al., 1989), or no changes (Tissue and Oechel, 1987).

Thus, these responses to elevated CO₂ remain very poorly understood, and accurate predictions about the response of natural ecosystems to global increase in the levels of CO₂ still require much additional data on the mechanistic bases of the response of ecosystems (Strain, 1985; Dahlman, 1984). Thus, we still cannot answer whether rising levels of global CO₂ will have significant effects on overall plant productivity and the structure and function of natural ecosystems. Furthermore, we do not know whether any enhancement of primary productivity will be sustained over time or whether it will diminish, or even disappear, after a few years because of limitation by other environmental resources. Working scenarios can be proposed for the response of some ecosystems to climate change (Figs. 1, 2, and 3). These are based on the limited data available and on the general knowledge of these

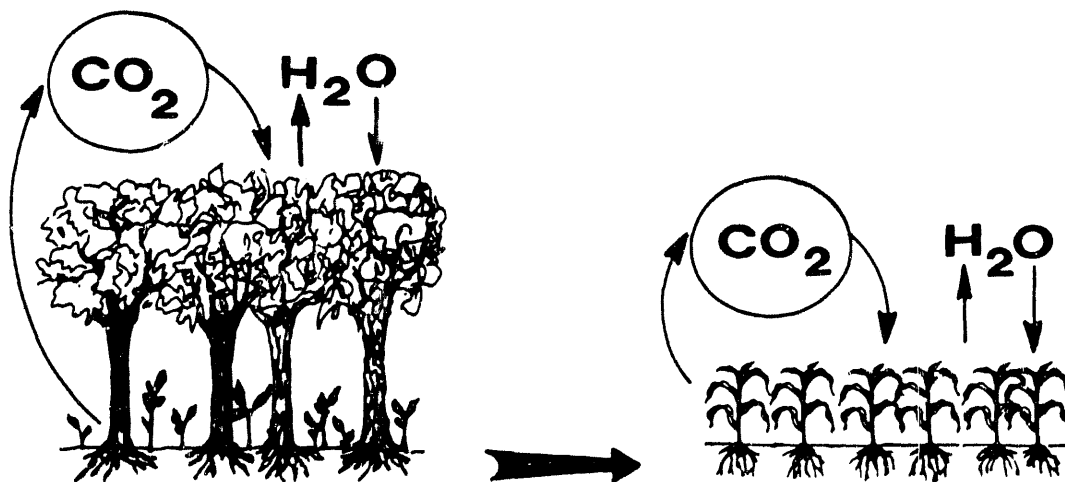
systems by several specialists. For example, a group of scientists meeting recently at Harvard proposed that the parameters shown in Table 1 are important for the impact of climate change. Therefore, a successful research program must consider the climate change, the direct effects of CO₂ on plants, and other components of ecosystems.

PHYSIOLOGICAL RESPONSES OF PLANTS TO ELEVATED CO₂ LEVELS

From an extensive body of literature on crop plants (Kimball, 1983a,b; Cure, 1985; Acock and Allen, 1985), annual plants (Bazzaz et al., 1984; Bazzaz and Garbutt, 1988; Bazzaz et al., 1989), and other species of plants (Strain and Cure, 1985; Strain, 1987), several generalizations about the response of plants to rising CO₂ can be confidently made (Bazzaz, 1990).

Within a certain range of concentrations of CO₂, the photosynthetic rate per unit leaf area for C₃ plants (i.e., all trees) will increase. Stomatal conductance and transpiration will decrease, and therefore, the efficiency of water use will increase. The enhancement of photosynthetic rates appears strongest early in the life of a plant or a leaf, and this enhancement may

Tropical Rainforest



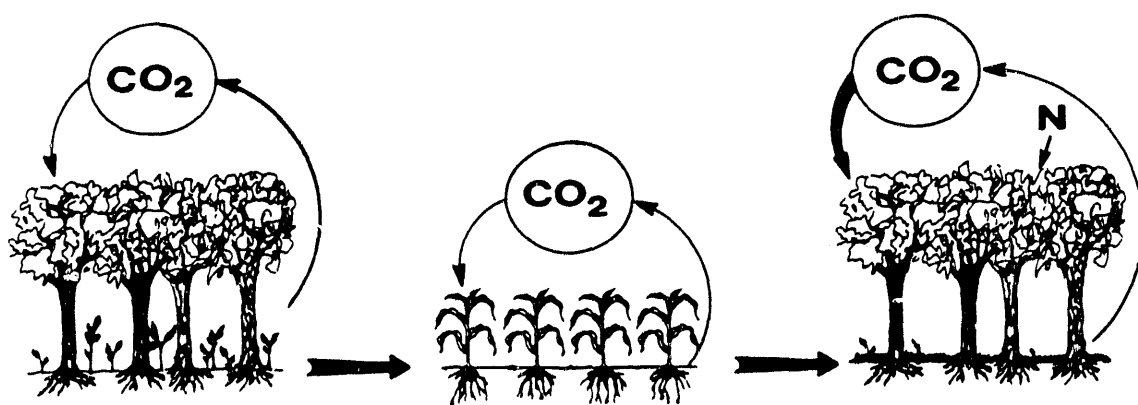
Changed hydrology

Fig. 1. A scenario of the response of a tropical rainforest to climate change.

CONCLUSIONS

Through our research program, we should investigate the effects of elevated CO_2 and other aspects of the climatic change on plants and other components and processes of ecosystems, particularly for ecosystems that seem especially sensitive to global change. We need to understand the controls on primary productivity and other aspects of the structure and function of ecosystems. Finally, we should combine our knowledge on the response of ecosystems with general circulation models, so that we will have an integrated atmosphere-biosphere approach to predict the impacts of global change.

Temperate Forests



- Changed tissue chemistry (C/N)
- Decreased litter decomposition
- Increased litter depth
- Lower nutrient availability?
- Changed consumption by herbivores
- Decreased seedling recruitment
- Changed ecosystem structure & diversity

Fig. 2. A scenario of the response of temperate forests to climate change.

Table 1. Potential for Feedback Regulation of Atmospheric CO₂

	NPP	Biomass	Soil Carbon	Response to Climate Change	Contribution to Atmosphere
Arctic	L	M	HH	HH	HH
Boreal	M	HH	H	HH	HH
Temperate	M	H	M	H	HH
Tropical					
Rainforest	H	HH	L	H	HH
Grassland	H	M	M	H	H?

L-Least, M-Moderate, H-High, HH-Highest

decrease or disappear altogether over time. Furthermore, plants can adjust their photosynthetic rate so that they do not differ when grown under various levels of CO₂.

Species differ substantially in the magnitude of their responses. Effects on respiration are less well documented, yet data show that there are increases and decreases with increasing levels of CO₂.

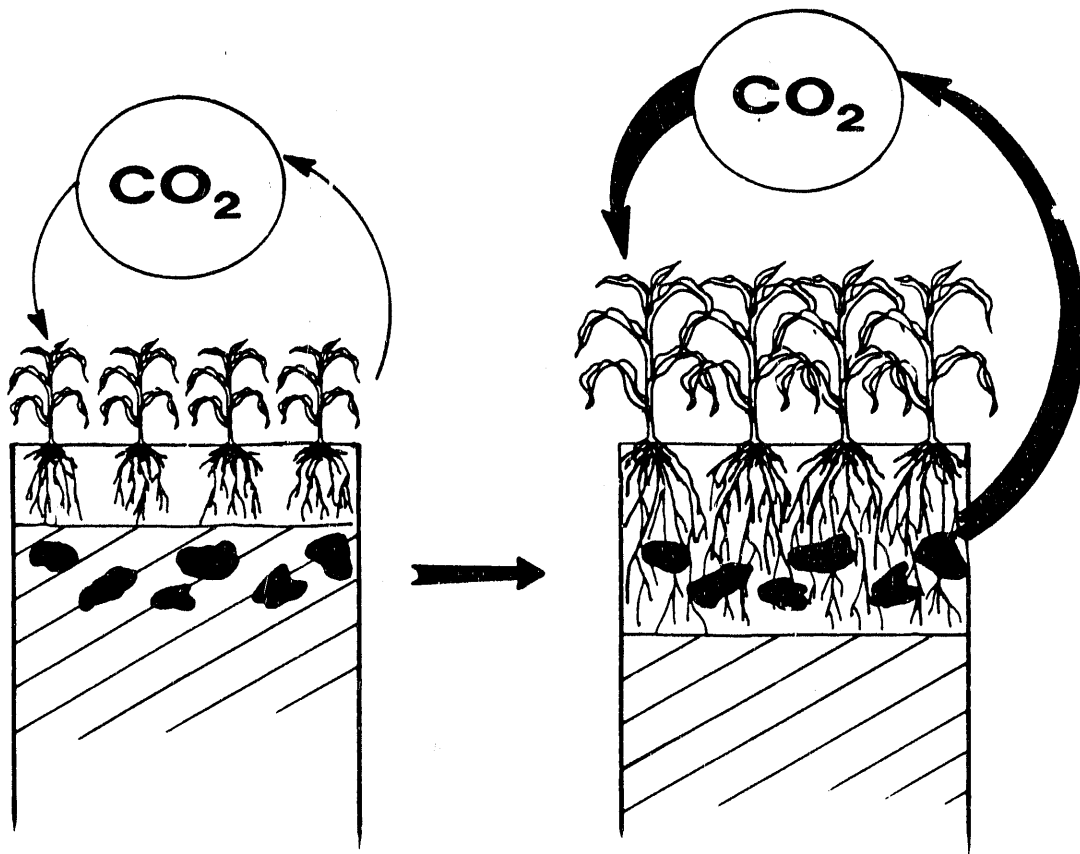
With increased levels of CO₂ plant morphology, resource allocation patterns, and reproductive biology can change. There is strong dependence of how plants respond to CO₂ when other environmental factors vary, especially light, soil moisture, nutrients, temperature, and some gaseous pollutants, such as SO₂ and nitrous oxides.

Because of differences in the responses among species of the same community, competitive interactions among them may change and lead to changes in community structure and ecosystem function. Community composition of ecosystems that are very nutrient-limited (e.g., serpentine grasslands) is less affected by elevated concentrations of CO₂ than that of nutrient-rich communities, even though some species in the nutrient-limited ecosystem show increased biomass accumulation.

The ratio of N to C, especially in leaves, declines when plants are grown in elevated CO₂. Herbivore consumption may then decline because of poor food quality, or increase because the herbivores must consume more foliage to grow and reproduce. Poor tissue quality in plants caused by growth in high CO₂ environments can, in turn, reduce the fitness of some herbivorous insects.

The critical areas of response that are important for understanding whether ecosystems are sources or sinks for atmospheric carbon are the potential for photosynthetic acclimation, dark respiration, growth and allocation, and interaction with other environmental factors.

Arctic Ecosystems



- Deeper thaw
- Lower water table
- Higher photosynthetic rates
- Higher nutrient availability
- Much higher decomposition rates
- Net CO_2 release

Fig. 3. A scenario of the response of arctic ecosystems to climate change.

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SEA ICE RESPONSE TO GLOBAL CLIMATIC CHANGE

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INTRODUCTION

The response of the polar regions to climatic change is significantly affected by the presence of sea ice. This sea ice cover is very dynamic and, hence, has a variety of thicknesses ranging from open water to pressure ridges tens of meters thick (reviewed by Hibler, 1989). In addition to undergoing deformation, the ice pack is typically transported from one region to another with melting and freezing occurring at different locations. Such transport tends to create net imbalances in salt fluxes into the ocean. On a large scale, an important factor is the amount of ice drifting out of the Arctic Basin through the Fram Strait. The variability of this export can be estimated from satellite data on ice drift. However, estimates of the thickness distribution of the ice are difficult to obtain. Moreover, the spatial and temporal variability of the transport is substantial. Also, there is the issue that the mechanical characteristics of the ice cover, which dictate many of its outflow characteristics, may vary in response to climatic change, and hence, may cause feedback effects that could affect the response of the system.

As a consequence, it is clear that while statistical models and observations of present-day ice extent are important for validating physically based models of sea ice growth drift and decay, reliance on such observations to ascertain the sensitivity of the high latitudes to climatic variations may leave out important feedbacks which could affect the response. Instead, it appears important to develop physically based models to successfully explain observed features of sea ice growth drift and decay and then to use some version of these models in numerically based studies of climate.

In this regard, it is important to develop models that include ice drift and dynamics in climatic studies and to begin to examine the response of such models to simulated, rather than observed, atmospheric forcing. These comments need to be viewed in light of the fact that present climatic studies usually only include thermodynamic sea ice models, which do not even come close to including the major sea ice processes relevant to climatic change. As a consequence, inclusion of any level of verified data on ice dynamics would be an improvement provided the forcing wind fields of the atmospheric circulation models have an acceptable level of correctness. Some simple robust sea ice dynamic models developed by the author and coworkers are discussed below.

Overall, there are three broad areas where determining the physical mechanisms is important for developing a physically based understanding of the response of the high latitudes to climate change: Sea ice dynamics and thermodynamics, the thickness distribution of sea ice and its evolution, and the coupling of sea ice with the ocean. Aspects of these features that are relevant to climatic change are discussed below, including some perspectives on recent research.

SEA ICE DYNAMICS

General Characteristics of Sea Ice Drift

From observations the overall characteristic of ice drift is that on short time scales, ice drift tends to follow the wind, with the drift approximately following the geostrophic wind with about one-fiftieth of its magnitude. This general feature of ice drift has been known for many years, beginning with Nansen's expeditions to the Arctic at the end of the 19th century. Since wind variations are larger than currents on a short time scale, then fluctuations in ice drift will be dominated more by wind than by currents, except for shallow regions. However, on a long time scale, the steadier currents can play a significant role.

These characteristics are illustrated in Fig. 1, which shows short-term and long-term drift rates using a linear ice-drift model. As seen, including current effects has only a minor effect on short-term variation. However, in examining cumulative drift over several years, significant differences occur if currents and ocean tilt are neglected. Basically, although smaller, current effects are steady. On the other hand, wind effects, while large, tend to fluctuate out over a long time, leaving a smaller constant value.

Ice-Ice Interaction

The need for a non-linear ice rheology is an important facet of ice dynamics. Considering sea ice to be a two-dimensional isotropic continuum, a general constitutive law applicable to non-elastic deformation is

$$\sigma_{ij} = 2\eta\dot{\epsilon}_{ij} + [(\zeta - \eta)(\Sigma\epsilon_{kk}) - P]\delta_{ij} \quad (1)$$

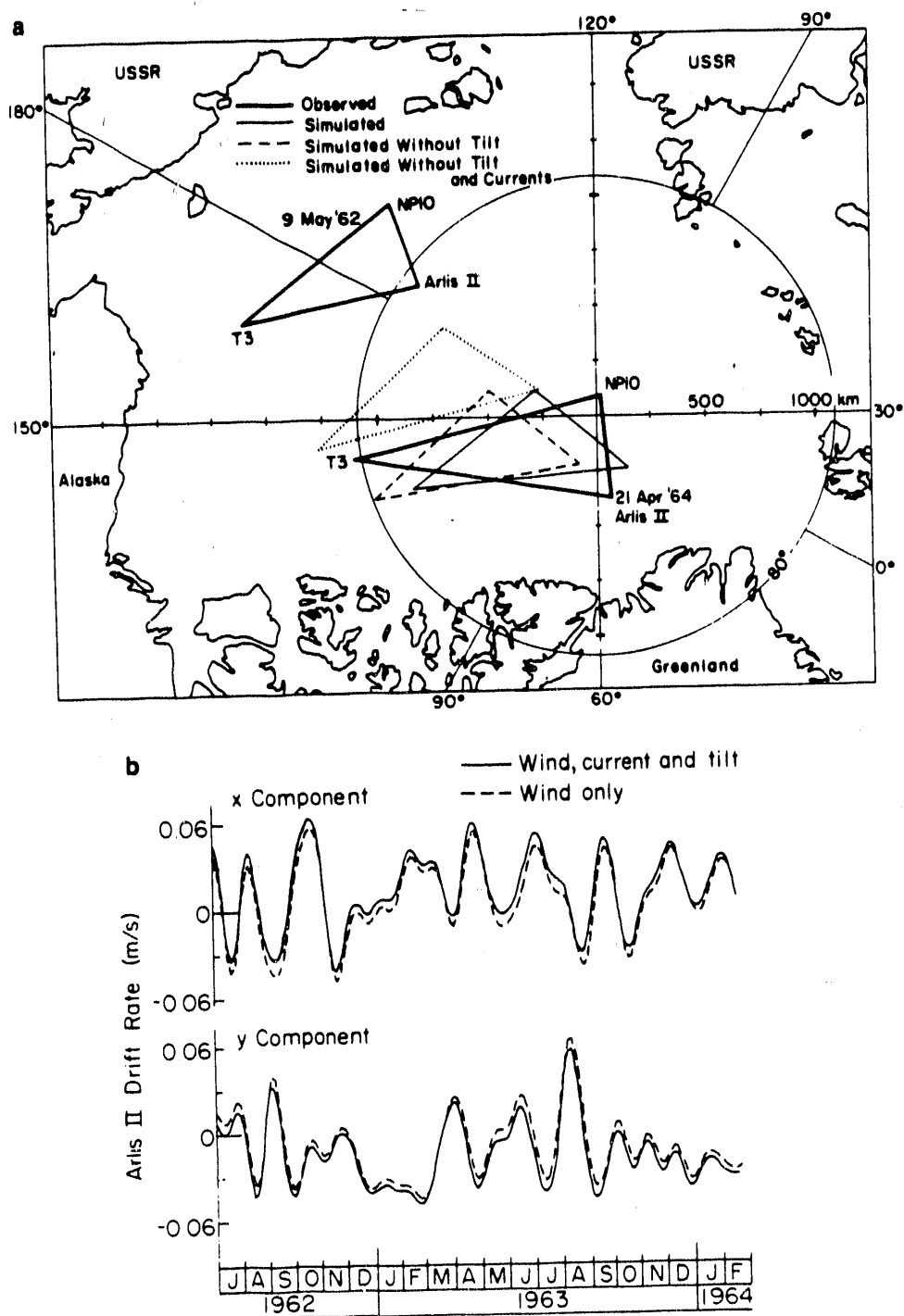


Fig. 1. (a) Predicted and observed net drift of three drift stations, with and without current and ocean tilt effects. (b) Effect of currents and ocean tilt on the 25-day smoothed predicted drift rate of the Arlis II ice island. (From Hibler and Tucker, 1979).

where σ_{ij} is the two-dimensional stress tensor, and $\dot{\epsilon}_{ij}$ is the strain rate tensor. In constructing a law beginning with this general framework, one would like to take into account the following observational characteristics and intuitively reasonable assumptions: discontinuous slippage near shore; rigid ice motion and lack of ice motion under considerable wind forcing; small or zero tensile strength, both for uniaxial divergence as well as real two-dimensional dilation; and high compressive strengths. These last two points basically say that sea ice should offer very little resistance to being pulled apart, while, at the same time, offering considerable resistance to being pushed together. In addition to these characteristics, viewed from above, sea ice clearly takes on the appearance of a closely packed, granular medium (Fig. 2). Based on granular medium studies in other materials, this correspondence suggests some type of constitutive law where the stresses are largely affected by contact stresses between floes rather than by stresses transmitted by eddy diffusion.

