Annals of Glaciology 5 1984

MODELING THE OCEAN IN CLIMATE STUDIES

by

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ABSTRACT

A number of processes in the ocean must be modeled properly in order to produce valid estimates of oceanic heat transport, sea-surface temperature, and sea-ice extent in climate studies. These include: wind-driven turbulent mixing and water transport in the surface layer, internal vertical mixing due to several small-scale mechanisms, horizontal and vertical exchanges by mesoscale eddies, mixing along isopycnals, large-scale transport by currents, deep convection in polar regions, and boundary exchanges with atmosphere, ice, and land. Techniques to model these processes are described. Prospects are given for parameterizing the effects of phenomena that cannot be resolved in climate studies, particularly mesoscale eddies. Past simulations of the ocean in climate studies are reviewed. A modeling strategy is outlined for an improved treatment of the ocean, consistent with the computational power soon to be available.

1. INTRODUCTION

The purpose of this paper is to review the methodology and results of modeling the ocean in a climatic context and to suggest future modeling improvements. The paper summarizes a presentation to an audience of glaciologists and climatologists at an international symposium on ice and climate modeling, and it is not intended to be a comprehensive review of ocean climate modeling. Literature references on technical issues have been omitted for the sake of brevity, except in regard to the figures and tables. Those interested in review articles on various oceanic processes are referred to Warren and Wunch (1981).

The author has been involved primarily in modeling the three-dimensional circulation of the oceans, and this review reflects that orientation. It is assumed that most glaciologists and climatologists are interested in modeling climate on time scales as short as the seasonal cycle (because of the importance of summer melting) and as long as the replacement time of the deep world ocean (≈ 800 a). It is also assumed that, although properties of ocean temperature and heat transport are desired mainly in the polar regions, nevertheless the oceanic connections of polar oceans with mid-latitude gyres and the linkage between the Arctic and Antarctic oceans through deep-water circulation will ultimately require ocean modeling on a global scale. The construction of a valid world ocean model is therefore taken as the methodological objective of ocean climate modeling. In the meantime, a number of simpler models will be employed to improve understanding of various issues; but these models will not be discussed here.

A number of important physical processes must be treated properly in a three-dimensional ocean model. The first column of Table I lists these processes in five categories: (a) processes at the upper and lower ocean boundaries, (b) internal vertical mixing, (c) effects of mesoscale ocean eddies, (d) large-scale transports, and (e) miscellaneous high-latitude effects. The next five sections of this paper will discuss these categories in turn from three viewpoints: how the processes are modeled in isolation, how well they are represented today in ocean models, and how they might be treated in improved models of the world ocean. Of the five categories, the one dealing with mesoscale eddy effects will be given the greatest emphasis.

2. UPPER AND LOWER BOUNDARY PROCESSES

It is well known that the ocean is capped by a relatively homogeneous layer which is stirred by wind forcing and by surface heat losses. Process models generally employ fine vertical resolution and governing equations based on higher-order turbulence closures to predict the time-dependent behavior near the surface. Most ocean models for climatic studies employ a constant-thickness mixed layer as a simplification. Variations of upper ocean heat capacity which affect surface temperature are neglected. An improvement is to embed a variable-thickness mixed layer in circulation models. A bulk formulation can be based on simple principles of mechanical energy conservation. The detailed vertical structure below the mixed layer need not be retained if an approximate rule to give the temperature jump at the base of the layer is substituted. Figure 1 shows a test by the author of such a mixed-layer model, driven by observed atmospheric forcing over a two-year time period. There is reasonable agreement between predicted and observed temperatures on all time scales. Many of the short-term temperature fluctuations result from entrainment of colder water during episodes of mixed-layer deepening. Even the amplitude of the seasonal cycle is directly related to summertime average depth of the mixed layer. These fluctuations cannot be predicted by a constant-thickness mixed layer. It therefore seems that prognostic variable thickness should be included in ocean climate models. It is also physically reasonable to regard momentum as well-mixed to this depth, and the Ekman transport perpendicular to the direction of wind stress should be uniformly distributed through the mixed layer.

The bottom boundary layer is often neglected in ocean climate models. However, studies with resolved ocean eddies indicate that bottom friction can be the dominant mechanism of energy dissipation in some circumstances. Observations and process models suggest that bottom friction can be adequately represented in large-scale models by a quadratic drag law.

https://doi.org/10.3189/1984AoG5-1-133-140 Published online by Cambridge University Press

