A HIGH RESOLUTION ICE-OCEAN MODEL
WITH IMBEDDED MIXED LAYER

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JINLUN ZHANG

Thayer School of Engineering
Dartmouth College
Hanover, New Hampshire
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A numerical investigation of the North Polar ice-ocean system is carried out in two stages. In the first stage, a high-resolution ice-ocean model for the Arctic Ocean, the Barents Sea and the Geenland-Iceland-Norwegian (GIN) Sea is developed. This model has a horizontal resolution of 40 km x 40 km and 21 vertical levels. It is one of the highest resolution models yet applied to the seas of the North Polar Region and can easily resolve the coastline, bottom topography, narrow passages and currents in the region. In addition, ocean open boundaries are used in the model for the Bering Strait, Denmark Strait and the Faroe-Shetland Passage to obtain more realistic ocean inflow/outflow conditions. Also introduced into the model is observed monthly varying precipitation as a supply of fresh water or addition of ice to maintain salt balance. This model is then used to simulate the ice and ocean circulation in the Arctic Ocean, Barents Sea and GIN Sea.

In the second stage, an oceanic mixed layer model of Kraus and Turner (1967) is embedded into the ice-ocean model to examine the effects of vertical mixing processes on the sea ice distribution and ocean circulation in the North Polar Region. A method, mainly based on Adamec et al (1981) and Resnyanskiy and Zelen'ko (1991) with some modifications, is used to couple the mixed layer model with the ice-ocean model. A one-dimensional salinity mixed layer model was created and tested to ensure a success of the
coupling efforts. The behavior of the coupled ice-ocean-mixed-layer model is examined and compared to the previous ice-ocean-only model through a parallel simulation of the ice-ocean system in the North Polar Region.

The simulation results show a two-gyre ocean circulation pattern in the Arctic Basin. One is the well documented Beaufort Sea anti-clockwise gyre, the other is the less known clockwise gyre in the Eurasian Basin. This two-gyre system is found to be forced by the bottom topography resolved by the high resolution model. The results also show the necessity of using a high resolution grid to enhance the simulation of the Atlantic inflow to the Arctic in the intermediate levels, which is predicted by the models to be around 1 Sv (10^6 m^3/s). In addition to simulating the major currents in the region under consideration, the models also resolve the narrow Beaufort Sea undercurrent and create a realistic flow reversal there by the help of the inflow from the Bering Strait open boundary and the ice cover's influence on the ocean surface stress input. The incorporation of the open boundaries into the models is also partially responsible for the one Sv Atlantic water inflow at the Fram Strait, a basically balanced heat budget in the whole region under consideration, and improved ice drift statistics.

The intensive vertical mixing processes at the ice edge generated by the mixed layer model considerably improve the prediction of the Greenland Sea ice edge location and its interannual variability compared to observations. The existence of the variable depth mixed layer is found to be of less impact to the Arctic Basin and the Barents Sea in terms of the variations of ice extents. However, the Arctic ice thickness simulated by the fully coupled ice-ocean-mixed-layer model is increased in closer agreement with observations. The mixed layer model only slightly changes the ocean surface circulation in the Arctic, but does not change the overall circulation pattern there.
The interannual variations of the ice concentrations in the three regions, the Arctic Ocean, Barents Sea and GIN sea, are closely correlated to the variations of the surface air temperature. The effects of ocean circulation are most prominent in the Barents Sea ice extent in terms of interannual variability. An ice outflow at the Fram Strait is predicted which is close to observational estimates. This ice outflow is important to the ice conditions in GIN Sea where the melted ice is basically compensated by the ice outflow. The interannual variations of the ice outflow and the mean distributed ice thickness in the GIN Sea are closely correlated statistically.
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Contents

List of Figures ........................................................................ vii
List of Tables .......................................................................... xii
Chapter 1. Introduction .............................................................. 1
1.1 Previous Studies ................................................................... 2
1.2 Thesis Work ....................................................................... 8

Chapter 2. Description of the North Polar Region ......................... 11
2.1 Physical Features ............................................................... 11
2.2 Ice Characteristics ............................................................. 13
2.3 Ocean Features .................................................................. 16