A rheology that satisfies these characteristics is some type of plastic rheology. This fact was first noted by Coon (1974), who based much of this proposition on the similarities between sea ice and a granular medium. In the central Arctic, it is also possible to base plasticity on the physics of pressure ridging in that mechanistic pressure ridge models show that the

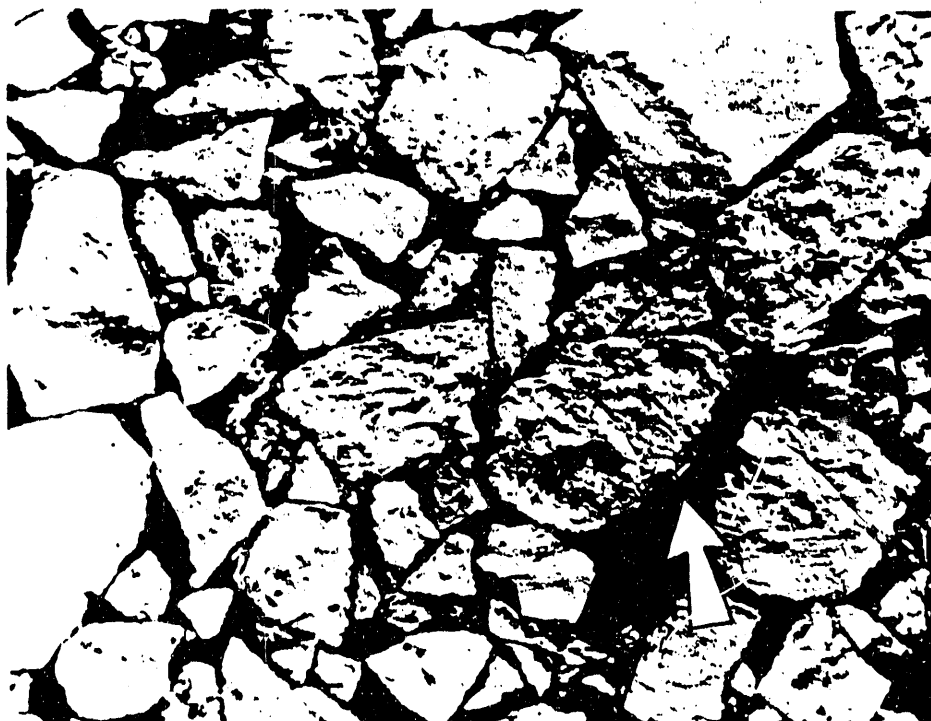


Fig. 2. Typical aerial view of pack ice in the central Arctic in summer. The ship shown in about 50 m in length.

work needed to build ridges is relatively independent of how fast the ridge is formed (Parmeter and Coon, 1972; Hopkins and Hibler, 1989). By dimensional consideration this leads to a relatively rate-independent stress. It is easier to visualize the essential differences between different rheologies in one dimension (Fig. 3). This figure shows that the linear viscous rheology differs essentially from a plastic rheology in that the stress is rate-independent. Figure 3 also sketches the viscous-plastic rheology approach introduced by Hibler (1979) to model non-linear flow. Here rigid flow is approximated by a state of very slow creep. Also, collision-induced rheology (Shen et al., 1987), while non-linear, has the stress more rapidly increasing with convergence.

In two dimensions, the basic idea in a plastic rheology is to introduce some type of yield curve in stress space, where the stresses lie on the yield curve when flow is occurring (Fig. 4). The ratio, but not the magnitude, of the principal components of the strain rate tensor dictates where the stress lies on the ice yield curve. Based on analogies to elastic plastic systems, the associated flow rule has often been used, which specifies that the ratio of principal components is the same as a normal to the yield

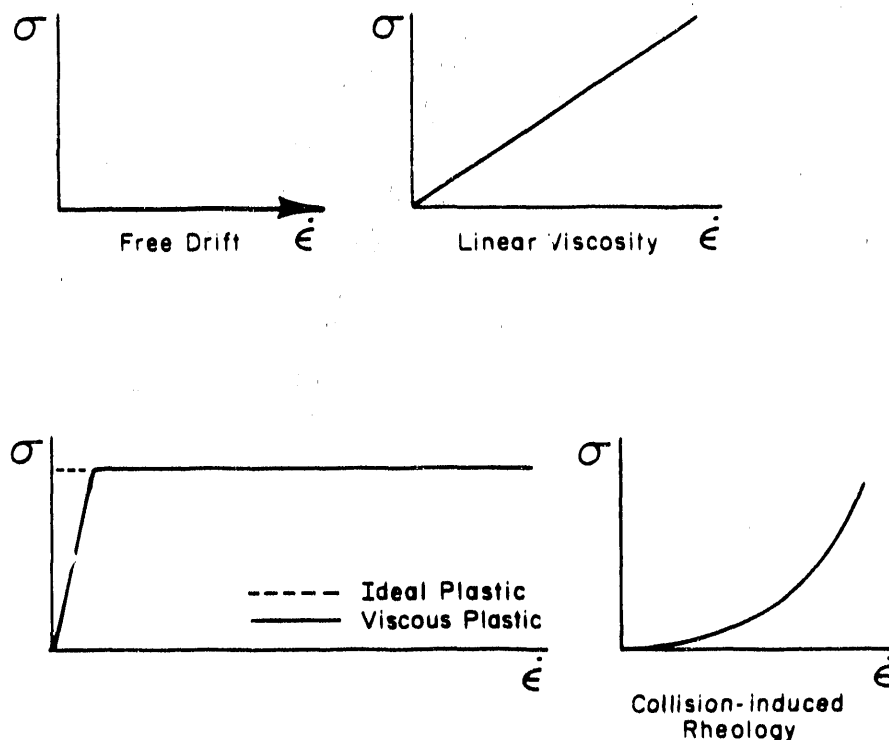


Fig. 3. Comparison of sea ice rheologies in one dimension (a positive $\dot{\epsilon}$ here represents convergence).

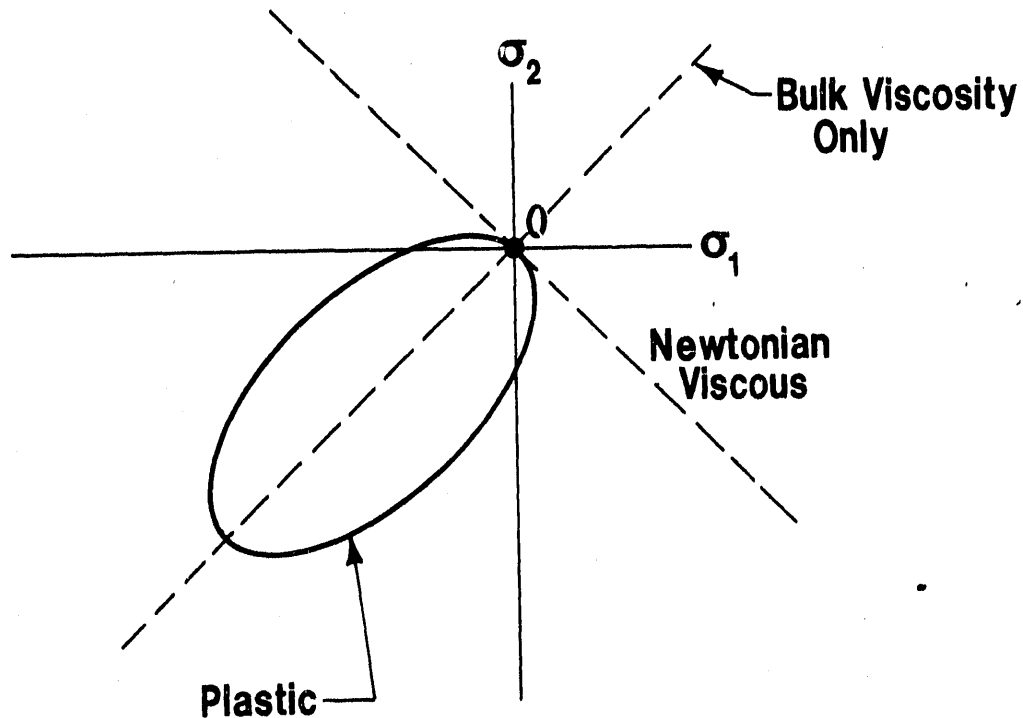


Fig. 4. Allowable stress states for a Newtonian viscous rheology with only a shear viscosity, a linear viscous rheology with only a bulk viscosity, and an ideal plastic rheology with an elliptical yield curve. The stress states are plotted as a function of the principal components of the two-dimensional stress tensor. In the linear viscous cases, the stress states can lie anywhere on the dotted lines, whereas in the plastic case the stress state is restricted to be inside or on the yield curve.

surface. Since elastic behavior is doubtful for sea ice, such a flow rule is not absolutely necessary. However, as shown below, application of this rule yields an intuitively acceptable rheology.

Probably the best smooth yield curve for sea ice is an ellipse, as proposed by Hibler (1979) (see Fig. 4). This curve allows only a very small tensile stress and has a larger compressive stress than shear stress. This smaller shear to compressive stress ratio is consistent with analysis of ice deformation data (Hibler, 1974) and is also reasonable intuitively. Applying the normal flow rate to this curve yields (Hibler, 1977) values of η , ζ , and P in equation 1 of

$$P = P^*/2$$

$$\zeta = P^*/2\Delta$$

$$\eta = \zeta/e^2$$

where,

$$\Delta = [(\epsilon^2_{xx} + \epsilon^2_{yy})(1 + 1/e^2) + (4/e^2)\epsilon^2_{xy} + 2\epsilon_x \epsilon_{yy}(1 - 1/e^2)]^{1/2}$$

and $e = 2$. Note that a given stress can be produced by various combinations of strain rates, so that strain rate is not a unique function of stress.

One of the strongest justifications for using a plastic law is the behavior of ice slippage near the shore. An example of the application of remote sensing imagery to address this problem is the study of Hibler et al. (1974), an illustration from which is shown in Fig. 5. (This study is a good example of the use of satellite imagery to better understand ice dynamics.) This figure shows observations of ice motion over one day off the coast of Alaska. The basic nature of the deformation is effectively an almost discontinuous slippage near the coast with a more rigid motion further out. This type of behavior is practically impossible to simulate with a linear viscous law. In particular, simulating this behavior requires that the viscosities in Equation 1 vary as a function of deformation rate. Analysis of seasonal dynamic-thermodynamic simulations shows this type of discontinuous slip behavior is nicely modeled by a plastic model.

Plasticity and Free Drift: The Necessity for High Ice Interaction in Seasonal Simulations

Observationally we know that on short time scales ice drift tends to follow the wind. While not obvious, analysis of numerical and analytic ice models shows that this feature is characteristic of a plastic rheology. However, thickness build-up is totally different with a plastic rheology compared to free drift (where the ice is merely taken to follow the wind), with free drift yielding totally unacceptable results on ice build-up for climatic studies.

To demonstrate the necessity for using an ice interaction in seasonal simulations, a series of comparisons of coupled dynamic-thermodynamic simulations were carried out for the Arctic Basin, using the two-level dynamic-thermodynamic model of Hibler (1979). The ice strength constant P^* and grid are the same as those used by Hibler and Bryan (1987) in a coupled ice-ocean model. As can be seen from the velocity fields in Fig. 6, the average annual ice velocity looks similar in both the simulations. However, the ice build-up in the full drift case is totally unacceptable. In particular, even after only one year, the free drift simulation results in unrealistic build-ups of over 19 meters off the Canadian Archipelago, together with excessive ice convergence in the center of the Beaufort Gyre. On the other

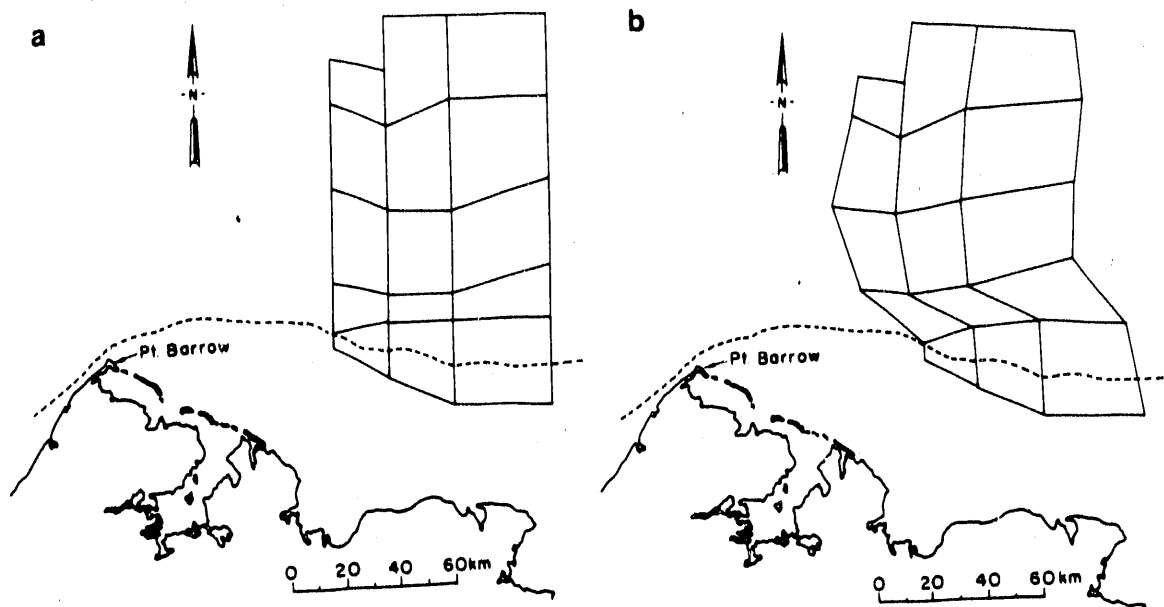
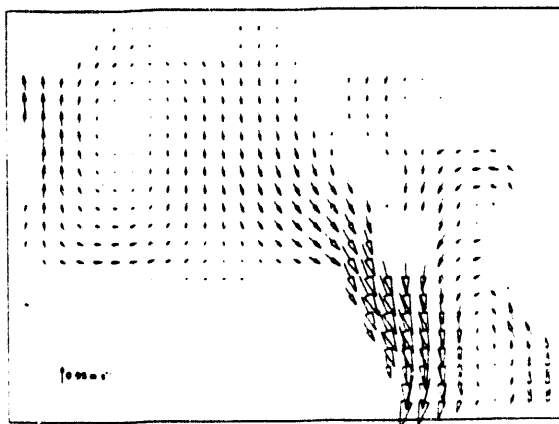


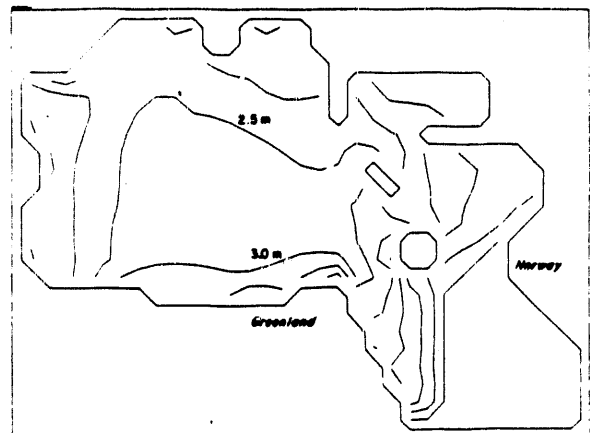
Fig. 5. Observed ice motion over a one-day interval during shear activity in the near-shore region: (a) 19 March 1973; (b) 20 March 1973. The observations were made from Landsat imagery and the dashed line marks the boundary between the shore-fast ice and the pack ice. (From Hibler et al., 1974b).

hand, the strong resistance to convergence in the plastic model yields a relatively realistic build-up. Basically, the deformation field of the ice velocity is a much more sensitive test of a model's performance. This fact, in turn, infers that to optimally aid in model development, satellite remote sensing should either be capable of resolving ice deformation or providing some type of measure of thickness or ridging. These results also illustrate the danger of just using the statistics of predicted ice drift as a test of model performance (see e.g. Thorndike and Colony, 1982). If a model yields good drift statistics but unrealistic ice build-up, for example, then its utility and physical basis are clearly limited.