TABLE I. PHYSICAL PHENOMENA OF OCEANS AND METHODS OF MODELING THEM

	Physical phenomenon	Process model approach	Usual climatic model approach	Improved climatic approach
	Surface oceanic mixing	higher-order closure	constant-thickness mixed layer	prognostic-thickness bulk mixed layer (Heald and Kim 1979)
(a)	Surface ageostrophic transport	vertically resolved	Ekman transport in upper level	Ekman transport in bulk mixed layer
	Bottom boundary layer	vertically resolved	free-slip	quadratic drag law (Weatherly 1972)
	Vertical shearing instability	vertically resolved		
(b)	Double-diffusive effects	vertically resolved	K_H = constant	$K_H = fn(Ri,N)$
	Small-scale vertical mixing	vertically resolved		(Pacanowski and Philander 1981, Sarmiento and others 1976)
	Barotropic instability		$A_M = 10^6\ m^2\ s^{-1}$ $A_H = 2 \times 10^3\ m^2\ s^{-1}$	$A_M = A_H = 0(10^3\ m^2\ s^{-1})$
(c)	Baroclinic instability	horizontally resolved with Δx = 20 km		$-A_M = A_H = 0(10^3\ m^2\ s^{-1})$
	Vertical form drag by eddies		K_M = constant	$K_M = fn(\Delta T,\Delta u)$
	Large-scale advection	done by time-mean currents	poorly resolved by 500 km, 5-level grid	moderately resolved by 100 km, 10-level grid
(d)	Mixing along isopycnals	isentropic coordinates	z coordinates	diffusion along isopycnals
	Interaction with topography	sigma coordinates	flat bottom	variable number of levels (Bryan 1969)
	Density relation	Knudsen formula	linear/quadratic formula	higher order polynomial (Bryan and Cox 1972)
(e)	Gravitational instability	plume models	convective adjustment	modified convective adjustment
	Sea-ice dynamics	viscous/elastic/plastic	no motion	empirical ice motion (Thorndike and Colony 1982)
	Ice thermodynamics	storage of sensible and latent heat	no heat storage	storage of sensible and latent heat (Semtner 1976)

3. INTERNAL VERTICAL MIXING

There are a number of physical processes which mix heat and salt vertically. One of these is an instability caused by vertical shear combined with weak stratification. Others are related to the different molecular diffusivities of heat and salt and to breaking internal waves. All of these phenomena have usually been modeled in climate studies by a constant vertical eddy diffusivity. A number of investigators have suggested functional forms for the dependence of the diffusivity on Richardson number Ri and Vaisala frequency N. Such formulations should be adopted in future ocean climate studies.

4. EFFECTS OF MESOSCALE OCEAN EDDIES

In multilevel models of the ocean, it is necessary to include horizontal diffusion of momentum, heat, and salt to account for the effects of unresolved scales of motion. Failure to prescribe adequately large diffusion coefficients causes unrealistic checkerboard patterns to appear in the solution fields. The larger the grid spacing is in a model, the larger are the required coefficients. A

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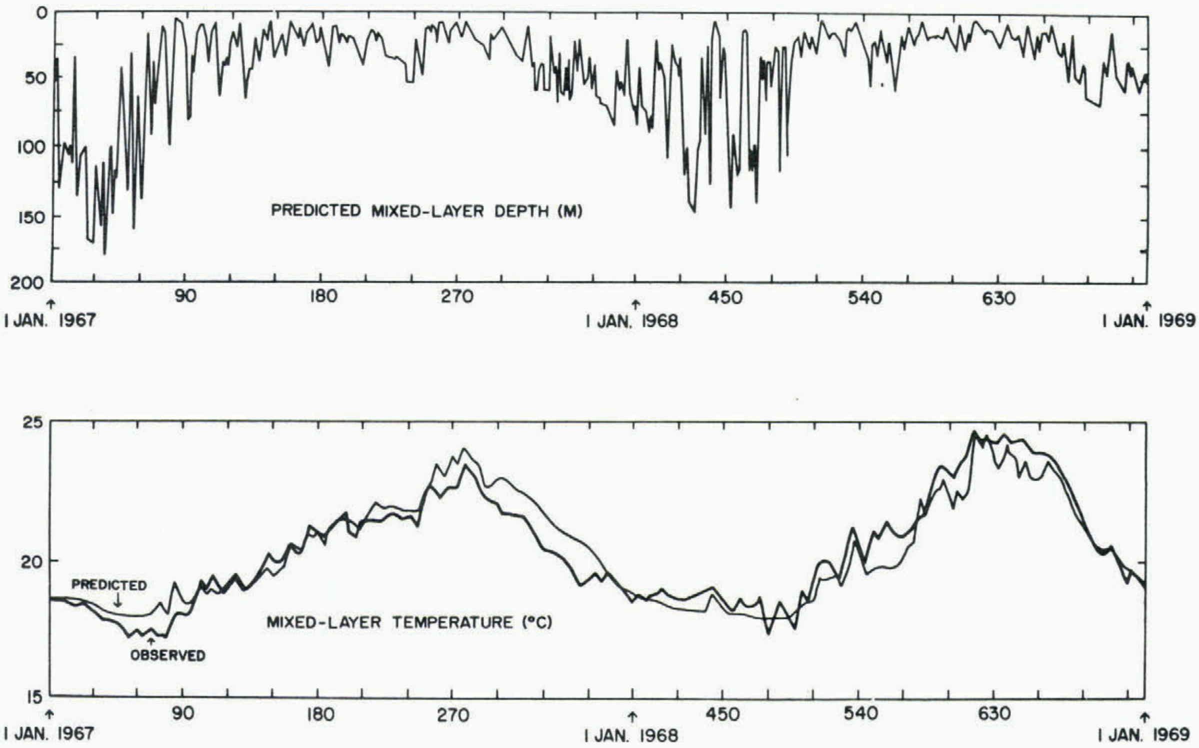


