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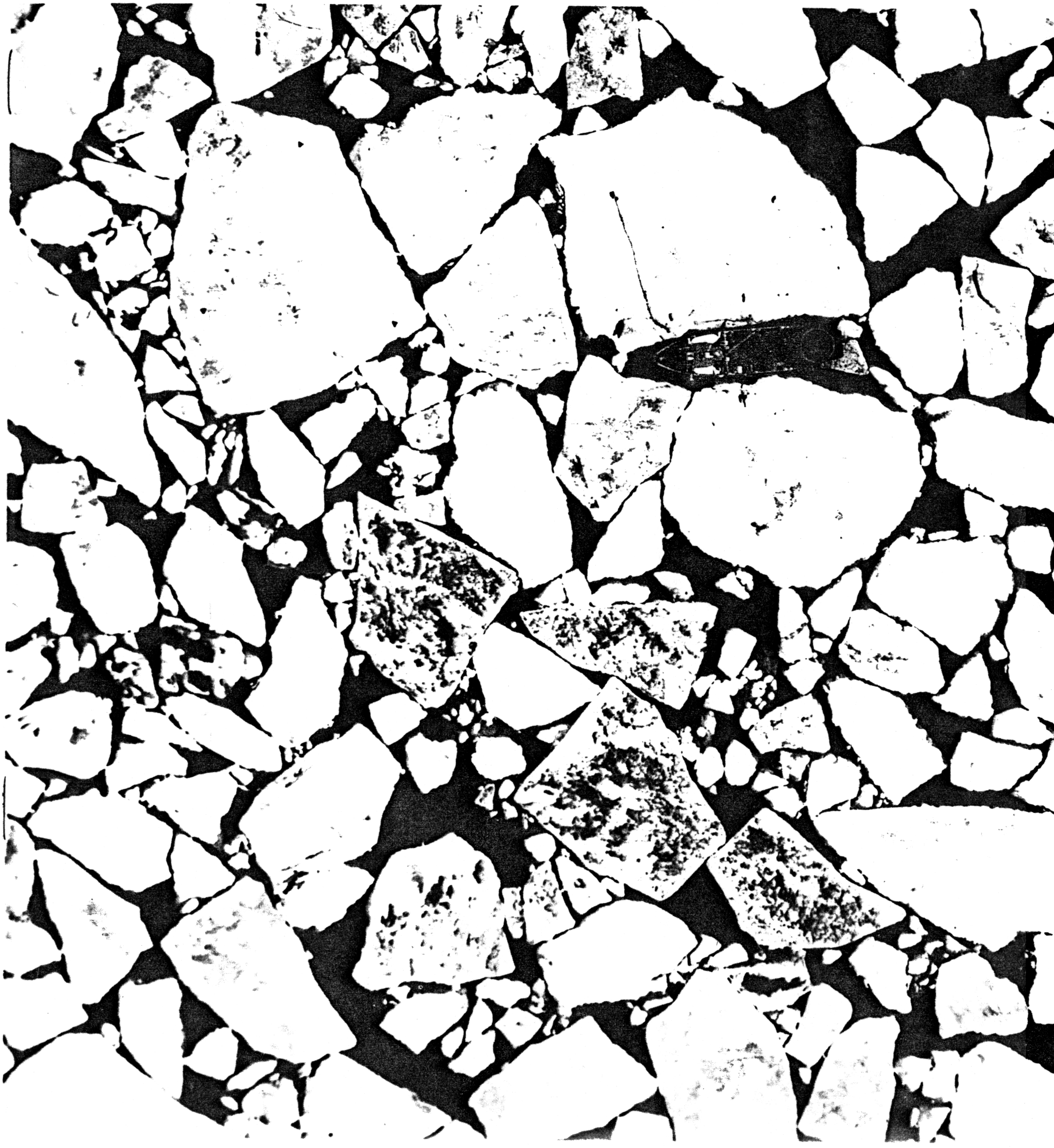

COVER

Polarbjorn (ice-strengthened ship, 50 meters long) moored to an ice floe at 81.1°N, 7°E in the marginal sea-ice zone in the Fram Strait region near Svalbaard (Norwegian Arctic). Photograph was taken at the end of June 1983. See page 489. [Vernon Squire, Scott Polar Research Institute, Cambridge, England]