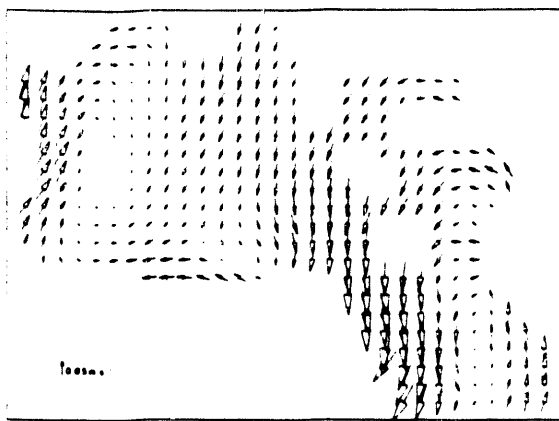
As can be inferred from Fig. 6, the plastic model offers a consistent way to construct a weakly damped ice velocity field without excessive convergence. The computer time needed for this is not excessive. The year-long integration at one-day time steps for Fig. 9, for example, required 6 minutes of Cyber 205 computer time, of which about 70% was used to solve the momentum equation. Considering that the equilibrium time scale of the ice model is about 3 to 5 years, then the total computer time needed for equilibrium



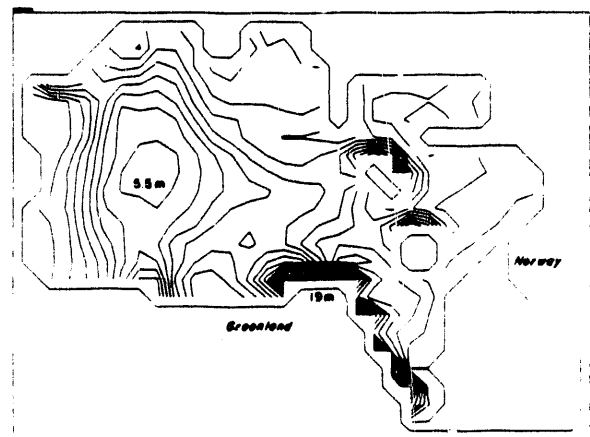
a. Full model.



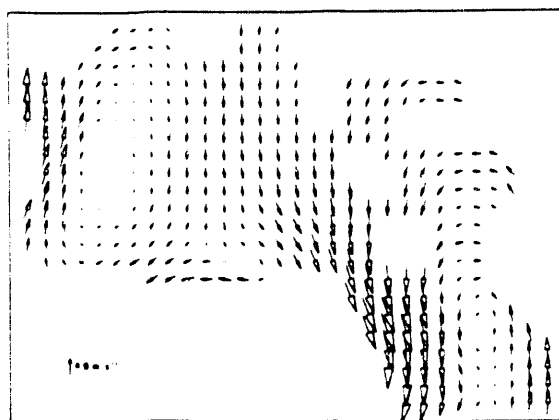
d. Full model, first year.



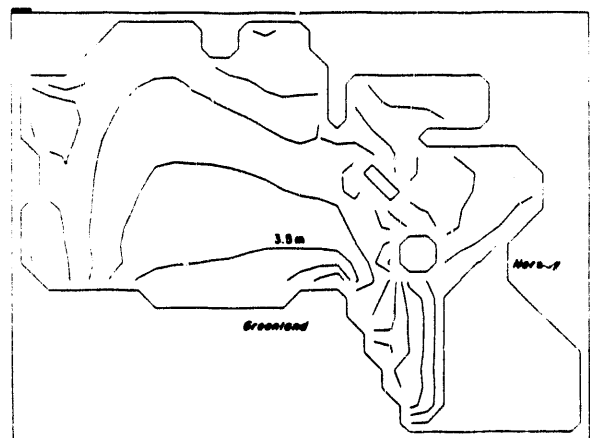
b. Free drift.



e. Free drift, first year.



c. Annual average wind velocity.



f. Full model, fifth year.

Fig. 6. Simulated annual average ice velocities (a,b) and ice thickness at the end of December (d,e,f) for a plastic model versus a free-drift model. Also shown is the average annual wind field (c).

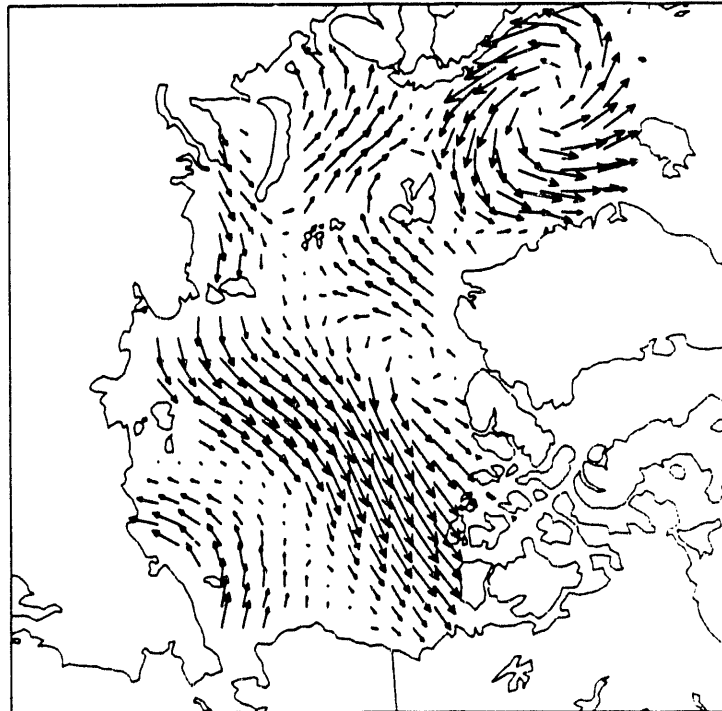
simulation of both polar covers is substantially less than the equilibrium computer time of an atmospheric or oceanic General Circulation Model (GCM).

Simpler Parameterizations of Sea Ice Dynamics: Feedback Effects

The two-level dynamic-thermodynamic model of Hibler (1979; 1980), which employs an elliptical yield curve, has been used for a variety of large- and small-scale studies in both the northern and southern hemispheres (Hibler and Walsh, 1982; Hibler and Ackley, 1983; Walsh et al., 1984) for examining both the seasonal cycle and the interannual variability of ice thickness and extent. Recently it was also coupled to a one-dimensional mixed layer model and used to investigate the seasonal cycle of sea ice and ice thickness in the Weddell Sea (Lemke et al., 1989). Some of the dominant physical results of these investigations are the identification of the role of sea ice dynamics and ice transport in creating spatial imbalances of salt fluxes (Hibler, 1980; Hibler and Ackley, 1983) which can have a major effect on the ocean circulation, and the effect of leads and ice advection on the rapid decay of the ice cover. Specifically, Hibler and Ackley (1983) found that with full ice dynamics, a precipitous ice-edge decay could be simulated for the Weddell Sea without the need to use large amounts of heat flux to the ice, which had been argued in earlier studies on the basis of thermodynamic considerations alone. This result was justified on a more fundamental level by Lemke et al., (1989) who found that with an interactive mixed layer model, vertical heat fluxes were relatively small, except near the ice edge. These authors also demonstrated that the exclusion of dynamics rendered a coupled ice-mixed layer model much less sensitive to perturbations in the oceanic characteristics due to the ice interaction. This negative feedback effect also was identified by Hibler (1984) in a series of warming sensitivity studies of a dynamic-thermodynamic sea ice model in the Weddell Sea.

While probably usable in its present form (Hibler, 1979) for coupled atmosphere ice-ocean models, there has been some need to develop a simpler sea ice model including full dynamics that has certain practical computational and physical advantages. (For example, we might like to make use of mean monthly rather than daily wind fields.) Work on a "cavitating fluid" modeling is underway at Dartmouth College with initial results on its application to an ice edge and to the Arctic Basin ice cover (Flato and Hibler, 1989; 1990). The basic idea in this model is to allow ice to freely diverge but to have substantial resistance to convergence with negligible shear strength. A momentum conserving and computationally efficient scheme for model this behavior was developed by Flato and Hibler (in preparation) and the characteristics of this model are being investigated. This model offers several particular advantages for large-scale modeling of climate. For one, because of the lack of shear strength, it is less affected by artificial islands imposed at the North Pole and by irregular coast lines often used in global ocean circulation models to handle the consequences of meridian convergence near the North Pole. Also, it is usable at longer time steps, or with mean monthly forcing used to drive some ocean models. Typical drift results obtained from this type of model are shown in Figs. 7 and 8.

Velocity Field (m/a) - Free Drift



Velocity Field (m/a) - Incompressible Cav. Fluid

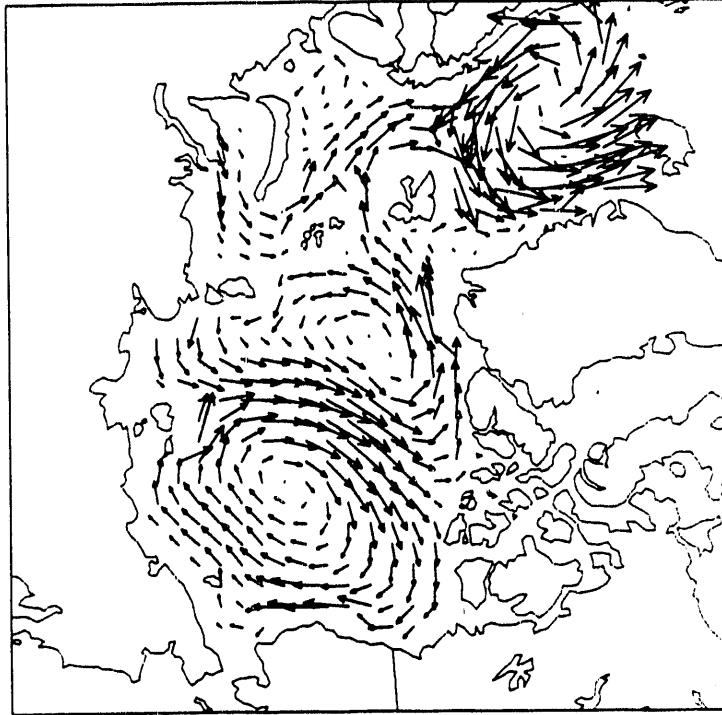


Fig. 7. Ice drift results obtained by the application of a cavitating fluid model to ice drift while without ice interaction is shown in the "free drift" figure. A velocity vector one grid cell long is approximately 0.12 ms^{-1} . Fig. 8 also shows some results from a spherical coordinate version of the model with an island at the pole as often used in global ocean models. Due to the lack of shear strength with the cavitating fluid, reasonable results can be obtained on ice drift, outflow (through the Fram Strait) and ice thickness, even with the island present.

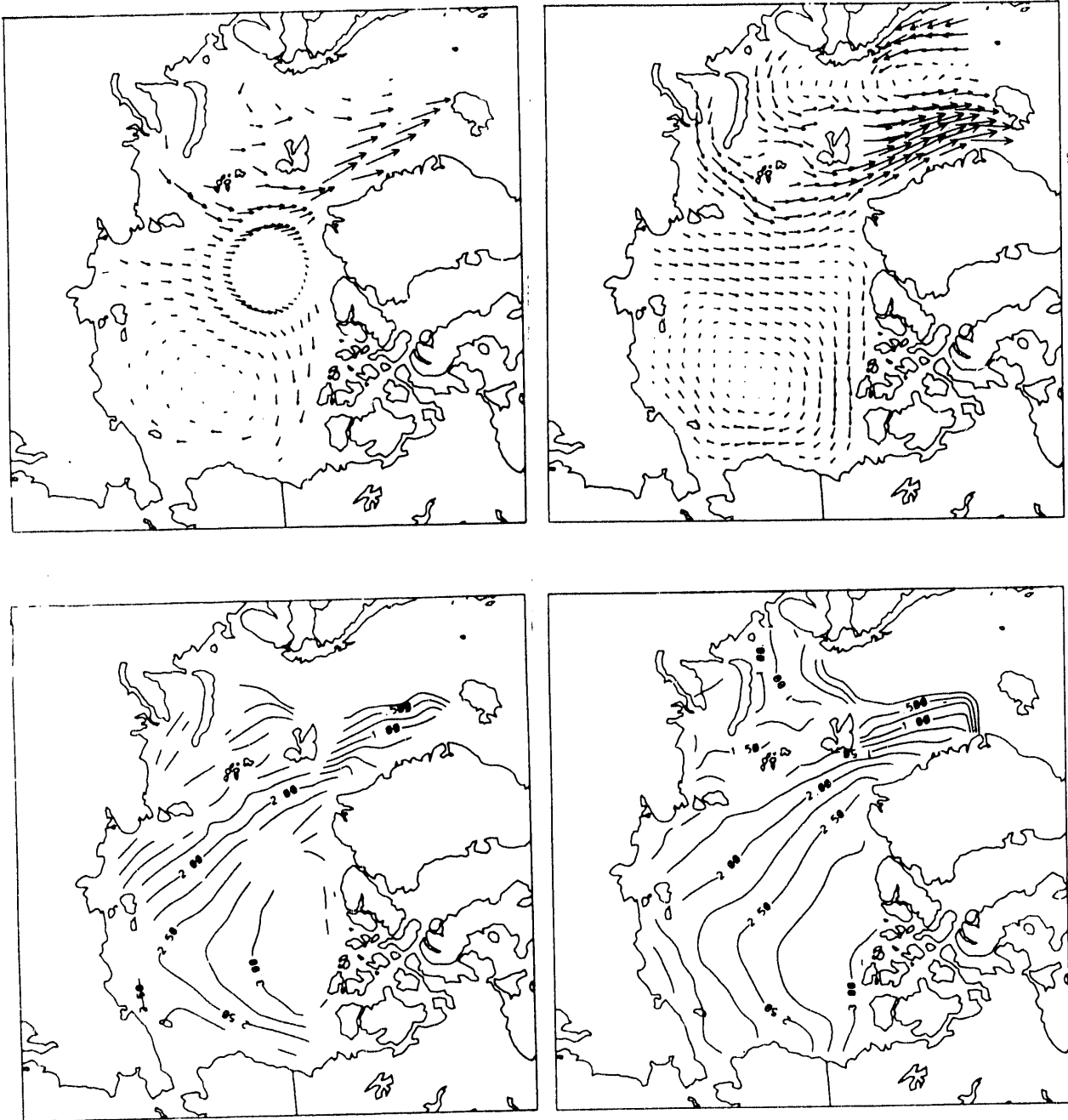


Fig. 8. Average January ice velocity fields and ice thicknesses for a cavitating fluid model in spherical co-ordinates with an Island at the pole and a rectangular co-ordinate system.

Some preliminary analyses of this model are shown in Fig. 9 and Table 1 where comparisons between buoy drift and simulated buoy velocities are given. Due to the lack of shear strength in the cavitating fluid model case, the drift is somewhat extensive. Also, when the correlation coefficients between the various simulated and observed time series were calculated, they were almost identical. However, there are substantial differences between the different model results with, for example, the predicted net drift being substantially larger where shear strength is neglected.

On the negative side, however, it should be emphasized that this type of model does not yield as accurate an ice drift field as the full viscous plastic sea ice model (Hibler, 1979), and, in many cases, may predict velocities that are too large. Also, the details of the thickness build up

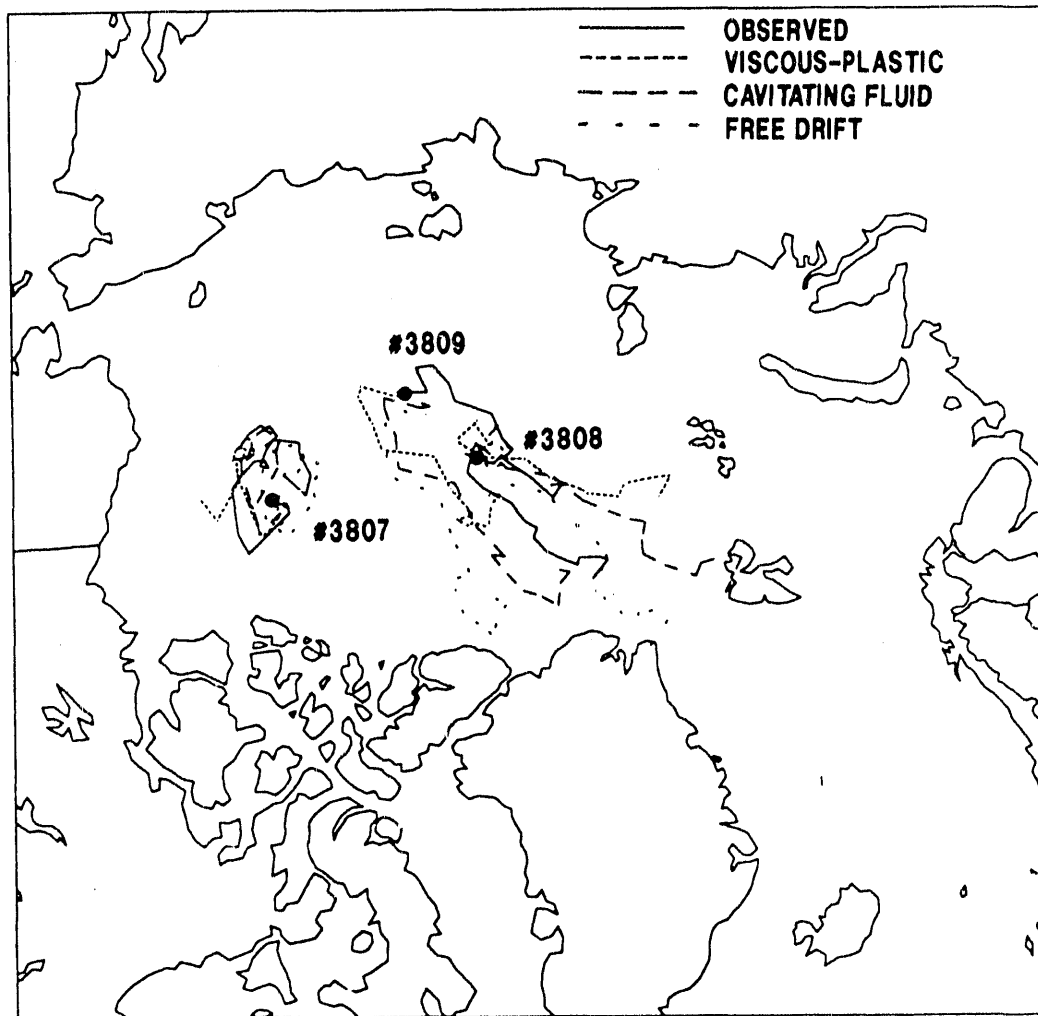


Fig. 9. Simulated and observed buoy drift velocities for three buoys for the period June 81 to June 83.