Fig.1. Performance of an embedded mixed-layer model (developed by Heald and Kim 1979) using observed atmospheric forcing from an ocean weather ship. (From Semtner in press.)

number of numerical studies of the world ocean have been carried out with grid spacings of 200 to 500 km. The grid spacing is generally dictated by computational considerations: cutting the horizontal grid size in half increases the number of calculations by a factor of eight or more. Table II indicates the diffusion coefficients that have been used in several studies. Over the range of grid spacing, the momentum coefficient varies from 10^5 to 10^6 m^2 s^{-1} and the the coefficient for heat and salt varies from 10^3 to 2×10^4 m^2 s^{-1}. Usually, minimal values in a particular study are found by trial and error rather than being dictated by physical considerations. For each of the studies listed in Table II, the values were constant in space and time.

Over the past decade, increased understanding of the physical basis for horizontal redistributions of momentum, heat, and salt in the ocean by mesoscale eddies has been obtained by observational studies, as well as by numerical studies with horizontal grid spacing adequate to resolve the eddies. The redistributions are complex, in that the order of magnitude and even the sign of the effective diffusion coefficient can change on relatively small spatial scales. Oceanographers who are interested primarily in mesoscale eddies have used this fact to argue that modeling studies with constant diffusion coefficients and without resolved eddies are invalid.

Eddy-resolving models have shown that, in certain types of oceanic flows, instabilities related either to the horizontal structure of the currents or to the vertical shear of the currents lead to downgradient diffusions of heat, with an order of magnitude of 10^3 m^2 s^{-1}. Observational studies in oceanic currents give similar results (see the last column of Table III). The order of magnitude for eddy diffusion of momentum is also 10^3 m^2 s^{-1} (cf. A_M values in Table III), but the sign can be negative in some cases. Model eddies have also been found to drive deep circulations by processes such as form drag, so that the effective vertical momentum diffusion can be as large as 1 m^2 s^{-1}.

The specified values of A_H in the global ocean studies of Table II are of magnitude 10^3 m^2 s^{-1} (except for one), whereas the specified values for A_M are much larger. This suggests that coarse-grid models may distort the predicted motion field more than the thermal field. Putting both coefficients at order 10^3 m^2 s^{-1} would require a grid size of 100 km or less, while still not resolving eddies. At this point, it seems reasonable first to examine global simulations made with the coarse grids and

TABLE II. HEAT AND MOMENTUM DIFFUSION COEFFICIENTS A_H AND A_M (m^2 s^{-1}), GRID SPACING Δx (km), AND CIRCUMPOLAR TRANSPORT $\Delta\psi$ (10^6 m^3 s^{-1}) FROM WORLD OCEAN SIMULATIONS

	Cox (1975)	Takano (1975)	Bryan and others (1975)	Washington and others (1980)	Meehl and others (1982)	Han (in press)
A_H	10^3	2.5×10^3	2.5×10^3	2×10^4	2×10^3	2×10^3
A_M	0.8×10^5	10^5	0.8×10^6	10^6	10^6	0.8×10^6
Δx	200	400	500	500	500	400
$\Delta\psi$	186	35	22	50	105	86

see how well or badly they reproduce large-scale features of the observed ocean circulation and second to examine the extent to which a grid-size reduction may improve the representation of eddy effects.

Figure 2 shows the differences between computed and observed sea surface temperature (SST) from two coupled ocean-atmosphere models. One might assume that the large positive errors in SST at high latitudes result from a large grid size. However, the main source of error in the Geophysical Fluid Dynamics Laboratory (GFDL) simulation is misrepresentation of southern-hemisphere winds by the atmospheric model, while in the National Center for Atmospheric Research (NCAR) simulation it is an overly large value of A_H.