Chapter 3. Model Description ...................................................... 32
3.1 Ice Model ........................................................................ 33
3.2 Ocean Circulation Model ................................................... 36
3.3 Mixed Layer Formulation .................................................. 42
3.4 Coupling of Mixed Layer Model with Ocean Model ............... 49
3.5 Coupling of Ice, Ocean and Mixed Layer Models ................. 55
3.6 Numerical Framework ....................................................... 57
3.7 Atmospheric Forcing Fields ............................................... 60
3.8 River Runoff and Precipitation ......................................... 61
3.9 Boundary and Initial Conditions ....................................... 62

Chapter 4. Experiment with Ice-Ocean-Only Model ....................... 70
4.1 Ocean Surface Velocity Fields ........................................... 70
4.2 Water Transport Patterns.................................................................72
4.3 Trace of Atlantic Water in the Arctic.............................................75
4.4 Beaufort Sea Undercurrent...........................................................75
4.5 Analysis of Integrated Vorticity......................................................79
4.6 Ocean Temperature and Salinity Fields........................................81
4.7 Ocean Heat Budget.........................................................................82
4.8 Oceanic Heat Flux..........................................................................85
4.9 Ice Compactness and Thickness Fields.........................................88
4.10 Ice Drift and Outflow..................................................................89

Chapter 5. Experiment with the Ice-Ocean-Mixed-Layer Model........141
5.1 Characteristics of Variable Depth Mixed Layer..............................142
5.2 Ice Compactness and Thickness......................................................151
5.3 Ocean Circulation..........................................................................153

Chapter 6. Interannual Characteristics of the Models......................193
6.1 Ice Concentration Variations.........................................................193
6.2 Ice Budget in GIN Sea.................................................................197

Chapter 7. Summary and Conclusions .............................................205
7.1 Summary and Conclusions.............................................................205
7.2 Suggestions for Future Work.........................................................211

Appendix 1. Numerical Treatment of Open Boundary Conditions........213
Appendix 2. Test of One-Dimensional Salinity Mixed Layer Model......215

References..........................................................................................222
List of Figures

Figure (2.1.1) Map of the North Polar Region identifying major geographical features .................................................. 22

Figure (2.1.2) Contour map of the bathymetry of the North Polar Oceans .................................................. 23

Figure (2.2.1) The average and extreme seasonal limits of the Arctic sea ice extent for ice concentrations ≥ 1/8 .................................................. 24

Figure (2.2.2) Extreme sea ice conditions at the end of February and August from 1966 to 1975 .................................................. 25

Figure (2.2.3) Record of ice conditions off the North Slope of Alaska .................................................. 26

Figure (2.2.4) The pattern of mean ice drift and surface water flow in the Arctic Ocean .................................................. 26

Figure (2.2.5) Recorded tracklines of buoys deployed in the Arctic in 1981 and 1982 .................................................. 27

Figure (2.3.1) Vertical profiles of temperature and salinity at selected locations in the North Polar Region .................................................. 28

Figure (2.3.2) Summer surface (5 m) salinity of the North Polar Oceans .................................................. 29

Figure (2.3.3) Pattern of surface circulation in the North Polar Oceans .................................................. 30

Figure (2.3.4) Pattern of the Atlantic water circulation in the North Polar Oceans .................................................. 31

Figure (3.4.1) Illustration of vertical mixing .................................................. 51

Figure (3.4.2) Illustration of mixed layer shallowing .................................................. 52

Figure (3.6.1) Grid configuration of the 40 km resolution ice-ocean model .................................................. 67

Figure (3.6.2) Contour map of bottom topography represented in the 40 km resolution ice-ocean model .................................................. 68

Figure (3.6.3) Bottom topography represented by numbers of ocean levels in vertical columns of T and S points .................................................. 69

Figure (4.1.1) Mean ocean surface velocity .................................................. 93

Figure (4.1.2) Annual (1983) mean pressure contours at ocean levels 2 (18 m)
and 3 (37 m) .................................................................96

Figure (4.2.1) Monthly mean streamfunction contours ..................................97

Figure (4.2.2) Annual (1983) mean velocity and temperature distributions at
the transect across the Fram Strait .....................................................98

Figure (4.2.3) Annual mean Atlantic water into the Arctic through the Fram Strait ...... 99

Figure (4.2.4) Contours of potential vorticity, f/H .........................................100

Figure (4.3.1) Annual (1983) mean velocity and temperature distributions at
transect J63 .................................................................101