Table 1. Buoy Drift Statistics Compared to Three Different Models

Monthly Average Drift (Model vs. Observed)	
Model	Correlation Coefficient
Viscous-Plastic	0.841
Cavitating Fluid	0.829
Free Drift	0.821

Average Drift Results for 3 Buoys			
Model	Buoy Drift (km)	Model Drift (km)	Error Radius (km)
Viscous-Plastic	98.1	107.3	66.1
Cavitating Fluid	98.1	134.4	76.1
Free Drift	98.1	131.7	73.6

and the degree to which ice thickness characteristics are successfully simulated are not totally clear. Some of these features are being investigated and analyzed (Flato and Hibler, in preparation).

It should also be noted that this type of model is similar to a special case of the Hibler model (1979) in that there is no shear strength. On my suggestion (private communication, 1984), such a "bulk viscous" model (with a slightly different numerical closure) was used by Semtner (1987) in an ice-ocean circulation model driven by mean monthly forcing. However, analysis of the bulk viscous numerical method (Ip et al., in preparation) shows it to be much less computationally efficient than the cavitating fluid procedure. The procedure developed by Flato and Hibler (1990, and in preparation) could also be used in a situation where one starts with an approximation to free drift and modifies the velocity field in a momentum-conserving manner to derive an ice velocity field that prevents excessive convergence. Under these circumstances, the initial velocity field might be derived from empirical wind ice drift correlations that include some form of ice interaction.

An additional interesting feedback induced by ice pressure (see Flato and Hibler, 1989) is its tendency to cause an ice edge to expand, even with a symmetric wind forcing. Basically, the ice interaction induces internal pressure which ultimately pushes against the coast. By Newton's third law, the coast pushes back on the ice which can lead to a net outward ice flux.

SEA ICE THERMODYNAMICS AND THE ICE THICKNESS DISTRIBUTION

An item of some importance to climate response is the need for including a full multi-level ice thickness distribution in ice models used for climate studies (Thorndike et al., 1975). This is important because of the much more realistic treatment of the ice thickness distribution that can be affected by such a model, and hence, better treatment of the sea to air heat exchanges and the concomitant salt and freshwater fluxes to the ocean. Investigation of such a model also leads to a more fundamental inquiry into the nature of ice rheology due to the explicit parameterization of the ridging process in such models. A particular active area of investigation is the amount of energy dissipated in the ridging process. This was recently examined via a discrete element granular media model to provide a more fundamental physical basis for the rheology and ice strength used in coupled dynamic-thermodynamic sea ice models (Hopkins and Hibler, 1989). This investigation also has some relevance to climate simulations vis-a-vis the ice drift patterns since with a full rheology, higher strengths can actually prevent ice from moving out of the Basin during winter. In particular, Hibler (1980) found that using a full variable thickness model, and assuming that the total energy dissipated in ridging was 10 times the potential energy loss, there was a stoppage of the ice outflow out of the Basin during April and May.

The Role of Rheology in Simulating Ice Build-Up

By coupling an ice thickness distribution and ice strength to a plastic rheology it is possible to carry out an equilibrium seasonal simulation in which ice thickness changes reach an equilibrium seasonal change in three to five years. This simulation can be done using a two-level ice thickness distribution, a procedure used in most dynamic thermodynamic model studies today (Hibler, 1979; Hibler and Walsh, 1982). However, it is also possible to treat the ice thickness distribution more completely using a multi-level formulation. This procedure was used by Hibler (1980), in conjunction with a full heat budget-based thermodynamic model, for a seasonal simulation of the Arctic ice cover following Thorndike et al. (1975). In a multi-level thickness distribution, one uses the governing equation for ice thickness distribution:

$$\frac{\partial g}{\partial t} + \nabla \cdot (\mu g) + \frac{\partial (fg)}{\partial h} = \Psi$$

where gdh is the fraction of area covered by ice with thickness between h , and $h+dh$, f is the vertical growth rate of ice of thickness h , and Ψ is a redistribution function (depending on h and g) that describes the creation of open water and transfer of ice from one thickness to another by rafting and ridging (see Hibler, 1986 for a discussion of different formulations specifying that work done in building ridges is equal to deformational work (first noted in two dimensions by Rothrock, 1975) via the equation:

$$C \int_0^{\infty} h^2 \psi(h) dh = \sum_{i,j} \sigma_{ij} \epsilon_{ij}$$

where C is related to the gravitational and buoyancy forces on ice.

Fig. 10 shows the results of ice thickness build-up from the Hibler (1980) study, together with observed build-up. As can be seen, the ice build-up in a dynamic-thermodynamic model is in much better agreement with observed build-up than that in thermodynamic-only models. (An additional obvious shortcoming of thermodynamic-only models is that no outflow from the basin, as we know occurs, is simulated.) As can be seen from the average annual velocity field in Fig. 10, this thickness build-up is due to the ice dynamics, which tend to pile up ice off the Canadian archipelago while removing it from other regions. As this ice gets thicker it gets stronger and tends to prevent further convergence.

The Effect of a Full Thickness Distribution on Average Ice Thicknesses

Of more relevance to the issue of climate is the mean ice thickness with and without a multi-level sea ice model. Fig. 11 shows the time series of Arctic Basin averaged thickness with a two-level model and with a full variable-thickness sea ice model. As seen for the full multi-level model, the increased growth of the full ice thickness distribution makes up for the transport of ice out of the Basin and yields a mean ice thickness very close to what would be obtained by thermodynamics alone. On the other hand, in the two-level case the transport of ice out of the Basin is not fully balanced by the somewhat increased growth of the ice cover due to a fraction of open water present. It may be possible to tune the two-level model in such a way that increased growth occurs by prolonging the lifetime of the fraction of open water or modifying the growth of the thick ice.

As mentioned before, it should also be emphasized that such transport means that the net ice growth, and hence, net salt flux, at a given location will differ substantially from zero. How big an effect this can be is illustrated in Fig. 12, which shows net growth contours from seasonal equilibrium dynamic-thermodynamic simulation of the Arctic ice cover (Hibler and Ackley, 1982). This net imbalance in local salt fluxes can be especially important in ice-ocean coupling and modeling, as net melt will tend to stabilize the ocean, whereas net freezing will lead to increased overturning and convection.

OCEAN-ICE COUPLING

This is such a major topic that we cannot do it justice here; in fact, the topic is complex and involves a combination of issues related to both ice and ocean dynamics. What we can point out at this juncture is that the salt and momentum fluxes between a dynamic melting/freezing ice cover and the ocean drive much of the complexity of the system, and, as a result,

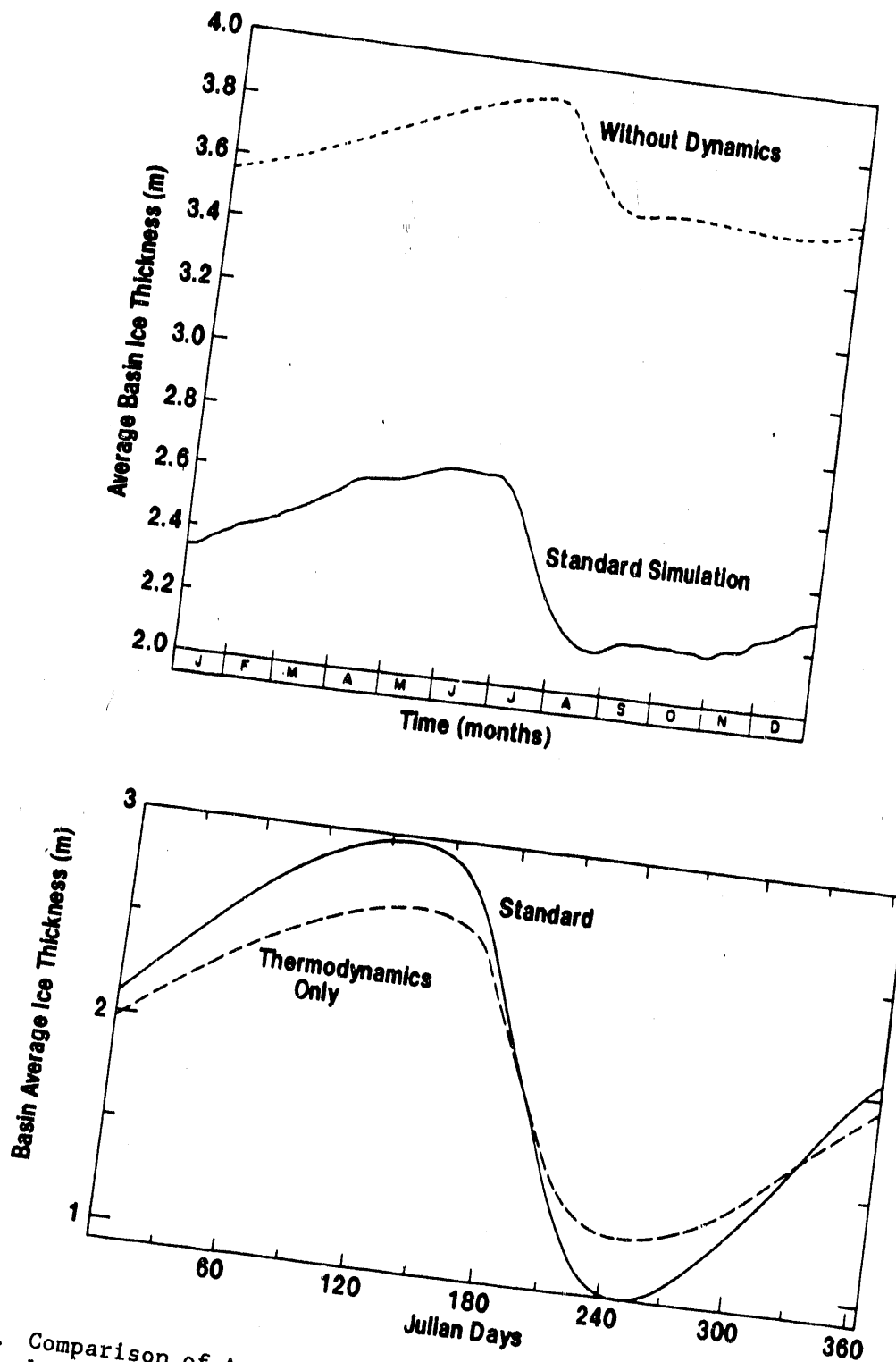
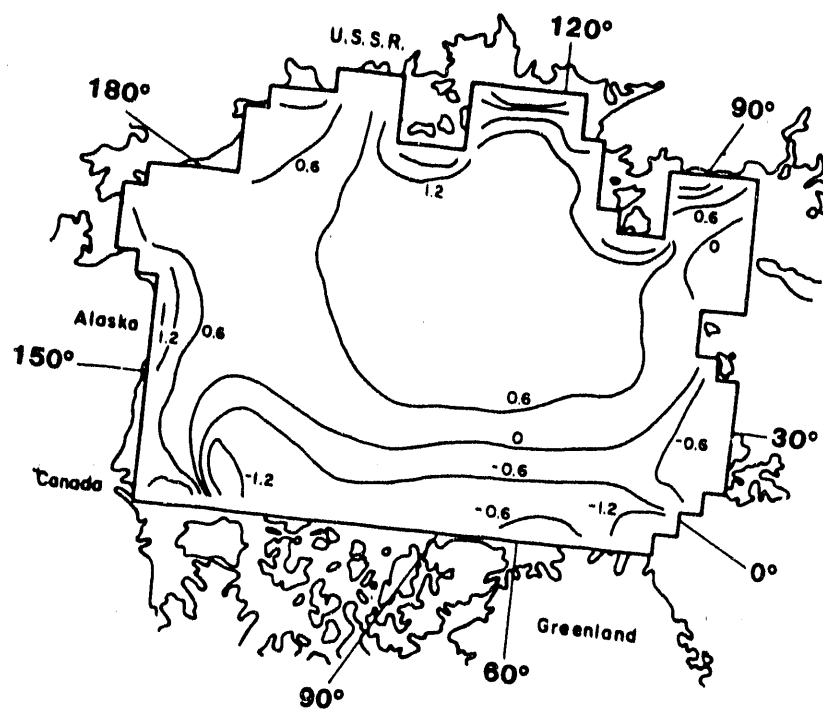
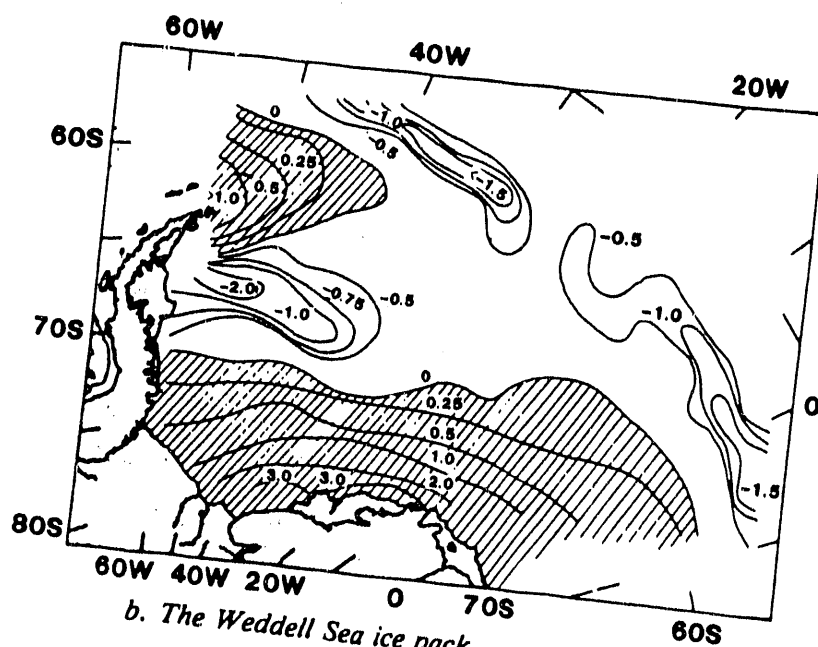


Fig. 11. Comparison of Arctic Basin averaged ice thicknesses for a) a two-level model with spatially invariant growth rates (Hibler, 1979) and b) a multi-level sea ice model with full thermodynamics (Hibler, 1980).



a. The Arctic Basin ice cover



b. The Weddell Sea ice pack

Fig. 12. Net annual ice growth from a dynamic-thermodynamic simulation.

the degree to which modification of surface salt fluxes and surface wind stresses by the ice cover affect the ocean circulation. A framework for addressing this issue was developed by Hibler and Bryan (1987), consisting of a combined ice-ocean circulation model that coupled a baroclinic 14-level ocean model with a two-layer dynamic-thermodynamic sea ice model. The results of this model (in a "robust diagnostic" form whereby the model was weakly relaxed to observed temperature and salinity fields) were analyzed by Hibler and Bryan (1987). Their results show some of the major effects of ocean circulation on the ice cover of the northern hemisphere. Particularly notable is the improvement in the location of the ice margin location shown in Fig. 13 with the full ocean model included. This figure also shows that sensitivity studies without lateral ocean circulation but with convective overturning produce a much less realistic ice edge. This, in turn, implies that the full lateral ocean circulation is needed to get realistic results in the Arctic, Greenland, and Norwegian Seas. The issue of the role of the ice cover on ocean circulation is less clear and requires inclusion of a full ice rheology to understand all the effects.

The interannual variability of this type of model has also been investigated by Ranelli and Hibler (in preparation). Of importance to climatic studies is the effect of the ocean circulation on the ice thickness and areal extent through oceanic heat flux variations. To show some of the overall characteristics of the ice edge results, Fig. 14 presents the average ice thickness contours in February, together with the simulated and observed ice edges for three years. Of particular note in the results on the ice edge is a more extensive open water region in the Barents Sea region than was obtained by Hibler and Bryan (1987). This ice edge result is consistent with the observed ice edge and is made possible by a very substantial oceanic heat flux in the Northern portion of the Greenland Sea which leads to a more realistic ice edge than could otherwise be simulated. However, the ice-ocean model does not successfully simulate the "Odden" like extrusion feature south of Iceland in 1982 and 1983. There is also a slightly excessive simulated ice edge in the southern part of the Greenland Sea, possibly due to sluggish transport through the Denmark Strait.

The role (or lack of role) of sea ice on bottom water formation in the North Atlantic is an important issue with respect to climatic response. This subject is likely tied in closely to the hydrological cycle in the sense that increased precipitation may create a much greater ice extent which could, in turn, lead to salinity anomalies such as were recently observed in the North Atlantic. It has also been hypothesized (Myzak et al., 1989) that North Atlantic salinity anomalies may have a feedback to high-latitude precipitation, resulting in interdecadal cycles roughly consistent with 20th century data sets. While such arguments are speculative, they provide a basis for physically based model studies to see if such mechanisms hold up.