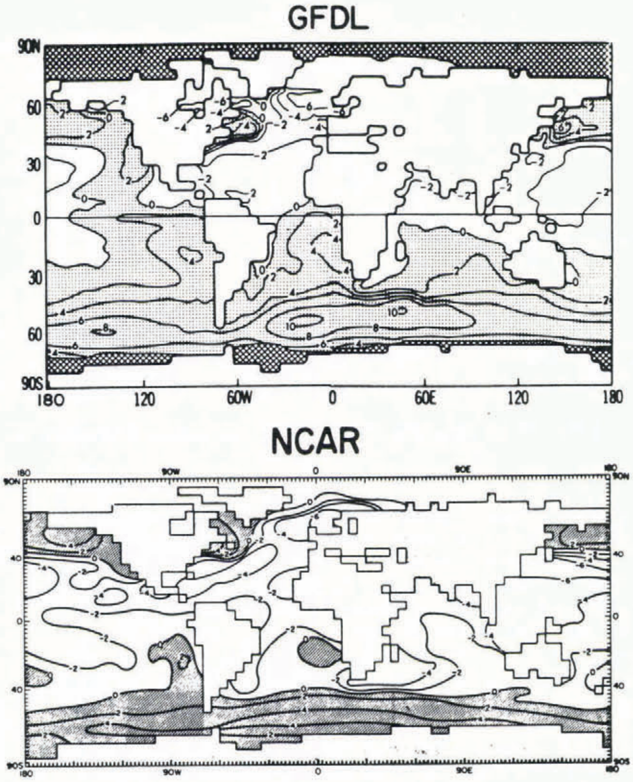
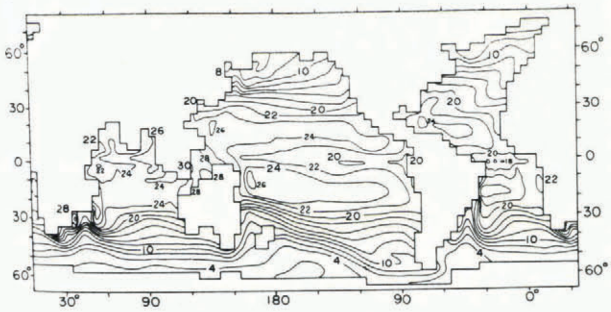


Fig.2. Annual average SST differences (computed minus observed) from coupled model studies at GFDL and at NCAR. (From Washington and others 1980.)

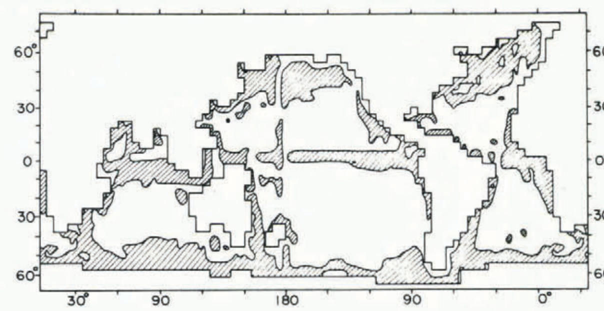
A fairer assessment of ocean model performance can be made by looking at cases with observed atmospheric forcing and A_H of order 10^3 m^2 s^{-1}. Figure 3 shows fields from a global model driven by observed winds but by surface heat fluxes computed from a zonally invariant atmospheric temperature. The SST in the top panel would consist of horizontal lines in the absence of ocean circulation. Obviously, circulation is important, and many features of the predicted SST qualitatively resemble those observed. The middle panel shows a rather realistic pattern of equatorial and polar upwelling, together with mid-latitude downwelling. The bottom panel shows significant surface heat fluxes, which are required to balance oceanic heat transport.

Figure 4 shows simulated heat transport and heat storage from an NCAR model with observed forcing and reduced A_H, versus the observed quantities. Except for a diffusive contribution which is still too large in high latitudes, the meridional heat transport is well modeled in terms of both the annual mean and the seasonal cycle. Seasonal heat storage and meridional heat transport by individual oceans are fairly well modeled. Since heat transport and

Temperature distribution at a depth of 20 m.



Vertical component of velocity at a depth of 70 m.



Heat flux (ly d^{-1}) computed from the surface water temperature.

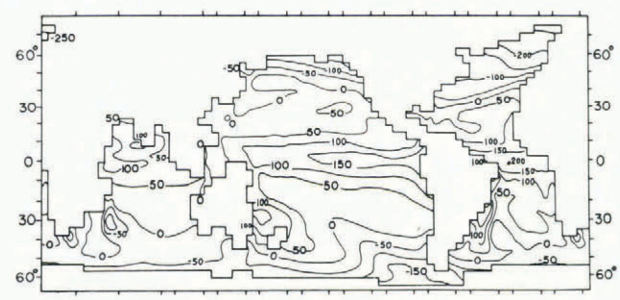


Fig.3. Predicted fields from the world ocean model of Takano (1975).

heat storage are the major components of the oceanic influence on SST and climate, the ability of the coarse grid models to simulate the ocean properly in climate studies may be better than eddy modelers have suggested. A fairly realistic simulation of SST from another coarse-grid model has been obtained by Han (in press).