Ocean Circulation: Its Effects on Seasonal Sea-Ice Simulations

Abstract. *A diagnostic ice-ocean model of the Arctic, Greenland, and Norwegian seas is constructed and used to examine the role of ocean circulation in seasonal sea-ice simulations. The model includes lateral ice motion and three-dimensional ocean circulation. The ocean portion of the model is weakly forced by observed temperature and salinity data. Simulation results show that including modeled ocean circulation in seasonal sea-ice simulations substantially improves the predicted ice drift and ice margin location. Simulations that do not include lateral ocean movment predict a much less realistic ice edge.*

The growth, drift, and decay of sea ice are closely related to the circulation of polar oceans. This is especially true in the Greenland and Norwegian seas in winter where warm currents flowing northward encounter rapidly cooling atmospheric conditions together with sea ice advancing southward. In earlier work on modeling the seasonal cycle of Arctic sea ice, the ocean has been approximated by a motionless mixed layer of fixed depth (*1, 2*) with possibly a small constant heat flux from the deeper ocean, and more recently by a one-dimensional mixed layer of variable thickness (*3*). This approach has also been used in most global-climate-model sensitivity studies of the effect of increasing atmospheric carbon dioxide.

In this report we examine the dominant effects of a more realistic treatment of the three-dimensional ocean circulation on seasonal sea-ice simulations. For this purpose we have constructed an ice-ocean model and used it to carry out a series of seasonal simulations of the Arctic, Greenland, and Norwegian seas. The results show that including the ocean circulation yields first-order improvements in the predicted ice margin location and in the ice velocity fields. Moreover, this improvement in the ice margin prediction requires inclusion of the full three-dimensional circulation of the ocean.

Our basic approach in this study was to couple an existing dynamic thermodynamic sea-ice model (*2, 5*) with a multilevel baroclinic ocean model (*6*). The sea-ice model supplies heat flux, salt flux, and momentum-exchange boundary conditions for the top of the ocean. The ocean model, in turn, supplies current and heat-exchange information to the ice model. Since our main concern here is examining the effect of ocean circulation on sea ice, mean annual observed oceanic temperature and salinity data (*7*) are used to weakly force the ocean model (all terms that are externally specified "force" the model) so that its equilibrium time scale is similar to that of the ice model (3 to 5 years). This "diagnostic" (*8*) method allows the ocean model to be relaxed to available climatological ocean data, while at the same time allowing considerable adjustment in the upper ocean as a result of the effects of ice-ocean interaction. In addition, the barotropic mode of the ocean is fully simulated so that time-varying currents due to surface stress fluctuations are part of the model predictions.

The details of the coupling may be outlined as follows. The sea-ice model of Hibler (*2, 5*) is used to calculate ice drift, thickness, and compactness. We determined the momentum transfer from the ocean to the ice (i) by using the ocean velocity at the second level as the ocean current in the ice calculations (the first ocean level of 30-m thickness is a de facto mixed layer) and (ii) by allowing the surface-pressure term in the rigid-lid ocean model to act on the ice in the same manner as sea-surface "tilt" in conventional ice dynamic calculations. For the momentum transfer into the ocean, the force due to ice interaction is explicitly calculated by the ice model and subtracted from the wind stress. The remainder is taken to be the "wind stress" term transferred into the ocean. An important feature of this procedure is that, in general, the ice will drift in a direction different from the average "Ekman" motion of the upper 30-m layer of the ocean.

With regard to thermodynamic exchanges, heat transferred by the ocean into the upper mixed layer of the ocean is used to either melt ice (until the mixed layer returns to freezing) or warm the mixed layer if no ice is present. The ice model in turn is used to calculate energy exchanges with the atmosphere, based on the use of a complete surface heat budget (2). For salt fluxes, the amount of melting or freezing of ice is used to supply a salt flux into or out of the ocean (9). Because the ice model includes advection, the annual average salt fluxes at given locations will differ substantially from zero.