It should be emphasized that in understanding the ice edge advance and retreat, boundary layer processes play an important role. As shown by Hibler and Bryan (1987) and Hibler and Ranelli (in preparation), the overall variations of the ice edge can be substantially improved by a large scale ice-ocean model. There are many discrepancies generally with a problem of

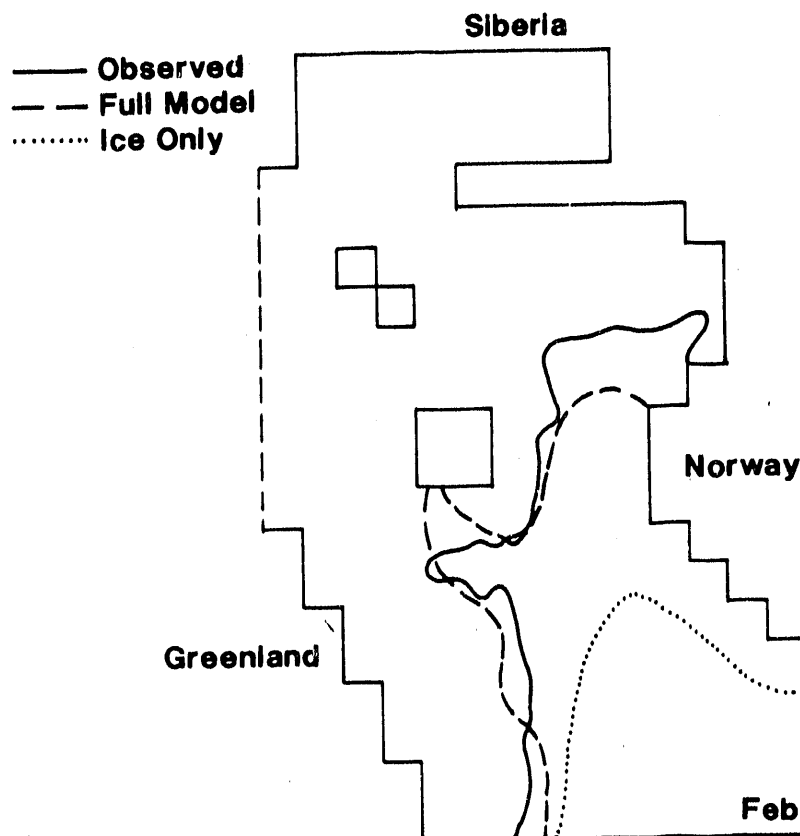


Fig. 13a. Simulation results for the February 50% concentration limit for a coupled ice-ocean model and for an ice-only model.

too great an ice extent. The role of penetrative convection on retarding the ice advance and retreat was recently demonstrated by Houssais and Hibler (in preparation) in a two-dimensional coupled ice ocean model including boundary layer dynamics. The major results are shown in Fig. 15, where a time series of the ice margin with and without ice advection and with and without boundary layer physics is depicted. Essentially, when lateral ice advection is included, the inclusion of penetrative convection greatly reduces the ice edge advance that occurs as the ice moves outward and stratifies the ocean. In the absence of ice motion (thermodynamic sea ice model) the effects of convection have a lesser effect on the locations of the ice edge.

A COMMENT ON THE SENSITIVITY OF ANTARCTIC SEA ICE MODELS TO ATMOSPHERIC WARMING

To examine the effect of ice dynamics on the response of sea ice to atmospheric warming, a hierarchy of Antarctic simulations was carried out and analyzed by Hibler (1984). In these simulations, atmospheric warming

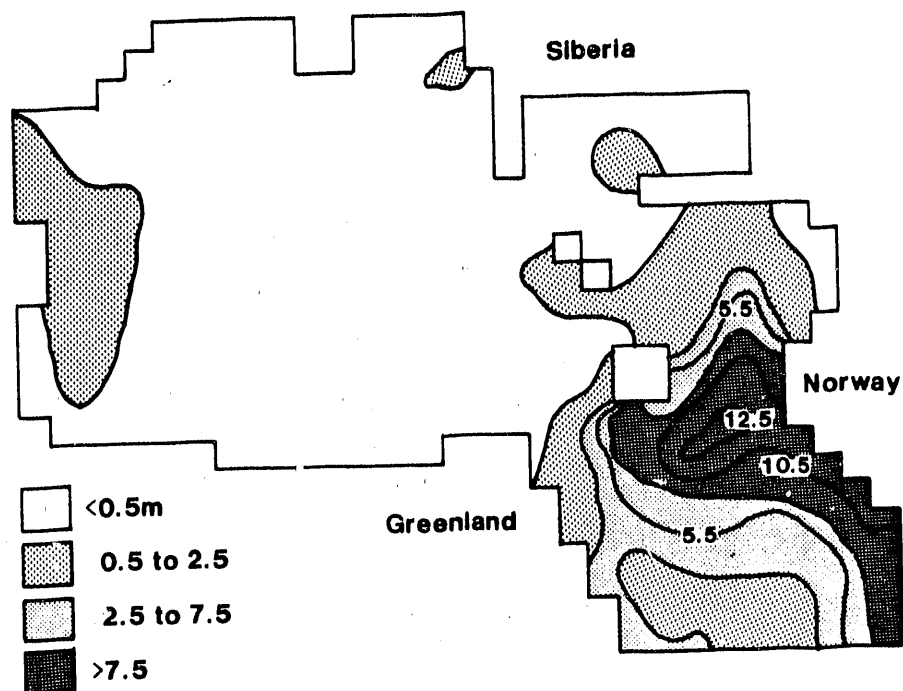
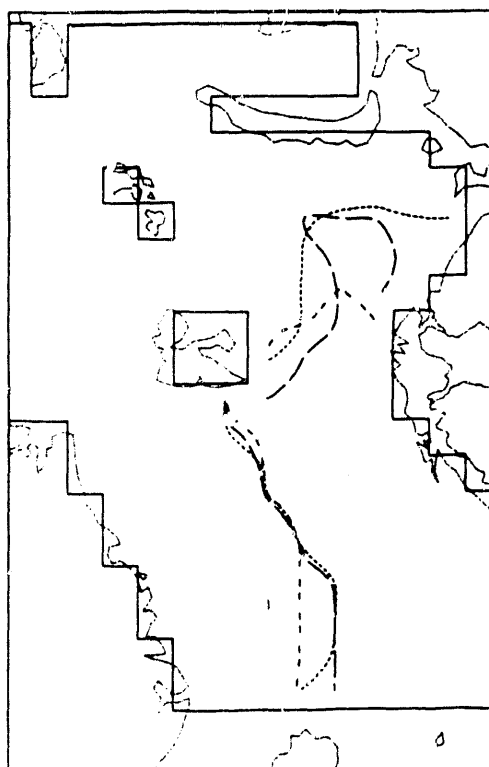


Fig. 13b. Average annual heat gained by the upper layer of the ocean from the deeper ocean and by lateral heat transport. The contours are in terms of the melting capacity of the heat (in meters of ice per year). A melt rate of 1 m of ice per year is equivalent to 9.57 W m^{-2} .

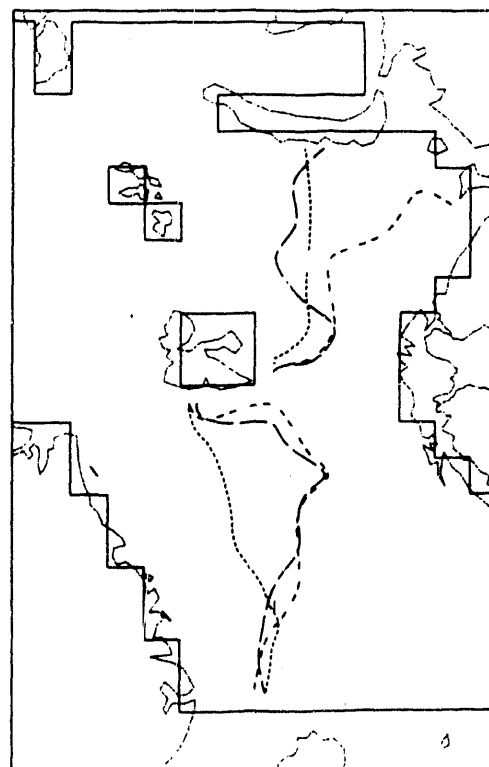
was crudely modeled by allowing the atmospheric temperature to rise by 4 degrees Centigrade. Some of the results are outlined here. Overall, the findings show how various combinations of different dynamic and thermodynamic effects can interact. Three basic models were examined: a fully coupled dynamic-thermodynamic model, a static thermodynamics-only model, and a static model with *in situ* leads. In the "leads-only" model, the leads are specified to be uniform spatially with values equal to the spatially averaged daily results from the full model (Table 2).

While all models tended to yield a reduced ice extent under warming, which increased the overall sea to air heat transfer, it was notable that a prominent feature of the results was the decrease in net ice growth through the ice cover for models including dynamics, as opposed to an increase in ice growth for the thermodynamic-only models. Thus, for the ice-covered region the thermodynamics-only model has an increased heat loss under atmospheric warming, while both the full- and fixed-leads models have a decreased heat loss.

Modeled Ice Edge(Diag) 1981-83



Observed Ice Edge--1981-3



- FEBRUARY

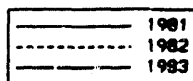


Fig. 14. Simulated and observed ice edges for February based on a 20% concentration limit.

Consequently, not only do the dynamics cause a quantitative change in the air-sea heat exchanges, but they cause a qualitative modification, in that the sign of the heat loss change is different. This decreased heat loss in the two models that include dynamical effects is due to the dominance of ice growth in leads. Growth rates over open water tend to decrease as the atmosphere gets warmer and, hence, there is less heat loss.

CONCLUDING REMARKS: UNDERSTANDING THE RESPONSE OF SEA ICE TO CLIMATE CHANGE

Because of almost constant motion and deformation the dynamics and thermodynamics of sea ice are intrinsically coupled. In addition, salt and freshwater fluxes into the ocean due to the freezing and melting of ice, together with ice transport yielding unbalanced fluxes, tie the ice and ocean circulation together. As a consequence, understanding the response of the high latitudes to climatic change requires considering the coupled ice ocean system (including ice interaction) in the polar regions. The results and

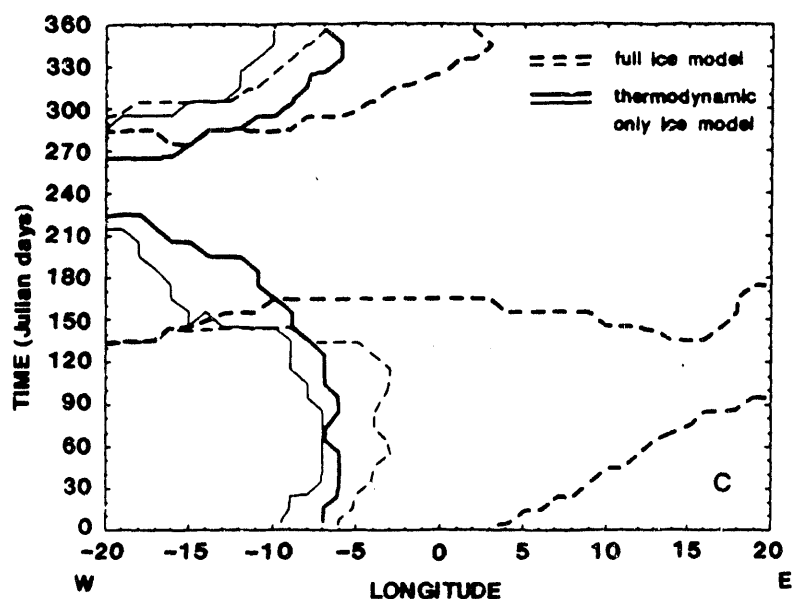


Fig. 15. Ice edge versus time. The light solid and light dashed lines are for the mixed layer model.

theory reviewed there have indicated the complexity of different thermodynamic and dynamical effects and the role they can play in air-sea interaction. This complexity makes it difficult to guess the correct ad hoc treatment of sea ice to use in climate models. Instead, the results emphasize the importance of including a more realistic treatment of sea ice vis-a-vis a fully coupled ice-interaction-based dynamic-thermodynamic sea ice model. These models, at least, contain the main first-order aspects of the sea ice system, whereas simple thermodynamics-only models clearly do not.

Table 2a. Net Heat Loss to Atmosphere, Day 60 to Day 240 Over the Whole Grid, in Units of Ice Thickness (m) Change^a

	Coupled model	Fixed leads	Thermo-only
Control temperatures	722	718	588
Warm temperatures	833	846	752
% change	122	118	128

^aAn average thickness change of 1000 m of ice over this time period is equivalent to an average sea-to-air heat flux of 114 W m^{-2} over the same 180-day period.

Table 2b. Net Ice Growth, Day 60 to Day 280, in Units of Ice Thickness Change^a

	Coupled model	Fixed leads	Thermo-only
Control temperatures	133	169	27
Warm temperatures	87	100	42
% change	65	59	156

^aAn average thickness change of 1 m of ice per year is equivalent to 49.3 km³ of ice per year or, equivalently, a yearly average exchange rate of 0.472 x 10¹² watts.

By coupling such models with treatments of the ocean, quantitative insights may be obtained. It also appears that due to the variety of complex dynamical processes, specifying ice fluxes and transport for use in ocean circulation modeling will leave out many major feedbacks that may occur, and hence, sweep aside the interesting feedback effects of climatic change.

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COMMENTARY ON SEA ICE FEEDBACK

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Perhaps the most studied mechanism by which sea ice affects estimates of global warming is the so-called "ice-albedo" feedback. The basic mechanism of ice-albedo feedback, recognized by Croll in 1875, according to Ingram et al. (1989), owes its existence to the simple facts that snow and ice have higher average short-wave albedo than land and water, and that snow and ice exist only at freezing and sub-freezing temperatures. These ideas are illustrated roughly by the data in Table 1, showing short-wave albedos for various surface conditions. If the surface insolation is fixed, the surface temperature T_s at a point increases if the albedo increases. If the temperature reaches the freezing point, energy balance is achieved by melting. The surface area covered by highly reflective snow and ice decreases as melting proceeds, thereby decreasing the area-average albedo of a fixed region that initially contains ice. If these surface albedo changes manifest themselves as change in the planetary albedo, then there is a positive feedback loop (Fig. 1).

The simple ideas embodied in this feedback may be illustrated mathematically with the aid of a zero-dimensional model for planetary radiation balance:

$$C \frac{\partial T}{\partial t} + \sigma T^4 = \frac{S_0}{4} (1 - A_p), \quad (1)$$

where T is the equivalent black-body temperature of the planet, C is the planetary heat capacity appropriate to T , A_p is the planetary albedo, σ is the Stefan-Boltzmann constant, and S_0 is the solar constant. Greenhouse effects may be illustrated in this simple framework by introducing the effective optical depth τ for absorption of long-wave radiation in the

Table 1. Total Albedo as a Function of Ice Type and Cloud Cover During the Summer Melt Season (After Genfell and Maykut, 1977). From: Maykut (1986).

Ice Type	Clear	Overcast
Old snow	0.63	0.77
White ice	0.56	0.70
First year blue ice	0.25	0.32
Mature melt pond	0.22	0.29

atmosphere. At equilibrium, the black-body temperature T may be related to the surface temperature T_s , according to:

$$T^4(1 + \tau) = T_s^4. \quad (2)$$

To complete the specification of the feedback, one postulates some dependence of planetary albedo on the average surface albedo, and therefore, on the surface temperature, i.e., $A_p = f(T_s)$, where f is a function, perhaps like the curve shown in Fig. 2. The negative slope of the curve reflects the hypothesis that planetary albedo decreases with increasing temperature, because the surface area covered by ice and snow decreases as the surface temperature increases. Therefore, the global average surface albedo decreases.

The climatic equilibrium temperature for this model is found by setting the time derivative term to zero:

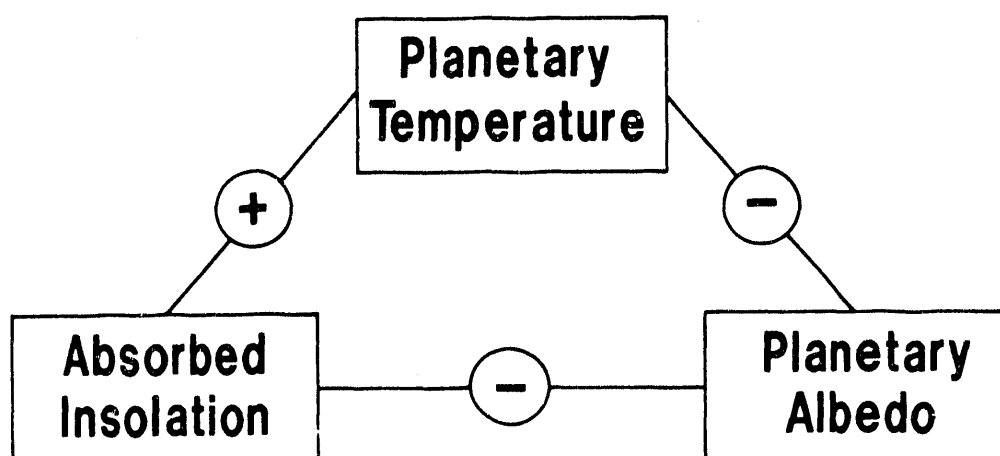


Fig. 1. Elementary albedo feedback.

$$\frac{\sigma T^4}{(1 + \tau)} = \frac{S_0}{4} (1 - A_p), \quad (3)$$

where the subscript s has been dropped for brevity. The two sides of equation (3) are referred to as "LHS" and "RHS." If LHS and RHS are plotted as curves on the same graph, with T as abscissa, the equilibrium climates correspond to points where the curves intersect. If the planetary albedo varies as shown in Fig. 2, there may be three such intersections, that is, multiple climate equilibria, a result first noted independently by Budyko (1969) and Sellers (1969). Fig. 3 illustrates their basic ideas. Three LHS curves are plotted on the figure, corresponding to three values of τ , and yielding one (cold), three and one (warm) climate, respectively. The equilibrium values of temperature T and planetary albedo A_p for the case $\tau = 0.7$ are denoted T_1 , T_2 , T_3 , A_1 , A_2 , and A_3 on the figure.