The proper representations of heat transport and storage in the coarse grids may be due to the dominant effects of meridional overturning and mixed-layer heat capacity, respectively. As indicated earlier, the momentum diffusion coefficient A_M can be expected to misrepresent horizontal circulation in coarse grids. Table II gives values of the mass transport by the Antarctic Circumpolar Current through the Drake Passage. It is clear that all the simulations give low values relative to the observed value of (185 ± 35) x 10^6 m^3 s^{-1}, except for one study with a finer grid and significantly smaller A_M than the others. However, models with large A_M do show similar patterns of horizontal circulation to those of the finer grid and of the real ocean (Fig.5). The study with the fine grid still underestimates the observed strength of mid-latitude gyres (≈ 100 x 10^6 m^3 s^{-1}), although the meridional transport is better represented (Fig.6).

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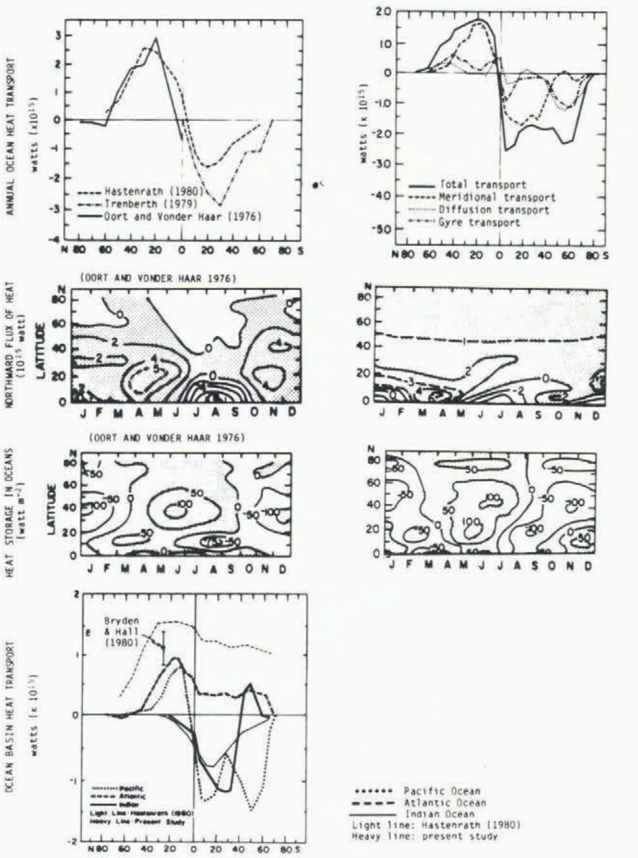

Fig.4. Aspects of oceanic heat transport and heat storage, as observed (left) and as predicted (right) in a model driven by atmospheric data. The bottom panel shows the predicted and observed transports in individual ocean basins. (From Meehl and others 1982.)

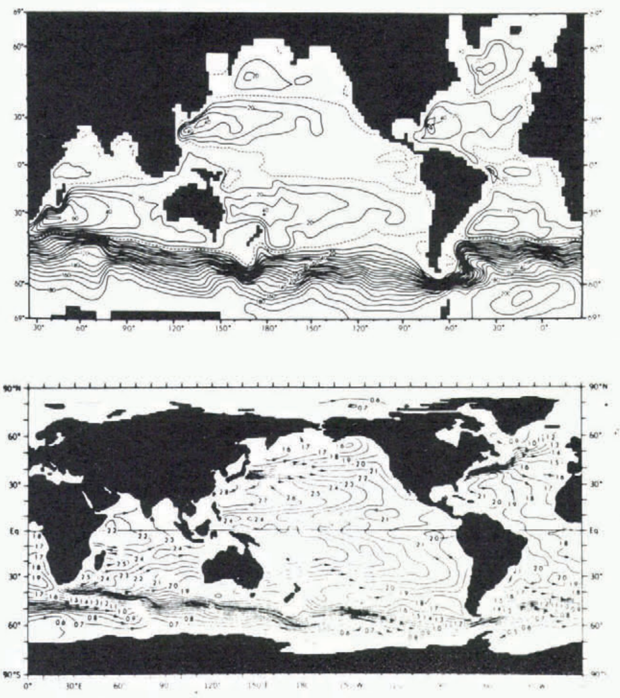

Fig.5. Above: streamlines of mass transport from the world ocean study of Cox (1975). Below: streamlines of upper ocean currents inferred from the density field by Levitus (1982).