Figure (4.3.2) Annual (1983) mean temperature distributions across transect J72
and J80 ............................................................................102

Figure (4.3.3) Annual (1983) mean ocean velocity at a depth of 407 meter ..........103

Figure (4.3.4) Annual (1983) mean pressure fields at 300 m and 407 m deep ..........104

Figure (4.4.1) Monthly mean velocity distributions across transect J63 off the
Alaska coast ........................................................................105

Figure (4.4.2) Annual (1983) mean ocean surface velocity ............................106

Figure (4.4.3) Annual (1983) mean streamfunction contours .........................109

Figure (4.4.4) Daily ocean surface stress and water transport at grid cell (14,69) ......110

Figure (4.4.5) Comparison of monthly mean Atlantic water into the Arctic through
the Fram Strait in 1979 among Ice-Ocean-Only, No-Fi and No-OB ..........112

Figure (4.5.1) Monthly (June, 1979) mean vertically integrated vorticity
components at transect J69 over the Arctic Basin ..................................113

Figure (4.6.1) Simulated 7-year (1979-85) mean vertical profiles of temperature
and salinity at selected locations in the North Polar Region .....................115

Figure (4.6.2) Simulated 7-year (1979-85) mean surface (5 m) salinity fields in
February and September ................................................................116

Figure (4.6.3) Temperature and Salinity distributions at Transect J10 .............117

Figure (4.6.4) Temperature and Salinity distributions at Transect J32 .............118
Figure (4.6.5) Salinity distributions at Transects J63 and J80 ........................................119
Figure (4.7.1) 7-year mean monthly heat transport (unit: $10^{12}$ w) in different areas
and the whole region ...........................................................................................................120
Figure (4.7.2) Annual heat transport in the whole region .................................................121
Figure (4.7.3) Annual heat transport in the Arctic Basin ....................................................122
Figure (4.7.4) Annual heat transport in the Barents Sea ..................................................123
Figure (4.7.5) Annual heat transport in the Greenland Sea .................................................124
Figure (4.8.1) Monthly mean heat flux components (w/m²) at grid cell (25,54)
in the Beaufort Sea ............................................................................................................125
Figure (4.8.2) Monthly mean heat flux components (w/m²) at grid cell (94,63)
in the Barents Sea .............................................................................................................126
Figure (4.8.3) Monthly mean heat flux components (w/m²) at grid cell (88,35)
in the GIN Sea ..................................................................................................................127
Figure (4.8.4) Monthly mean heat flux components (w/m²) at grid cell (96,35)
in the GIN Sea ..................................................................................................................128
Figure (4.8.5) Monthly mean heat flux components (w/m²) at grid cell (105,35)
in the GIN Sea ..................................................................................................................129
Figure (4.9.1) 7-year mean ice compactness fields .................................................................130
Figure (4.9.2) 7-year mean monthly ice concentrations .........................................................131
Figure (4.9.3) 7-year mean ice thickness distributions ............................................................132
Figure (4.10.1) Mean ice velocity distributions ....................................................................133
Figure (4.10.2) Comparison between simulated ice drift and observed buoy drift ............136
Figure (4.10.3) Annual mean areal ice transport out of the Fram Strait from the Arctic ....138
Figure (4.10.4) Annual mean ice volume transport out of the Fram Strait and
the Denmark Strait .............................................................................................................139
Figure (5.1.1) 7-year mean monthly mixed layer depth fields .............................................154
Figure (5.1.2) Transect (J37) ~ time plot of 7-year mean monthly
mixed layer depth .................................................................155

Figure (5.1.3) Transect (J37) ~ time plot of monthly mean mixed layer depth
for 1981 through 1983 ...............................................................156

Figure (5.1.4) Transect (J37) ~ time plot of 7-year mean oceanic heat flux ............159

Figure (5.1.5) Transect (J37) ~ time plot of 7-year mean ice edge in the
Greenland Sea ...........................................................................161

Figure (5.1.6) 7-year mean ice edge distributions in the Barents and GIN Seas .......162

Figure (5.1.7) (a) Transect (J37) ~ time plot of 7-year mean friction velocity
and 1981 entrainment velocity ..................................................164

Figure (5.1.8) Transect (J37) ~ time plot of 7-year mean surface salt flux ...............166