In this model, increasing atmospheric carbon dioxide corresponds simply to increasing τ , shifting the LHS curve downward. Therefore, the intersections between LHS and RHS shift to the right, that is, toward higher surface temperature. Wherever the RHS exceeds the LHS, there is a net energy input to the planet, so the temperature must increase, and vice versa. Hence, the intermediate climate (T_2 with $\tau = 0.7$) is unstable, and the two extreme climates (T_1 and T_3) are stable. A sufficiently large change in τ might cause the complete disappearance of a stable climate, and a discontinuous shift to a very different stable climate.

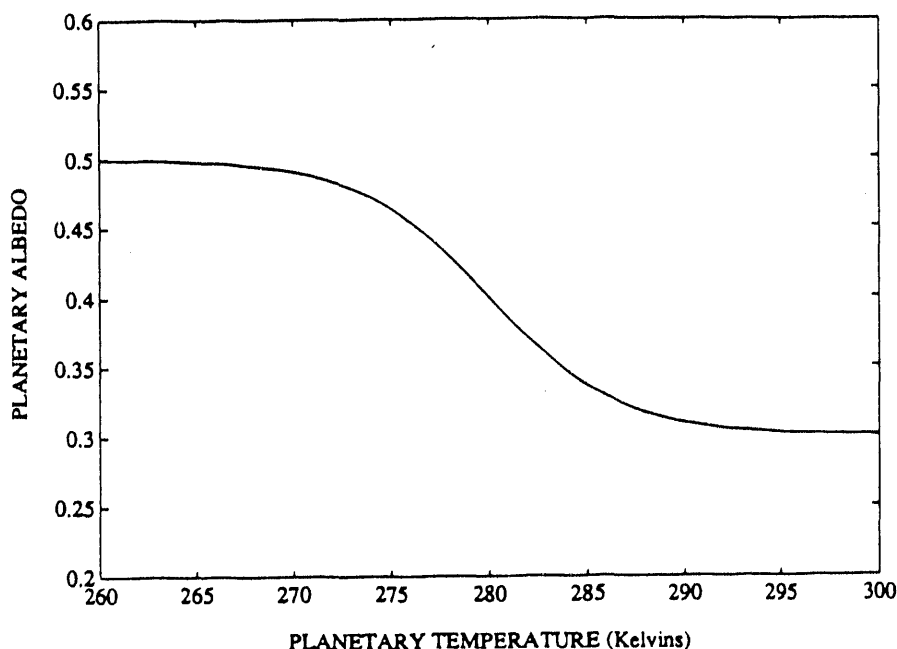


Fig. 2. Schematic albedo-temperature curve.

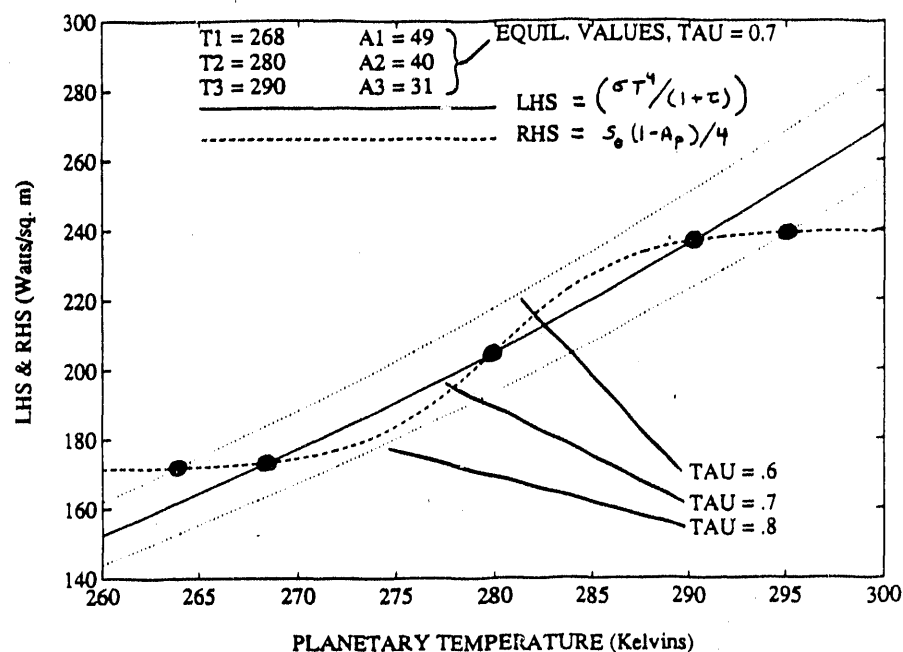


Fig. 3. The Budyko-Sellers Concept: intersections are climate equilibria. Three LHS curves, corresponding to three values of τ , yield one (cold), three and one (warm) climate, respectively. The equilibrium values of temperature T and planetary albedo A_p for the case $\tau = 0.7$ are denoted T_1 , T_2 , T_3 , A_1 , A_2 , and A_3 .

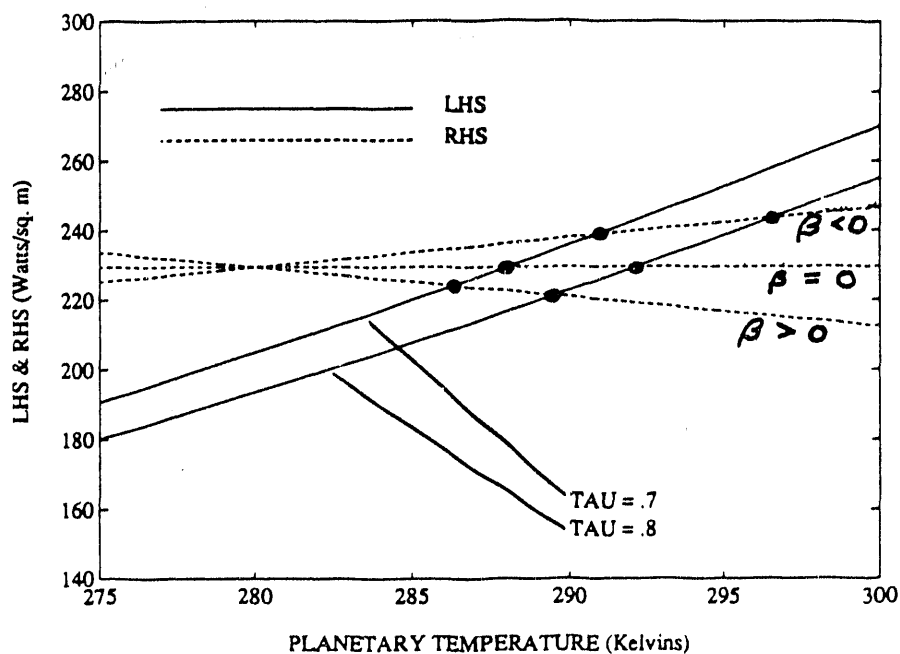


Fig. 4. Greenhouse warming for pos./neg./zero albedo feedback.

If, in the vicinity of a stable equilibrium climate, one linearizes the expression for A_p , one has:

$$A_p = f(T) = \alpha + \beta T, \quad (4)$$

so the RHS becomes a straight line. If β is negative, then the slope of RHS is positive, corresponding to positive feedback, and vice versa. The case $\beta = 0$ illustrates the absence of ice-albedo feedback. Consider two different LHS curves (two values of τ), corresponding, say, to atmospheric CO_2 concentrations equal to and double the pre-industrial global average levels. The difference between the equilibrium temperatures is proportional to the horizontal distance on the graph between the intersections of LHS and RHS (Fig. 4). Thus, the "global warming" in this simple model is largest when $\beta < 0$, that is, for positive ice-albedo feedback, and smallest when $\beta > 0$. In the latter case, other elements of the climate system, such as clouds, must interact in such a way as to over-balance the change in surface albedo due to melting snow and ice, leading to an overall negative feedback.

Having seen how greenhouse warming might be amplified by positive ice-albedo feedback, let us consider the relationship between equilibrium, in a climate model, and the sensitivity of the climate to an imposed change in concentration of CO_2 (here represented by τ). The elements of the physical model are as follows:

- (1) An optical depth, of the assumed form $\tau = \tau_c + \tau_o$, where τ_c is the contribution of carbon dioxide and τ_o is the contribution of all other gases, including water vapor, which we assume temporarily to be constant.
- (2) A planetary albedo $A_p = \alpha + \beta T$, with $\beta < 0$, assumed to represent the effects of sea ice melting at higher T .

S_0 and σ are taken to be physical constants.

Following the typical program, once the physical equations have been developed, one seeks an equilibrium solution for the present climate, to be used as a "control" simulation for comparison with later results. But we have one equation involving three unknown parameters: T , A_p , and τ . Given observed values for all three parameters, one could check the internal consistency of the simple model. Alternatively, if only two of the parameters are known, say T and A_p , then the modeler may "tune" his climate by choosing an appropriate value of τ . For instance, the planetary albedo of earth is about 0.33 and its mean surface temperature is about 288 K, which yields $\tau = 0.7$. If neither of the parameters τ and A_p are determined observationally, then the equilibrium equation (3) determines a two-parameter family of values that would fit the observed temperature T .

In the case of more complicated models, such as general circulation models (GCM) that are used to investigate global warming, there are many adjustable parameters and, in principle, many observational constraints. Typically, however, not all of them are investigated explicitly to determine the fidelity of the control climate. In typical models, the albedos assigned

to sea ice in its various states, and the fraction of precipitation that falls as snow are important adjustable parameters that affect the overall ice-albedo feedback.

The manner in which these uncertainties can affect estimates of feedbacks and climate sensitivity may be illustrated with the simple model. If T represents the climate, and r_o represents the effect of CO_2 concentration on long-wave absorption, then the quantity of primary interest is the derivative dT/dr_o , evaluated at the present climate. Making use of equations (3) and (4) yields:

$$\frac{dT}{dr_o} = - \frac{\frac{S_o}{4} (1 - A_p)}{4\sigma T^3 + \frac{S_o}{4} (1+r)\beta} \quad (5)$$

Suppose that we had regarded both A_p and r as tuning parameters, to be chosen so that the simple model mimics the present climate. Then the observed value of T would have determined the product $(1+r)(1-A_p)$, but not the individual values r and A_p that are needed in equation (5) to find the greenhouse sensitivity. This example illustrates how it is possible to have the relevant physical processes represented in the model, and to fit the present observed climate correctly, by canceling errors of opposite sign. In other words, one has the right result for the wrong reasons, and one does not, in general, obtain the correct derivative dT/dr_o from a model so fitted. In fact, there is a two-parameter family of sensitivities, dT/dr_o , consistent with both the model physics and the tuned equilibrium climate. Furthermore, an additional parameter β appears in equation (5). β is in a class of parameters that cannot, in principle, be determined from a single, steady state climate, no matter how well it is observed. That is, because β represents the response, during climatic change of planetary albedo to global mean temperature change. Therefore, one has either to observe two different climates, or to develop unassailable physical models for the relationship between planetary temperature and albedo, to evaluate the greenhouse sensitivity with confidence. In practice, parameters such as β are often assigned on the basis of observed spatial covariation between variables like temperature and albedo (Budyko, 1969).

So far I have concentrated on the feedbacks associated with A_p , that is, on the RHS. However, it is well established that the atmospheric optical depth r participates in feedbacks too (Raval and Ramanathan, 1989). For instance, $r_o = g(T)$ might represent the feedback between water vapor and the greenhouse effect. In evaluating the overall greenhouse sensitivity, as in equation (5), one would now obtain additional terms, such as:

$$\frac{\partial g}{\partial T} \frac{dT}{dr_o} \quad (6)$$

The dependence of g on T in a GCM involves many other variables and parameterizations, such as precipitation rates, cloud amounts, and convective adjustment. These processes would influence the LHS curve, so that the slopes of both curves are needed to assess the overall amount of greenhouse sensitivity.

According to the procedures outlined for this workshop, one should estimate the overall albedo feedback by computing the greenhouse sensitivity with β set to zero and then again with β set to some standard value β_0 . The difference $dT/dr_0|_{\beta=0} - dT/dr_0|_{\beta=\beta_0}$ is the sensitivity of greenhouse temperature change to ice-albedo feedback. The preceding illustrations show that feedbacks inherent in both the LHS and the RHS will affect this comparison.

The discussion of the simple zero-dimensional model shows that to evaluate the ice feedback, it is necessary, but not sufficient, to simulate a realistic control climate. Moreover, it is necessary to represent explicitly the physical processes that determine the state of the ice, because parameters such as β cannot be determined empirically from observations of a single climate equilibrium state. Therefore, it is essential to model the sea ice and its determinative processes in a way that facilitates comparison with a large number of observational constraints.

An important goal is to evaluate in more realistic settings the feedback due to sea ice, and to place narrower confidence limits on the estimates of this feedback. Several issues are related to this goal:

1. Direct estimates of feedback based on measurements.
2. Estimates of the feedback from GCM output.
3. Dependence of model feedback on the representation of the state of sea ice and physical processes affecting sea ice.
4. Steps needed to reduce the uncertainty in the feedback.

To the best of my knowledge, the direct estimation of overall feedback due to sea ice remains to be accomplished. The very effective methods employed by Raval and Ramanathan (1989) to evaluate water vapor feedback and cloud radiative forcing should be investigated for extension to sea ice. The approach might involve evaluating the net short-wave and long-wave radiative fluxes at the top of the atmosphere, using satellite data, and evaluating the surface long wave exitance, based on a surface temperature. These evaluations would be worked out for clear-sky pixels, in high-latitude, marine areas. Pixels with sea ice would be distinguished from those with open water. The objective of this work would be to estimate the direct spatial effect of sea ice, and all atmospheric phenomena associated with its presence, on the local greenhouse effect, the local planetary albedo, and the local net radiation of the earth-atmosphere system. The presence or absence of sea ice can be determined to sufficient accuracy from SMMR or SSM/I passive microwave data. Polar orbiters can provide the data from which to estimate fluxes at TOA. The surface temperature in clear sky areas can

be estimated from IR-window radiance measurements from space, and related to surface truth data from drifting buoys. It is difficult to distinguish cloudy and clear pixels in ice-covered regions, but this task can be accomplished with reasonable accuracy using interactive computer analysis of multiband satellite imagery. Presently, it is not possible to make this distinction accurately with a fully automated computer algorithm.

Many investigators note the sensitivity of their GCM results for CO₂ doubling experiments to sea ice effects (Manabe and Stouffer, 1980; Washington and Meehl, 1984). A recent study discusses the evaluation of a sea ice feedback parameter λ_i in different GCMs, and its sensitivity to the interactions among cloud and ice (Ingram et al., 1989). These authors report a range of λ_i from .09 to .31 Wm⁻² K⁻¹, comparing simulations from their own (UKMO) model that differ according to the interactive processes that are allowed to operate. Other models cited by Ingram et al. yield values as large as 0.7. Fig. 5 shows that the change in planetary albedo at individual grid points where sea ice occurs in the UKMO model is correlated with the change in surface albedo for a CO₂ doubling experiment. With fully interactive clouds, the slope or proportionality factor, is 0.178. If only those grid points with clear sky are considered, the slope becomes 0.775. So, while the effects of clouds in this model do not negate the ice-albedo feedback, they weaken it considerably. The main reason for this effect is that the clouds simply obscure the surface, in both the control and double CO₂ runs, so that changes in surface albedo have a relatively small effect on changes in the planetary albedo. These results show that GCM studies should address the accuracy with which clouds are simulated in high latitudes, so that the net feedback among cloud and ice can be evaluated with confidence. These results also show, in a more realistic model, how the proper evaluation of feedback requires realistic simulation of clouds in the control climate. To obtain accurate values for the surface albedo change, it is necessary to assess the accuracy with which the processes that determine sea ice areal coverage and ice surface albedo are simulated in the model. These processes include not only the surface energy and radiation balances, but also the mass balance sea ice.