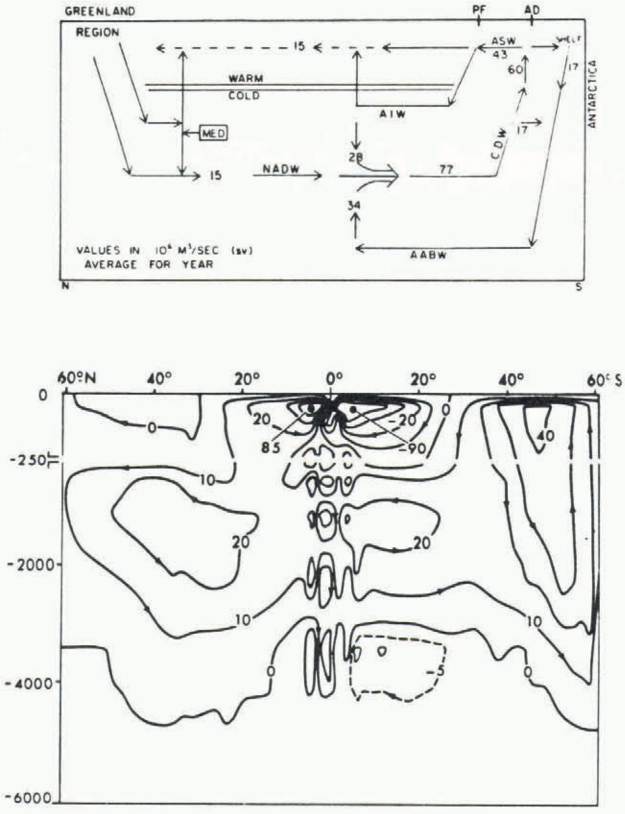

Fig.6. (above) Schematic view of the observed meridional circulation according to Gordon (1971) and (below) stream function of the meridional circulation in the model of Cox (1975).

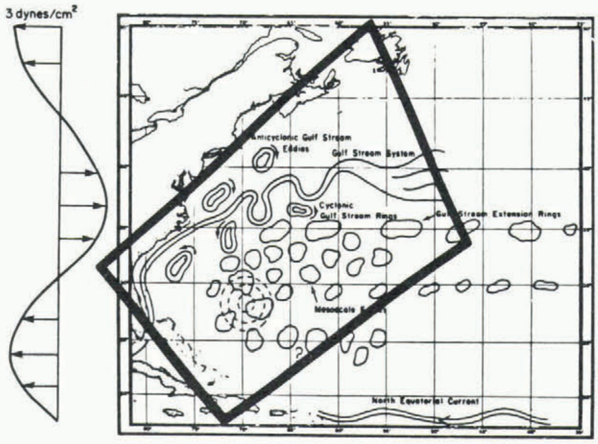

Fig.7. Schematic representation of phenomena of the Gulf Stream, as well as the model domain (bounded by the heavy line) and the wind stress used in the eddy resolving study of Semtner and Mintz (1977).

Let us now turn to the question of whether further reductions in grid size and in A_M without resolving eddies will adequately portray the horizontal circulation. Figure 7 shows the schematic design of an eddy-resolving study of the Gulf Stream. Figure 8 shows simulated fields of surface streamlines (height) and surface temperatures as follows: left, instantaneous fields from a 37 km grid; middle, time-averaged fields from a 37 km grid; and right: steady fields from a 75 km grid using $A_M = A_H = 10^3$ m^2 s^{-1}. From the point of view of large-scale influence on climate, the middle and left fields do not differ