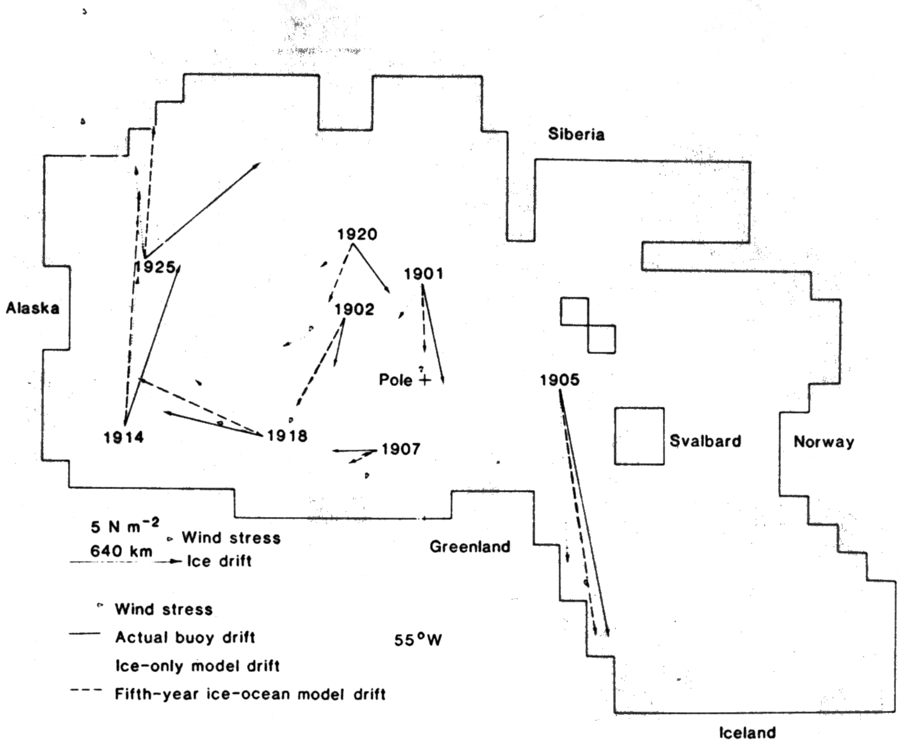

Fig. 1. Long-term average simulated and observed drift rates of ice buoys. To convert the drift rates to distances, we multiplied by time. The time intervals used were February through November for buoys 1901 and 1902, March through September for buoy 1905, May through November for buoy 1907, and March through November for the remaining buoys. Also shown are the mean surface wind stresses at the buoy locations averaged over the same time interval as the drift.

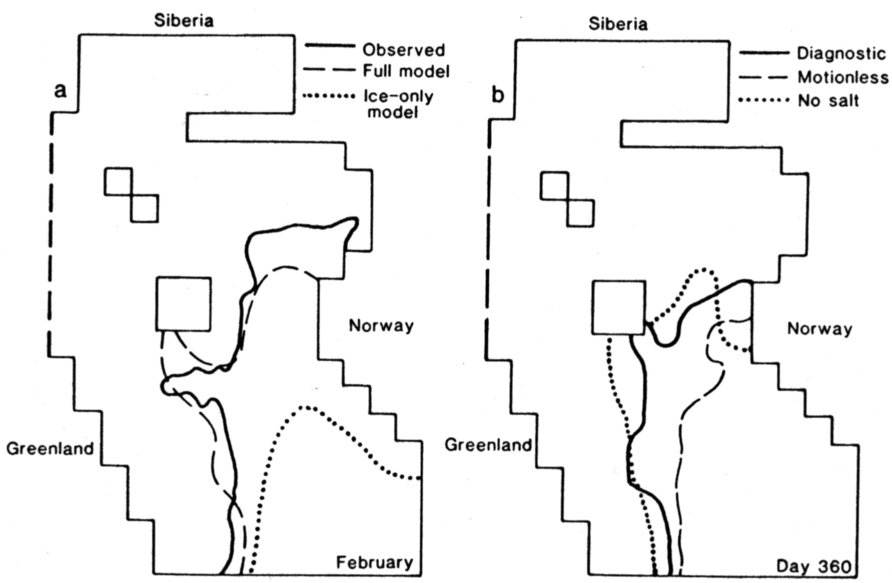

Fig. 2. (a) February 50 percent concentration limits from the full coupled ice-ocean model and for the ice-only model, which includes a fixed-depth mixed layer. The observed limit is taken from fleet weather charts (Suitland, Maryland, U.S. Navy) for the end of February. (b) Simulated 0.05-m thickness contours for day 360 for the full model, a coupled model with a "motionless" ocean, and a coupled ice-ocean model with no salt fluxes due to ice freezing and melting. The "motionless" model and the no-salt simulations were run for only 1 year. The starting points for these simulations were the ice thickness and ocean fields at the end of the fourth year of the full diagnostic model simulation.

Fourteen vertical levels are used in the ocean model, with deeper levels having increasing thickness. Bottom topography is resolved by the use of differing numbers of vertical levels at different locations. The diagnostic forcing (10) to observed data is done at all but the upper level of the ocean with a uniform 3-year relaxation time. In addition, at lateral boundaries of the ocean without land (11), all levels of the ocean are forced with a 30-day relaxation time over several grid cells closest to the boundary. This procedure follows techniques determined by Sarmiento and Bryan (12). To simulate river runoff, river inflow was specified seasonally at various boundary locations; this procedure yielded mean annual river runoff identical to that used by Semtner (13) in a prognostic (or predictive) simulation of the Arctic Ocean without explicit sea ice.