To model the climate response with confidence, one wants to know that the sea ice is simulated realistically in the present climate. A difficulty here is that the specification of sea ice state in typical GCMs is so different from the observable state of sea ice that only very few observational constraints can be imposed. For instance, the all-important average surface albedo over areas the size of a GCM grid box (>(200 km)²) changes dramatically through the spring and summer, but is not known to better than 10% accuracy. Moreover, real sea ice is fissured and cracked by leads and polynyas. These areas have radiative properties completely different from ice in both summer (lower albedo) and winter (higher surface temperature). In fact, the ice in a region comparable in size to a satellite sensor footprint or GCM grid box has a wide range of thicknesses, typically between zero and about 10 meters, as expressed by a thickness distribution function (Thorndike et al., 1975). The distribution changes in response to several processes, including ice transport across thickness gradients, melting and freezing, and mechanical redistribution, wherein thinner ice breaks and piles

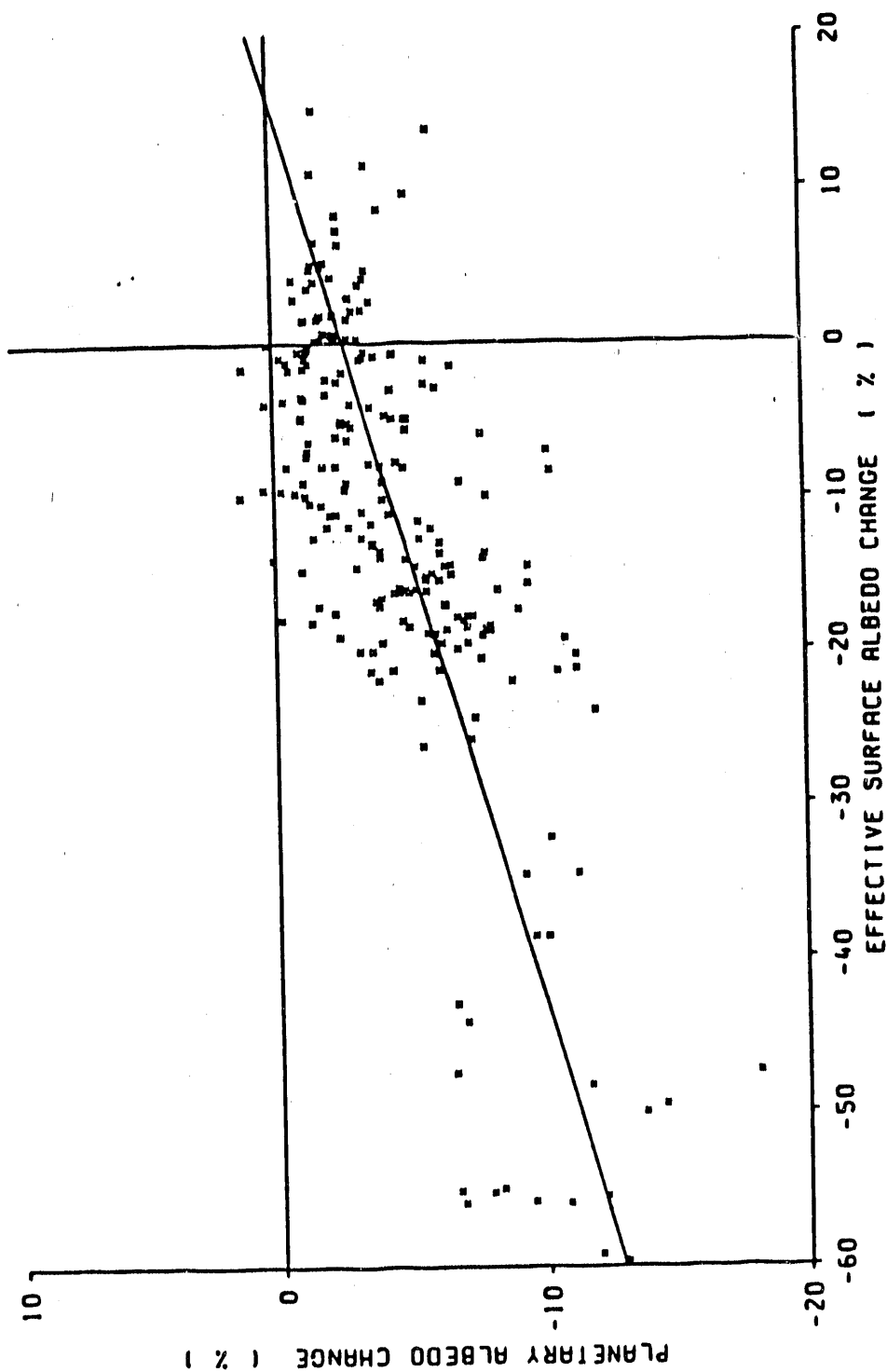


Fig. 5. Scatterplot of change at each sea ice point of the effective surface and planetary albedos, averaged over December, January, and February, in the experiment of WM87. All points which had sea ice and received sunlight at any time in either of the integrations are included. The area-weighted root-mean-square best fit straight line is also indicated. From: Ingram et al., 1989.

to form pressure ridges, and fissures widen to add new area of open water. To improve the evaluations of climate feedback using GCMs, it is essential that they begin to represent the ice state and processes more realistically, providing further points of comparison between model outputs in the control climate, and observational constraints.

From the viewpoint of surface energy balance, it is important to distinguish an area of open water in each grid box containing pack ice, and to model the transfer of heat between the open water and the ice. Another crucial element is the snow cover atop the ice, because its presence greatly reduces ice growth in winter (thermal insulation) and delays the transition to low-albedo melting ice in the spring.

Ice transport is important because the real ice edge in many seas of interest (e.g., the Greenland Sea, the Ross Sea) is determined by a balance among melting, freezing, and advection. The latter process is of the same order as the former at the ice edge (cf. Walsh et al., 1985; Moritz, 1988). If GCMs are tuned to produce a realistic ice edge position in the absence of ice transport, then the feedback due to sea ice area change has built-in inaccuracies: the ice balance in the control climate would be thermodynamically incorrect to first order. Inclusion of ice transport will become even more important as the modeling proceeds to coupled ocean-atmosphere GCMs. For example, the transports of salt, freshwater and heat by sea ice moving into the Greenland Sea from the Arctic Ocean are seen as important controls on the stratification where North Atlantic Deep Water is formed. It is perhaps the leading hypothesis that the "Great Salinity Anomaly" of the 1960s and 1970s in the North Atlantic (Dickson et al., 1989) originated as a sea ice transport and mass balance anomaly in the Greenland Sea (cf. Aagaard and Carmack, 1989). Fig. 6 (Untersteiner and Carmack, in press) illustrates schematically how transports of ice and water in the Arctic Ocean eventually influence the convective overturning in the high North Atlantic.

Cattle (1988) notes that the surface radiation fluxes over pack ice in the UKMO model vary by as much as 40 W m^{-2} , depending on whether Arctic clouds are prescribed or computed interactively. He also emphasizes the lack of reliable, concurrent climatological data on the surface radiation budget, and cloud cover, needed for comparison with the control climate simulations. This paucity of radiation and cloud data impedes progress on evaluating and improving model parameterizations of cloud processes and radiative transfer. Uncertainties of this size translate into large errors in the estimates of overall ice feedback, as illustrated, for instance, by the gross underestimation by most GCMs of the average sea ice thickness in the Arctic Ocean. The net surface radiation dominates the thermodynamic forcing on sea ice in all seasons. Therefore, it is important to pursue strategies for acquiring accurate data on the surface radiation budget in pack ice, together with concurrent atmospheric temperature and humidity profiles, and cloud conditions. These surface measurements must be acquired concurrently with satellite data, with the purpose of calibrating algorithms for monitoring the Arctic from space in the longer term. In the meantime, the surface-based data would provide extremely useful constraints for GCM modeling of sea ice feedback.

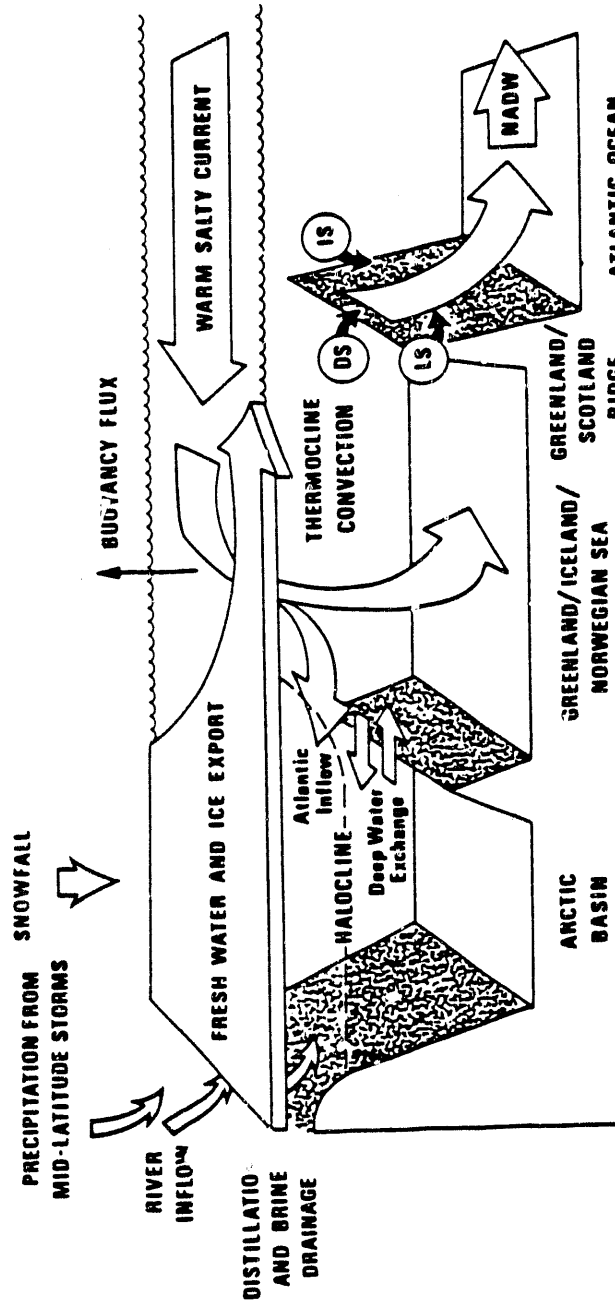


Fig. 6. Processes of ice and water transport in the Arctic Ocean, and their relation to North Atlantic Deep Water (NADW). From Untersteiner and Carmack (in press).

STEPS NEEDED TO REDUCE UNCERTAINTIES

The mass balance of sea ice in large scale regions, such as the entire Arctic Ocean, or individual seas such as the Greenland, Bering, Ross, Weddell Seas, should be estimated from observations, distinguishing local and advective contributions, and the roles of radiation, ocean heat flux, and other fluxes. Such a mass balance data set would provide a strong litmus test for calibrating GCMs based on the present climate: Do they get correct ice edge position for the correct reasons? Do they get correct ice thickness for the correct reasons? On passing such a test, the models could be used with more confidence for investigating the "What if?" scenarios of interest in the greenhouse gas question. At present, the most important element missing from the representation of sea ice state in GCMs is the thickness distribution, including thin ice and open water. Recent developments in moored instruments, such as Upward Looking Sonars, can monitor the thickness distribution to obtain statistics that could be used to constrain estimates of GCM-produced mass balance.

Direct estimates of the effects of sea ice radiative forcing should be made on the basis of existing satellite data and surface temperature observations. Although this kind of analysis refers to spatial variations rather than climatic change, they might put us on a much firmer footing in deciding whether ice-albedo feedback should indeed be considered positive, and how large it might be.

New measurements of surface radiation fluxes, atmospheric temperature and humidity profiles, cloud properties, and average ice-surface albedo and temperature are needed over the Arctic pack ice. These measurements would be used to establish a more reliable radiation climatology as a constraint for modeling efforts, and are essential for calibrating satellite algorithms, on which long-term monitoring efforts must be based. The elements of such a program should include one or more surface stations on the Arctic pack ice for at least one year, equipped with sets of calibrated radiometers that resolve short-wave, long-wave, and spectral irradiances, together with regular radiosonde measurements, cloud observations and all-sky photography. Such a station might be established in cooperation with Soviet scientists involved in the ongoing "North Pole" (NP) station series, or with the Canadian Polar Continental Shelf Project, which has outfitted drifting ice islands. Spatial average surface temperature and albedo estimates should be determined at a few pre-selected phases of the seasonal cycle, using low-flying aircraft with downward-looking sensors. The activity has been recommended by the WMO/WCRP Working Group on Sea Ice and Climate (WCP, 1989), and again by the Workshop on Arctic Systems Science, Ocean-Atmosphere-Ice interactions (JOI, in prep).

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GLOBAL CLIMATE FEEDBACKS: CONCLUSIONS AND RECOMMENDATIONS
OF THE JUNE 1990 BNL WORKSHOP

The issue of global change initiated by increases in the concentrations of CO₂ and other greenhouse gases is a scientific issue with major policy implications. The best means to examine the response of the Earth's climate to prospective perturbations in radiative forcing caused by such changes, and to other industrial activities, is modeling, specifically by means of general circulation models (GCMs) of the Earth's atmosphere and of the coupled atmosphere-ocean system. However, for these models to be useful, it is imperative that they accurately represent all the significant responses of the climate system to changes in forcings, not just direct responses, but also indirect responses, i.e., feedbacks. To the extent that some of these feedbacks are not represented in the models, or are inaccurately represented, the models will yield descriptions of the response of the climate system to perturbations that are erroneous in magnitude, and perhaps, even in sign.

At the most simplistic level, these feedbacks will alter the response of global mean temperature to a specific change in forcing. However, as discussed at the workshop, responses of the Earth's climate to a change in forcing are much more varied than can be captured in a single variable such as global mean temperature, encompassing changes in the cloud amount, cloud vertical and horizontal distribution, and cloud microphysical properties; the amount and pattern of precipitation; the snow and ice cover of land; the amount and distribution of sea ice; the atmospheric and ocean circulations patterns; the amount, type, and distribution of surface vegetation; photosynthesis and respiration rates. All of these changes will alter the response of the Earth's climate system to any primary changes in forcing from that caused only by the primary change in forcing.

The purpose of this workshop was to identify the feedbacks inherent in the Earth's climate that actually or potentially govern the system's response to perturbations, to identify gaps in knowledge that preclude the accurate representation of these feedbacks in models, and to identify research required to represent these feedbacks accurately in models. It should be noted that the United States and the international research community is already undertaking a greatly enhanced research effort to obtain a better

understanding of and an improved ability to describe global climate change. The U.S. effort is described in detail in the Committee on Earth and Environmental Sciences (CEES) document, "Our Changing Planet - The FY 1991 Research Plan of the U.S. Global Change Research Program."* The present workshop thus focuses on identifying research which, in the view of the participants, is important to an accurate description of global climate change but is inadequately represented in research that is presently planned.

The feedback processes are essentially a response to a perturbation. The reference perturbation for the workshop assumed an increase in the atmospheric concentration of greenhouse gases. If we assume a doubling of the CO₂ equivalent gases, the forcing function will amount to about 4 W m⁻².

The most significant readjustment to the perturbation is the increase in the amount of atmospheric water vapor. Further readjustments could occur due to changes in cloud, rainfall, and snow patterns, as well as increases in oceanic evaporation and terrestrial evapotranspiration. Rising temperatures may alter the ratio of plant respiration to photosynthesis; rising CO₂ levels directly affect plant productivity as well. The responses of the vegetation can alter terrestrial carbon storage, leading to further shifts in atmospheric gas concentrations. Sea ice can be transported from one region to another, with melting and freezing occurring at different locations. Such transport tends to create net imbalances in salt fluxes in the ocean. Resulting changes in ocean thermohaline circulation can themselves initiate climate changes. Climate changes, including not only temperature changes but also changes in precipitation/evaporation and wind regimes, will undoubtedly affect physical and biogeochemical processes in the upper ocean layers and hence affect the dynamics of air-sea CO₂ exchange processes.

Of the many feedback processes that were discussed, all phases of water were identified as being the most important contributors to moderating or amplifying climate perturbations that might be initiated by increased atmospheric trace gases.

The major conclusion of the workshop on feedbacks was that the only feedback where sign and amplitude is confidently known is that of water vapor; specifically, an increased concentration of water vapor leads to a positive amplification of the magnitude of the global temperature change in response to a heating perturbation by a factor of ~1.6. At present, there is no confidence in the sign of the feedbacks associated with changes in cloud cover and cloud properties, ocean sources and sinks of CO₂, changes in surface hydrology (including vegetation transpiration, and albedo, changes in land surfaces, and changes in sea ice cover); snow feedback is almost certainly positive, but the magnitude of the feedback is unknown.

*This document may be obtained from the Committee on Earth and Environmental Sciences, c/o U.S. Geological Survey, 104 National Center, Reston, VA 22092.

Within the next decade, the international program, and particularly the U.S. program as described in the Committee on Earth and Environmental Sciences (CEES) document "Our Changing Planet - The FY 1991 Research Plan of the U.S. Global Change Research Program," should provide more definitive values for feedbacks and materially reduce the present uncertainties.

The individual workshop panels made several recommendations on research to accelerate the needed reduction of feedback uncertainties, and indicated how the recommended research can be coordinated with the CES plan. The recommendations are given in priority order.

FINDINGS AND RECOMMENDATIONS OF THE PANEL ON ATMOSPHERIC FEEDBACKS

The increase in the greenhouse effect due to increased concentrations of CO₂ will be enhanced by increased concentrations of water vapor; the magnitude of the change is expected to depend on the extent of temperature rise, because water vapor is itself strongly controlled by surface temperature through the Clausius-Clapeyron equation. Other feedbacks involving water occur through clouds, surface hydrology, sea ice, and snow. Magnitudes and signs of these feedbacks are not certain. Surface hydrology includes considerations of changes in albedo and evapotranspiration. The sea ice feedback is especially uncertain because as sea ice retreats, and sea-surface albedo decreases, these conditions seem to give rise to low, highly reflective stratus clouds.

The panel recommended several field research projects and model inter-comparisons that would enhance our ability to describe these feedbacks. The intercomparisons would also provide tests of the accuracy of the treatment of these feedbacks in the models, thereby lending enhanced confidence in the use of these models to address "what-if" questions pertinent to the future climate. Emphasis was placed on projects that could realistically be undertaken on a time-scale of five years or less. Based on stated findings the following classes of research projects were recommended:

1. It is widely expected that the EOS program will be the primary satellite observation program. An important pre-EOS activity is the Earth Probes program that will place smaller, specialized platforms in orbit up to six years earlier than the projected 1998 first launch of an EOS satellite. The first DOE Atmospheric Radiation Measurement (ARM) site will be operational in 1992, with others following in subsequent years. Thus, there will be no overlap with at least the first ARM site for about six years. Therefore, ARM satellite observations will have to rely on other satellites for a significant part of the ARM program as presently planned.

The panel recommended that a coordinated measurement program be initiated to simultaneously observe atmospheric radiation at the surface, at the tropopause, and at the top of the atmosphere. ARM sites would be suitable for surface measurements. Top-of-troposphere measurements might be made with an unmanned aircraft platform; it is understood that DOE is evaluating use

of such platforms in ARM. Short- and long-wave radiation should be measured ideally using instruments closely modeled after the ERBE scanners.