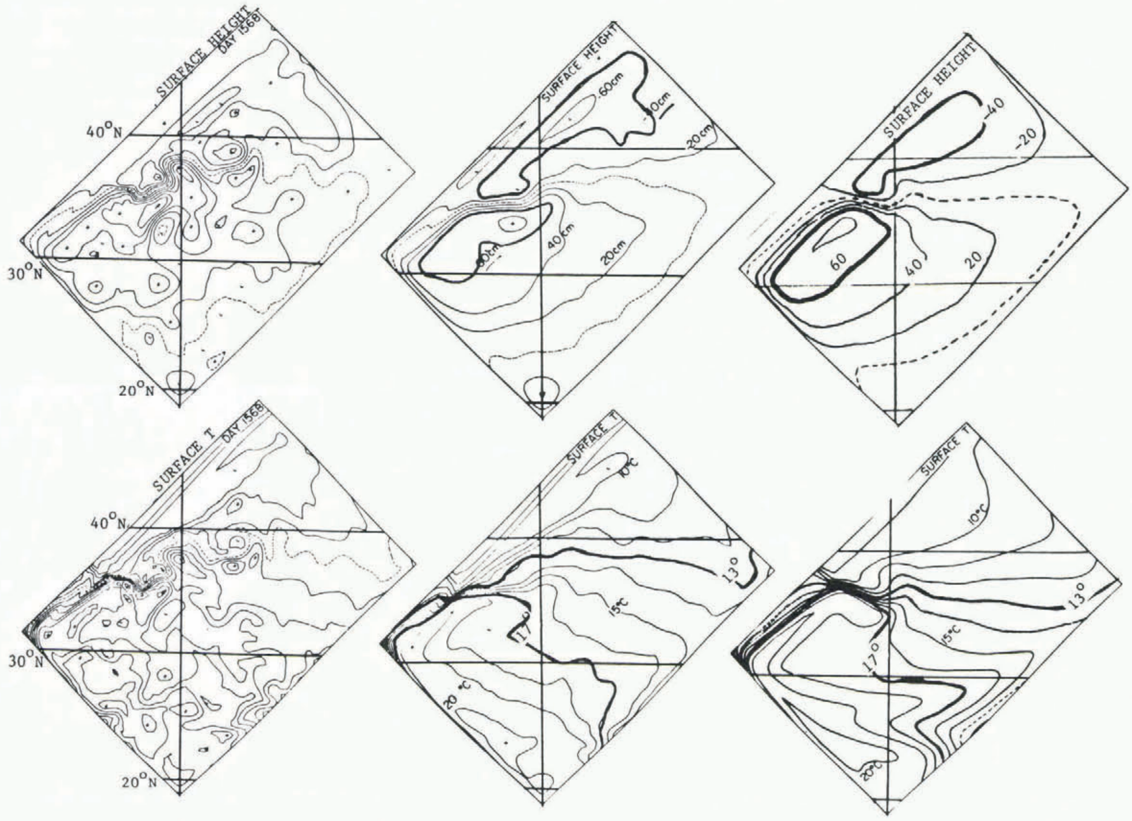

Fig.8. Left: maps of instantaneous surface height and surface temperature in the eddy resolving simulation of Semtner and Mintz (1977); center: time-averaged fields from that simulation; right: the steady fields from a coarse-grid simulation using $A_H = A_M = 10^3\ m^2\ s^{-1}$.

TABLE III. SOME OCEANIC HORIZONTAL DIFFUSIVITIES FROM OBSERVATIONAL (O) AND MODEL (M) STUDIES

Current system	Investigation	$A_M(m^2\ s^{-1})$	$A_H(m^2\ s^{-1})$
Gulf Stream and recirculation	(O) Bryden (1982)	$-9x10^3$	$11x10^3$
	(M) Semtner and Mintz (1977)*	---	$3-10x10^3$
westward interior flow	(O) Freeland and others (1975)	---	$0.8x10^3$
	(M) Semtner and Mintz (1977)*	---	$1-3x10^3$
equatorial currents	(M) Semtner and Holland (1980)	$2x10^3$	$2-10x10^3$
Circumpolar Current	(O) Bryden (1979)	---	$0.6x10^3$
	(M) McWilliams and others (1978)	$-3x10^3$	$0.9x10^3$

* based on unpublished further analysis.

much from each other. (In the absence of circulation, the surface height would be constant and the temperature would vary linearly with latitude from 30 to 0°C.)

Figure 9 shows instantaneous fields in another eddy-resolving experiment, this time for an equatorial simulation. Figure 10 compares several time-averaged fields versus the quasi-steady ones obtained by setting $A_M = A_H = 2 \times 10^3\ m^2\ s^{-1}$. Once again the comparison is favorable from a large-scale viewpoint. Such results suggest that much of the time-averaged oceanic circulation can be obtained from models with sufficiently fine grid to allow an A_M of order 10^3. Additional refinements, including local sign changes of A_M and enlarged value of vertical viscosity K_M can be included on the basis of developing theoretical understanding.

5. LARGE-SCALE TRANSPORTS

A related issue concerning grid size is whether oceanic time-mean fields can be adequately depicted. Figure 5 gives some indication that a 200 km grid size adequately represents the structure of the Antarctic Circumpolar Current (after taking account of the Mercator stretching of the model field). Figure 8 indicates that 75 km grid spacing allows western boundary features to be depicted. The scale of the simulated Gulf Stream jet is about 150 km, thereby indicating an upper limit for resolution. Thus the 100 km grid size required for realistic eddy viscosity is adequate for depicting the time-mean fields. Above 150 km, some degradation in certain regions can be expected.

In the real ocean, eddy diffusion may act more along isopycnal surfaces than purely in the hori-