Figure (5.1.9) Transect (J37) ~ time plot of 7-year mean surface heat flux ..........168

Figure (5.1.10) Transect (J37) ~ time plot of 7-year mean surface (5 m) salinity ........170

Figure (5.1.11) Transect (J37) ~ time plot of 7-year mean surface (5 m) temperature ...172

Figure (5.1.12) Transect (J63) ~ time plot of 7-year mean monthly mixed
layer depth ...............................................................................174

Figure (5.1.13) Transect (J63) ~ time plot of 7-year mean friction velocity ..........175

Figure (5.1.14) Transect (J63) ~ time plot of 7-year mean oceanic heat flux ..........176

Figure (5.1.15) Transect (J63) ~ time plot of 7-year mean surface salt flux ...........178

Figure (5.1.16) Transect (J63) ~ time plot of 7-year mean surface heat flux ............180

Figure (5.1.17) Transect (J63) ~ time plot of 7-year mean ice thickness ................182

Figure (5.1.18) Transect (J63) ~ time plot of 7-year mean surface (5 m) salinity (ppt) ..184

Figure (5.1.19) 7-year mean vertical profiles of temperature and salinity at
selected locations, from the ice-ocean-mixed-layer model .........................186

Figure (5.2.1) 7-year mean ice compactness and thickness fields from
ice-ocean-mixed-layer model .......................................................187

Figure (5.2.2) 7-year mean distributed ice thickness in the Greenland Sea, Barents
Sea and Arctic Basin ..................................................................188
Figure (5.2.3) 7-year mean ice concentration in the Greenland Sea, Barents Sea and Arctic Basin .................................................................189

Figure (5.2.4) 7-year mean fraction of area covered by ice with compactness greater than or equal to 0.2 in the Greenland Sea, Barents Sea and Arctic Basin ....190

Figure (5.3.1) Annual (1983) mean ocean surface velocity from the ice-ocean-mixed-layer model ..............................................................191

Figure (5.3.2) Monthly mean streamfunction from the ice-ocean-mixed-layer model ....192

Figure (6.1.1) Seasonal variations of ice concentration from Ice-Ocean-Only, Mean-Thermal, Ice-Only-Mean-Thermal and observation in the Arctic, Greenland and Barents Seas ......................................................199

Figure (6.1.2) Deviation of the seasonal ice concentration from the 7-year seasonal mean in the Arctic, Greenland and Barents Seas from 1979 to 1985 ..........200

Figure (6.2.1) 7-year mean distributed ice thickness, ice advection rate and ice melting rate in the Greenland Sea .................................................201

Figure (6.2.2) Ice outflow at the Fram Strait, the Barents Sea Opening and transect J7 .202

Figure (6.2.3) Seasonal variations of distributed ice thickness and ice advection in the Greenland Sea .................................................................203

Figure (6.2.4) Distributed ice thickness and ice advection deviations from their 7-year mean values in the Greenland Sea.................................204

Figure (A2.1) Ice melting rate.................................................................219

Figure (A2.2) Mixed layer salinity, mixed layer depth, and total salt from the salinity mixed layer model .................................................................219
List of Tables

Table (3.6.1) Vertical level thicknesses (in meters) for the ocean model..........................59
Table (3.8.1) River runoff.................................................................................................61
Table (3.8.2) Monthly net precipitation rates in cm/month.................................................62
Table (3.9.1) Streamfunction values (in Sv) at grid cells along the Bering Strait.............65
Table (3.9.2) Streamfunction values (in Sv) at grid cells along the Denmark Strait........65
Table (3.9.3) Streamfunction values at grid cells along the Faroe-Shetland Passage.......66
Table (4.10.1) Statistics of model ice drift and buoy drift over the whole region in 1979-1985.................................................................92
Table (6.1.1) Correlation coefficients between simulated and observed interannual ice concentrations .................................................................196
Table (6.1.2) Correlation coefficients from the prognostic study, the ice-ocean-mixed-layer simulation and the mixed layer sensitivity study...........196
Table (A2.1) Parameters used in different test cases.........................................................216
Chapter 1. Introduction