Much discussion was devoted to the need for high quality calibrated satellite measurements of top-of-the-atmosphere radiation with broad-band high spatial resolution scanners. These needs include: (1) establishing continuity between ERBE and CERES/EOS; (the ERBE satellite failed early in 1990; the CERES/EOS is not scheduled to be on line until 1998 at the earliest, but may well be delayed based on experience with earlier large NASA projects); (2) extension of measurements used by Raval and Ramanathan,* in both clear and cloudy regions; (3) determination of absolute shortwave flux to the surface as input to ocean models and coupled ocean-atmosphere models; (4) correlation of satellite measurements with ARM sites to establish the ability of ARM sites to serve as ground truth locations for the satellite measurements, and to complement radiation measurements at the surface and top of the troposphere; and (5) definition of aspects of later EOS flights, particularly to serve as prototype and test-bed cases for the scanners to be used on CERES.

The choice of a low polar orbit appears best for making most of the measurements and has the advantage of giving global coverage with a frequent revisit time. However, observing the development of convective activity would also be very useful, but would require a geosynchronous orbit, which would not give a global view. Both objectives could be met with multiple satellites. It may be useful to explore international cooperation to meet the needs of the coordinated measurement program in the nineties. The desirable measurement capabilities for the satellite are listed on the following page.

The acquisition of even a portion of these measurements equivalent to ERBE data would be very valuable if obtained simultaneously with ARM measurements.

2. The panel endorsed the Global Energy and Water Cycle Experiment (GEWEX; NSF/NOAA/NASA/DOI). The scientific objectives of GEWEX include describing and understanding the transport of water (vapor, liquid, and solid) and energy in the global atmosphere and at the underlying surface, and developing methods of predicting changes in the distribution of water within the global atmosphere and on the underlying surface, which may occur naturally or through human activities. However, only modest support is scheduled to be provided in FY 1991 for planning and development activities in preparation for GEWEX field projects, process studies, and modeling research. The panel recommended an enhancement of the GEWEX field program, possibly by coordination with DOE's ARM program on measurements of water (all phases) as function of position (3-dimensional); these measurements are especially important in the upper troposphere, where water vapor forcing of cirrus is the key to accurate understanding of cirrus forcing. Present knowledge above 500 mb is especially poor. Rawindsondes are notoriously unreliable at the low concentrations of water vapor characteristic of these

*Raval and Ramanathan, this volume.

<u>Parameter</u>	<u>Desired</u>
Water Vapor Profile	
<u>0-10 km</u>	
accuracy	±20%
vertical resolution	2 km
horizontal resolution	2.5°x2.5°
<u>10-30 km</u>	
accuracy	±50%
vertical resolution	2 km
horizontal resolution	2.5°x2.5°
Temperature Profile	
accuracy	0.5K
range	0-15 km
vertical resolution	5 km
horizontal resolution	2.5°x2.5°
Clouds	
cover	2%
top height	±0.25 km
emission temperature	±0.5K
albedo	±0.01
water content	±0.5 Kg/m ²
Radiation Flux	
short-wave acc.	2%
long-wave acc.	2%
long + short acc.	2%
Aerosols	
accuracy	5%
vertical resolution	1 km
horizontal resolution	2.5°x2.5°
Planetary Boundary Layer	
height	±0.05 km
horizontal resolution	50 km

altitudes. The recommended time-scales included resolution from daily to monthly-mean for extended periods.

3. The CEES report states that NSF, NOAA, DOE, and DOD will provide support for and conduct investigations of the chemical transformations and modeling studies that determine the fate of natural sulfur emissions. The

emphasis will be twofold: (i) fundamental laboratory characterization of the sequential transformation of the sulfur compounds, and (ii) testing the laboratory data against field observations of the abundance of these compounds. The workshop recommends a DOE enhancement of the program by examination of the relative role of Cloud Condensation Nuclei (CCN) derived from biogenic (dimethylsulfide, DMS) versus industrial (SO_2) emissions. The number density of CCN strongly influences shortwave forcing by clouds. It has been argued that CCN resulting from industrial emissions may have significantly perturbed shortwave forcing in industrial areas such as the eastern U.S., and may thereby be obscuring early detection of CO_2 induced warming. Wide areal coverage is needed because dimethylsulfide emissions are a strong function of ocean productivity and location. Chemical or isotope studies may distinguish industrial and biogenic sources.

4. It is recommended that the ARM program include the measurement of sub-visual cirrus clouds. Studies of the radiative properties of such cirrus by aircraft measurements above and below such cirrus layers are necessary, perhaps in conjunction with lidars. Western tropical Pacific areas are important for such studies, particularly extensive thunderstorm generated anvils which can act as climate forcers.

5. Measurement for photosynthetically active radiation, 0.4 to 0.7 μm , would be valuable in order to determine how cloud cover changes would affect agricultural productivity. It is recommended that the ARM program include a measurement of surface insolation, especially in the presence of clouds.

6. Large geophysical-scale and long-time monitoring of CCN and IFN is necessary to understand the present concentrations and the potential for future modifications. The panel suggested that within the Long Term Observing Planning effort of NOAA, attention be given to developing an improved climatology of CCN and Ice Forming Nuclei (IFN).

These studies, especially measurements of water substance, should be considered as a test-bed for models. The western tropical Pacific was suggested as an important site because of the large amount of convection, high water vapor concentrations, and sensitivity to El Nino perturbations. On the other hand, high latitude sites are highly sensitive to water vapor positive feedback, and thus might provide earlier indications of greenhouse warming. It was also suggested that there be a strong link to atmospheric chemistry programs. In addition to sulfate, it was observed that increments in tropospheric O_3 concentrations may be contributing to IR forcing, in an amount comparable to incremental changes in CO_2 and methane.

In addition to process studies, these measurements would provide data sets against which to compare GCM performance, enhancing the value of proposed GCM intercomparisons. Suggested intercomparisons include:

1. Snow feedback and its interaction with other feedbacks.
2. Distributions of water vapor, liquid water, and ice in the atmosphere.

3. Outgoing long- and shortwave fluxes derived from ERBE observations.
4. Latitudinal and longitudinal fluxes of water and energy in the atmosphere, and by implication, energy in oceans.
5. Radiative forcing in a 2 x CO₂ atmosphere, separately in clear and cloudy regions.

FINDINGS AND RECOMMENDATIONS OF THE PANEL ON OCEAN INTERACTIONS AND SEA-ICE RESPONSE

This panel considered the major feedback effects of the oceans on global climate which require an understanding to improve the predictive capability of climate models. It is assumed that advanced climate models will couple atmospheric and oceanic processes, and include realistic representations of biogeochemical processes which are crucial to climate feedbacks. Three processes were identified as being of high priority for research: the uptake and sequestering of CO₂ by the ocean; the transfer of heat across the air-sea interface and subsequent transport into the ocean interior; and the area, extent, and mass balance of sea-ice which dominates sensible heat fluxes at high latitudes. The uptake of CO₂ creates an indirect feedback on climate, modifying atmospheric forcing, due to anthropogenic emissions forcing, and potentially buffering atmospheric CO₂ levels on geological time-scales. Although heat transfer between the ocean and atmosphere is not a feedback process per se, it introduces a time constant into the relationship between atmospheric forcing and response. This delaying effect must be better understood to predict the timing and magnitude of the atmospheric response. The size of this time constant suggests that the ocean may serve as an effective calorimeter for global change. Ocean-atmosphere-sea-ice interactions are crucial to the evaluation of global climate response because current climate models represent a wide range of sensitivity.

The panel made a number of specific recommendations for observational and modeling efforts to increase an understanding of key processes.

1. It recommended that measurement of fluxes of CO₂ across the Air-sea Interface be supplemented.

The ocean influences climate by absorbing part of the CO₂ added to the atmosphere by human activities. The panel recommended research to better define the rate of CO₂ uptake and the mechanistic processes that regulate this rate, which must be understood if the evolution of the atmospheric CO₂ concentration is to be reliably predicted. This effort, as supported by DOE, has had, and should continue to have, two aspects.

The first aspect involves research to measure and model fluxes of CO₂ into and within the oceans. Current work should be continued, such as measuring the distribution of total CO₂, total alkalinity, and pCO₂ in oceanic surface waters, measuring the distribution of total CO₂ and total

alkalinity within the ocean, and development of three-dimensional ocean GCMs. Research should be extended to include ^{14}C and other tracers which can improve understanding of carbon transport, and can be used to validate and calibrate models. Studies of the time-varying distribution of O_2 in air should be undertaken because they will provide a definitive estimate of the net rate of CO_2 addition to the atmosphere and oceans. O_2 can be used in this way because it is linked with CO_2 by combustion, photosynthesis, and respiration, but is far less soluble in seawater, thereby eliminating uncertainties from gas exchange in mass balance calculations.

The second aspect involves experimental and modeling studies to achieve a mechanistic understanding of how ocean chemistry, biogeochemistry, and circulation fix the surface water pCO_2 field and thereby control air-sea fluxes. The southern ocean should have first priority. In addition to being an important site of anthropogenic CO_2 uptake, it is thought to have played a large role in regulating ice age pCO_2 , and is considered to be a region where humans can conceivably influence CO_2 removal. One focus of mechanistic research should be in the area of ecosystems and should focus on understanding how ecosystem dynamics limit primary production and inorganic carbon removal, as particulates and as dissolved organic carbon (DOC), from eutrophic surface waters. Regeneration of carbon in particles and DOC, which releases CO_2 that can be mixed back up to the surface, should also be studied. This research should place a heavy emphasis on moored instruments that can return a high-resolution, long-term record of ocean chemistry, biology, and circulation. We also recommend implementation of a large-scale sampling program for the study of the surface water nutrient distributions to be carried out in conjunction with measurements of the carbon system parameters. This work would yield maps of the time-varying nutrient distributions in oceanic surface waters. Such maps would be useful for calibrating and testing models that account for the changing concentration of surface water CO_2 , which is linked to nutrients by photosynthesis and respiration. Finally, the panel recommends *in situ* and satellite studies of ocean color. Measurements of ocean color provide the only possibility for long-term monitoring of variables related to biological productivity that give high spatial and temporal resolution.

The study of paleoclimate has also led to important insights into mechanisms that control climate and ocean chemistry. It is important to be alert to the implications of continued work in this area as a means of improving understanding of oceanic and climate processes.

2. The working group endorsed the NOAA Atlantic Climate Change Program Plan which has the following goals:

a. To monitor, describe, and model the space time variability of the large-scale meridional circulation of the Atlantic Ocean and its role in producing variability of SST, sea ice, and salinity over the Atlantic Ocean on seasonal, annual, and interannual time-scales.

b. To determine the response of the global atmosphere to persistent SST and sea ice anomalies in the Atlantic Ocean, and to develop coupled

ocean-atmosphere models to simulate and predict seasonal to decadal changes over and around the Atlantic Basin.

c. To design a program of observations to monitor the changes in the conveyor belt circulation, and to develop a suitable modeling program to assimilate these observations to help understand the mechanisms that determine the fluctuation of the conveyor belt circulation.

The NOAA program should be supplemented with transient tracer data to improve understanding of deep-ocean circulation.

The ocean offers special opportunities for the detection of global warming because of the favorable signal-to-noise ratio, compared to atmospheric measurements. Roemmich and Wunsch (1985)* have shown that substantial changes are taking place in the North Atlantic. A warming of 0.10° occurred between 1960 and 1980, between 1 km and 2 km over wide areas of the subtropical North Atlantic. It has been suggested that warming at the base of the thermocline can be monitored by remote acoustic methods. This strategy is being tested in ocean circulation models. Conventional, high-resolution hydrographic sections may be as effective for detection of these large-scale changes in ocean temperature.

It is recommended that model studies of greenhouse warming with coupled atmosphere-ocean models be undertaken to determine the best observational strategy. The study of transient tracer data will be useful, because the transient warming will penetrate the ocean in much the same way.

3. It is recommended that sea ice dynamics be included in atmospheric circulation models, especially those to be coupled to the ocean circulation. This will allow feedback effects due to ice transport and lead formation to be addressed, and will help to identify key inadequacies in the high latitude wind fields generated by atmospheric models. It will also clarify how to improve formulations of ice dynamics, thickness distribution, and transport in order to achieve more accurate representation of overall polar feedbacks in coupled models. The modeling experiments will require validation, using time-series data on sea ice thickness distribution and velocity obtained from measurements. Continued model development and verification using observed forcing fields will also be required.

The present uncertainties about ice-albedo and ice-cloud-radiation feedbacks must be reduced by analysis of *in situ* and satellite measurements. The purposes of this effort are: to acquire accurate estimates of surface radiation fluxes, cloud properties, and atmospheric soundings for at least one year on a station in the Arctic Ocean; to analyze these data together with periodic, aircraft-borne spatial surveys of solar radiation and surface temperature, and produce a data set for comparative and calibration analyses of satellite flux, sounder and imager data; and to establish accurate retrieval procedures to determine the net solar and long-wave fluxes at the

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

top of the atmosphere, the upward long-wave flux at the surface, and the net radiative effect of the intervening atmospheric layer. The established procedures should be applied to ERBE and other satellite data to estimate the net radiative feedback effect by comparison of ocean areas with and without sea ice cover. Subsequent studies should use the insights into polar radiative feedbacks, and the calibrated satellite-based radiation data, to improve and test parameterizations of polar radiative feedbacks in atmospheric general circulation models.

These aspects of high priority sea-ice response research should be linked to the developing programs of DOI/USGS (monitoring and research on the behavior of snow and ice, and their coupling with climate), NOAA (role of sea-ice in air-sea sampling), and NSF (Arctic Systems Science).

FINDINGS AND RECOMMENDATIONS OF THE PANEL ON TERRESTRIAL ECOSYSTEMS

Several aspects of biosphere-atmosphere interactions play a role in regulating the climate system. At rapid time-scales, evapotranspiration through terrestrial vegetation is a major factor in controlling atmospheric moisture content on regional-to-continental scales. The water vapor flux influences radiative properties of the atmosphere, surface energy balances, and cloud formation, which is a factor in global albedo. Terrestrial ecosystems play an important role in the global biogeochemical cycles. Seasonality in carbon exchanges between the atmosphere and terrestrial ecosystems cause an annual cycle in atmospheric CO_2 concentration. Terrestrial ecosystems regulate atmospheric CH_4 and N_2O concentrations, and strongly influence emissions of substances (e.g., nitric oxide and isoprene) that are precursors to tropospheric ozone. On longer time-scales, the structure of terrestrial ecosystems complexes (species composition, biomass, turnover rates) change, and these changes feedback on climate.

A broad research question stems from these interactions between the atmosphere and terrestrial ecosystems: How do terrestrial ecosystems modulate the response of the atmosphere to anthropogenic forcing from CO_2 emissions and land use? Specifically, how does the biosphere interact via short-term physiological and biological controls over water, energy, and albedo; and how does the terrestrial biosphere interact with global change via the carbon cycle?

Three areas of activity are timely when addressing this broad question: Whole-Ecosystem Controlled Experiments, Terrestrial-Atmosphere Flux Measurements, and Global Ecosystem Modeling.

1. Concepts, data, and models derived from laboratory studies must be extended to true field conditions, and this can be done with the current technology in Whole-Ecosystem Controlled Experiments. Particular attention needs to be given to those ecosystems with carbon storage and carbon fluxes great enough to have a significant impact on global carbon cycling, should they respond to CO_2 enrichment of the atmosphere or to climate change. In this regard, large stature ecosystems (i.e., forests with large biomass and

slower dynamics, and mid- to high-latitude soil systems with large organic carbon pools) require specific attention.

Because of their relevance to carbon cycle and climate issues, realistic field experiments with elevated CO₂ concentrations are a high priority. Such experiments will clarify the role of terrestrial systems as a sink for CO₂. They will provide data for parameterization and evaluation of models describing the effects of increasing atmospheric CO₂ concentration on terrestrial carbon and the more general characteristics of terrestrial ecosystem carbon dynamics, such as the allocation of photosynthetically fixed carbon when plant community interactions are modulating plant growth. Whole-Ecosystem Controlled Experiments are recommended to test the consistency and robustness of a number of general ecological principles derived during the last decade. Plant physiological and microbial concepts are already being tapped to make ecosystem models of substantially more generality, a critical requirement of global change studies.

2. It is recommended that Terrestrial-Atmosphere Flux Studies be included as a component of sites intended to evaluate the physical and chemical properties of the atmosphere. ARM can contribute to field ecological studies by adding measurements of surface fluxes -- sensible and latent heat, and CO₂ -- at long-term observation stations.

3. The field activities discussed above need to be set into a global perspective via a new emphasis on the development of Global Ecosystem Models. Global ecological models are recommended for predicting changes in the earth system over decadal time-scales due to effects of the biota on surface energy balance, hydrology, and the carbon cycle. The interactive linkage between model development and field studies is as essential to the development and parameterizations of Global Ecosystem Models as is that between GCMs and field-centered data acquisition activities such as ARM. Global ecosystem models with the following attributes are needed:

a. The models must be global in extent and geographically subdivided to about 50 km resolution.

b. The models must incorporate physiological and biophysical interfaces between the atmosphere and terrestrial ecosystems.

c. The models must describe biogeochemistry because of its importance in the global carbon cycle that controls atmospheric CO₂ and CH₄ concentrations.

d. The models must describe population and community processes with time constants ranging from several years to centuries. Community changes may dominate physiological-biophysical and biogeochemical cycling, so that they are critical to understanding options for human uses of ecosystem resources.

The development of global ecosystem models parameterized with field data meshes with the long-term goal of developing a credible 50-year simulation of the Earth system of global climate and global biogeochemical models.

Although many observational, process, and model studies of ecological systems are presently being organized by several agencies, the integrative concepts in the three areas of activity recommended here are not being addressed.

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