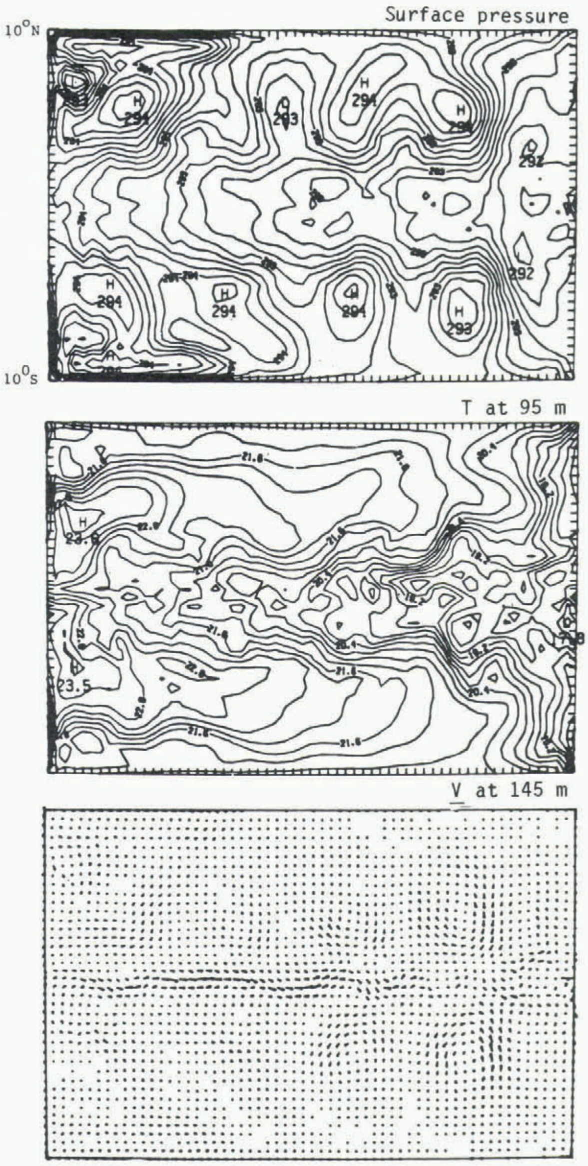

Fig.9. Instantaneous fields from the eddy-resolving equatorial experiment of Semtner and Holland (1980).

zontal. The choice of isentropic coordinates gives a better treatment of such diffusion but involves formidable numerical problems. For improved ocean climate modeling, continued use of z coordinates for dynamics but adoption of diffusion operators oriented along isopycnals may be the best compromise.

Topography is handled in atmospheric models by the use of sigma coordinates, whereby the lowest coordinate surface follows the terrain. The method tends to break down for the oceans, because the scale of the topography relative to ocean depth is too large. In some ocean climate models, a flat bottom is assumed, but this can distort currents, thermohaline circulation, and water-mass formation. The method of using z coordinates but allowing a variable number of vertical levels is probably preferable.

6. MISCELLANEOUS HIGH-LATITUDE EFFECTS

Brief mention should be made that accurate representation of the nonlinear equation of state for sea-water is important, especially in polar regions. Also, the usual method of dealing with gravitational instability by convective adjustment may treat bottom-water

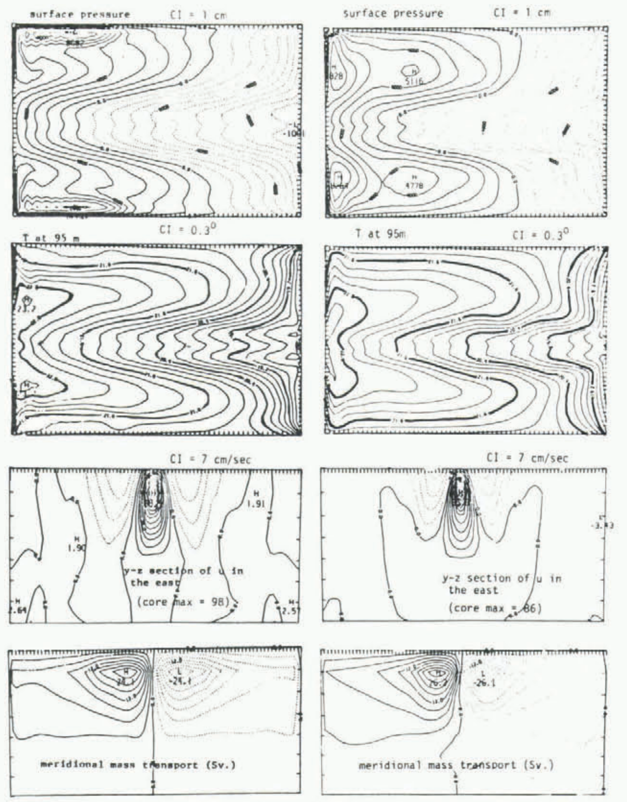

Fig.10. Left: Time-averaged fields from the eddy resolving experiment of Semtner and Holland (1980). Right: instantaneous fields obtained from introducing lateral diffusion of heat and momentum (using $A_H = A_M = 2 \times 10^3$ m^2 s^{-1}) and running the experiment an additional 150 days.

formation inadequately and require augumentation by plume models in some key regions. Finally, the thermodynamics and dynamics of sea ice, which are now treated rather poorly in most climate models, need to be improved.

7. CONCLUSIONS

To provide more accurate oceanic simulations in climate studies, a number of improved parameterizations have been suggested. Many of these are straightforward and cause little increase in computation. The largest increase would come from suggested grid-size reductions to allow more realistic eddy viscosity and smaller horizontal scales in the predicted fields. The need for resolving eddies in climate studies, as suggested by eddy modelers, has been disputed. The suggested grid-size reductions without resolving eddies will be feasible as faster computers become available in the mid to late 1980s. In the meantime, simulations with coarse grids can meaningfully portray many aspects of ocean circulation.

ACKNOWLEDGEMENT

The National Center for Atmospheric Research is sponsored by the US National Science Foundation.

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