To force the ice-ocean model, daily time-varying atmospheric pressure from the First GARP (Global Atmospheric Research Program) Global Experiment (FGGE) year (December 1978 to November 1979), together with monthly mean climatological temperature and humidity fields, were used (14). Empirical long- and shortwave radiation calculations (2) were used in conjunction with these atmospheric data to drive the thermodynamic portion of the model.

The horizontal resolution was taken to be 1.45° for the ocean model and 160 km for the ice model. The ocean model was formulated in a spherical grid system with the equator of the grid system going through the north geographical pole. The ice model was formulated in a rectangular grid. However, because of this particular spherical projection, there was little difference between the ice model grid and the ocean model grid (specifically, the ice advection terms will yield slight errors in the heat and salt exchanges). In light of the diagnostic forcing of the ocean, these small differences were not felt to be critical.

To obtain seasonally varying equilibrium results, we integrated the coupled model for 5 years, using 1-day time steps. For comparison, a 5-year simulation was also carried out with an ice-only model, which included only a motionless fixed-depth, 30-m mixed layer and no ocean currents. In addition, to examine the role of different processes, 1-year sensitivity simulations without surface salt fluxes and without ocean currents, respectively, were carried out.

Some of the main results from the model simulations are shown in Figs. 1 through 3. Figure 1 compares observed ice buoy (*15*) drift to simulated results both with and without inclusion of the ocean model. Although of considerable magnitude, the ocean currents account for less than 50 percent of the net ice drift. Although not apparent from Fig. 1, more detailed analyses show that both the ice motion and current structure have considerable temporal variability; the ice motion varies everywhere, whereas the main current variability is in shallower regions or near rapid topographic variations.

The main overall effect of the ocean model on the predicted ice drift is to substantially increase the East Greenland ice drift and to create more of a transpolar drift in the Arctic Basin. Basically, inclusion of the complete ocean model consistently improves the prediction of turning angles and especially improves the simulated drift direction and the magnitudes of ice drift in the East Greenland Sea. Specifically, for the drift vectors shown in Fig. 1, the average turning angle error in the full model is 23° as compared to 38° for the ice-only model. The drift magnitudes also show an improvement, with a cumulative error of 45 percent for the full model as compared to 58 percent for the ice-only model. The remaining errors in the simulated drift rates are probably due to the neglect of interannual variability (ice thicknesses from an earlier year can affect the ice interaction in a subsequent year), or more likely to errors in the mean annual wind stress. For example, the mean annual wind stresses differ substantially from the net observed ice drift (Fig. 1).

Probably the most noticeable effect of the modeled ocean circulation is to greatly improve the ice margin simulation (Fig. 2a). This improvement in the location of the ice margin is due to large amounts of oceanic heat that flows from the deeper ocean into the upper mixed layer in this region (Fig. 3). This heat flux occurs primarily in winter, and analysis shows that much of the enhancement is due to deep convection, which brings up warm water and prevents ice formation in early winter. This convection accounts for the absence of ice in the full-model simulation far from the ice margin.

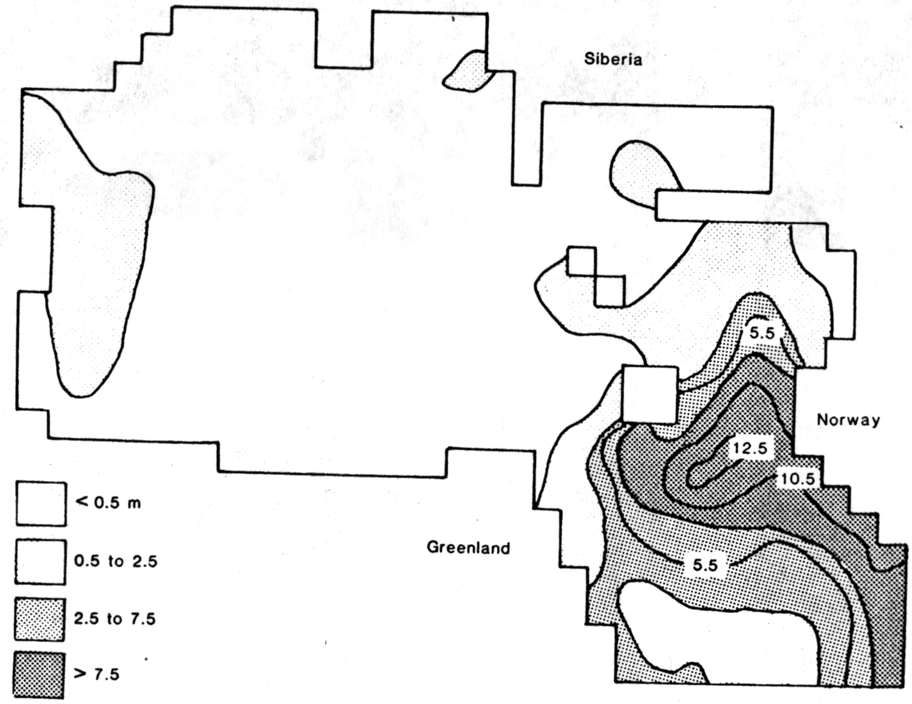

Fig. 3. Average annual heat gained by the upper layer of the ocean from the deeper ocean and by lateral heat transport. The contours are in terms of the melting capacity of the heat (in meters of ice per year). A melt rate of 1 m of ice per year is equivalent to 9.57 W m^{-2}

Near the ice margin where salt fluxes due to ice melting occur, a similar physics applies. But the precise location of the ice edge becomes more sensitive to the surface salt balance and to lateral effects in the oceanic circulation. Sensitivity simulations (Fig. 2b) indicate that the freshwater flux from melting at the advancing ice edge tends to seal off the ice margin to upward oceanic heat flux and allows a further advance than would occur otherwise. Moreover, a sensitivity simulation without lateral motion in the ocean (Fig. 2b) tends to produce a more excessive edge at the end of 1 year as the lateral transport of heat both to the deep and to the shallow layers is missing. This latter result demonstrates the difficulty that arises from the use of a one-dimensional model alone to simulate the seasonal cycle of sea ice.

A particularly notable feature of these results is the very large value of the oceanic heat flux (see Fig. 3) in the Greenland and Norwegian seas and the complex way in which this heat flux varies in space and time. These results underscore the need for the inclusion of a full ocean model in seasonal sea-ice simulations.

These simulations also provide a challenge to prognostic ocean models. The realistic results obtained here are due in part to the diagnostic forcing. To sustain these large heat losses in a fully prognostic ocean model requires very substantial northward heat transport. Whether such transports can be successfully simulated remains to be seen.

The sensitivity of these simulations to salt fluxes from the ice amplifies the need for more detailed examination of the processes dictating the advance and retreat of the ice margin and for the development and verification of more realistic boundary layer formulations (*16*) than used here. Such work is currently under way in the Marginal Ice Zone Experiment (MIZEX) (*17*).

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8. Throughout this report the term "diagnostic" refers to the fact that the mean annual temperature and salinity fields in the ocean are largely specified by observed data rather than predicted by the model. (This latter predictive case will be referred to as prognostic.) Shorter term seasonal variations in temperature and salinity are, however, predicted by the "diagnostic" model.
9. For simplicity and to ensure conservation, we calculated the salt flux by assuming that 100 percent of the salt content (35 per mil) was rejected immediately upon freezing. Similarly, melting was treated as a reduction in salt content of the top layer of the ocean proportional to the ice melt rate multiplied by a salinity of 35 per mil.
10. This forcing is done by adding additional source terms to the heat and salt conservation equations of the form $R(T_0 - T)$ and $R(S_0 - S)$. Here R is a relaxation constant taken to be 0.333 per year, and T_0 and S_0 are, respectively, the mean annual temperature and salinity fields.
11. The main open ocean boundary occurs five grid cells south of Iceland, where a wall has been assumed. Over the four rows of grid cells nearest this wall the diagnostic forcing is made stronger.
12. J. L. Sarmiento and K. Bryan, *J. Geophys. Res.* **87**, 394 (1982).
13. A. J. Semtner, Jr., *J. Phys. Oceanogr.* **6**, 379 (1976).
14. In the selection of forcing fields, the approach was to use climatological data wherever possible. However, it is important to use daily winds to force the ice model in order to simulate the deformation-induced ice growth. Since buoy drift data (*15*) were available for the FGGE year, this year was chosen for forcing winds. Although the mean climatological winds may differ from the FGGE mean winds, these differences will not have major effects on the baroclinic currents, which are largely forced by the diagnostic terms in the ocean model. However, for prognostic simulations it would be important to use wind fields whose annual mean is closer to climatological data.
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18. We thank M. Cox for advice and aid in implementing the ocean code. Figures for this report were prepared from computer printouts by K. Alversen. This work was supported by the National Oceanic and Atmospheric Administration and the Office of Naval Research.

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