The polar oceans are thought to play a significant role in the global climate system and have recently taken on particular importance in the context of possible climate changes due to carbon dioxide-induced warming in high latitudes. A prominent feature of the polar oceans is the presence of an ice cover over large areas. The ice cover on the ocean substantially alters heat, salt and momentum transfers between the atmosphere and ocean, and hence has the potential to alter atmospheric and oceanic circulation. The oceanic circulation, in turn, affects the ice dynamic and thermodynamic behavior, and seasonal growth and decay. These effects are particularly significant at the ice margins where seasonal characteristics are largely controlled by the oceanic heat flux. Parameterizing the close interaction between the ice and ocean is essential for examining the behavior of the polar ice-ocean systems and for identifying its effect on the global climate system. However, the ice-ocean interaction mainly occurs in the upper ocean which is in a state of turbulence due to wind, ice drift, surface cooling and salt rejection. Because of its importance in coupling the ice and ocean, this oceanic mixed layer requires special attention for a better understanding of the polar ice-ocean systems.

Although the last few decades have seen a continuing increase in the study of the polar oceans due to the realization of their importance in the global perspective, the study of polar oceans still lags behind the study of the rest of the world's oceans. This is because the remoteness and the harsh environment of the polar regions impose severe logistic constraints on field work and make data collection expensive and difficult. As a result, with a few notable exceptions, less observational and experimental data is available for the polar regions than the other regions in the world.
In addition to field data collection, another useful procedure for understanding the role of the polar oceans in the global climate system is to use numerical models to carry out computer simulations of the interactive ice-ocean system in these regions. As a complement to field work, numerical modeling has an advantage of being less expensive and easier to control and repeat than the field operations in the forbidding polar regions. Appropriate and comprehensive numerical modeling in conjunction with analysis of observations can increase our knowledge of the ice production, different ocean processes and the ice-ocean interaction. For this purpose, there has been an evolution toward better coupled ice-ocean numerical models for the ice-ocean system in the polar regions.

1.1 Previous Research

In early stages, ice and ocean models were pursued separately. For ice modeling where oceanic parameters were prescribed, Maykut and Untersteiner (1971) developed a one-dimensional thermodynamic ice model which was simplified by Semtner (1976a) to be more suitable for large-scale numerical modeling. Parkinson and Washington (1979) used Semtner's model, with ice dynamics included in a simple ad hoc manner, to simulate the ice variations in the Arctic and Antarctic. To examine the role of ice dynamics in sea ice growth, drift and decay, Hibler (1979) developed a full dynamic-thermodynamic ice model with a nonlinear viscous-plastic rheology which exhibited realistic properties in ice simulation in the Arctic.

On the other hand, there have been relatively few uncoupled ocean modeling studies for the polar oceans. Galt (1973) used a barotropic model and Hart (1975) used a two-layer model to simulate the Arctic Ocean dynamics and circulation. Semtner (1976b) conducted an extensive ocean modeling study of the Arctic Ocean wherein ice properties were prescribed, and the general ocean circulation was simulated using a full three-
dimensional baroclinic ocean model. This study brought to light some interesting features of the ocean circulation in that region such as the velocity fields, the integrated flow pattern and the temperature and salinity distributions.

Recently, considerable effort has been expended to develop coupled ice-ocean models to identify the interactive effects between the ice and ocean. Many of the coupled ice-ocean models were constructed for studies of the marginal ice zone on a short time scale (Roed and O'Brien (1983); Hakkinen (1986); Ikeda (1988) and Kantha and Mellor (1989)). However, Hibler and Bryan (1987) and Semtner (1987) have coupled certain ice models with a full three-dimensional baroclinic ocean model of the Arctic, Greenland and Norwegian Seas and carried out comprehensive studies of those regions. Hibler and Bryan (1987) coupled an ocean model similar to that of Bryan (1969) with the full Hibler (1979) dynamic-thermodynamic ice model. Their model is a robust diagnostic one with a 160 km horizontal resolution and 15 vertical levels of different thicknesses. The reason why the model is called 'robust diagnostic' is that the model's temperature and salinity are constrained weakly to climatological observation. This procedure allows seasonal variations in the ice ocean system to be modeled while constraining the climatology of the ocean to agree with observations. This model performed well in predicting the location of the ice edge and generating realistic ocean circulation, although it had certain limitations due to its diagnostic nature and crude resolution. The ocean model used by Semtner (1987) was also based on that of Bryan (1969) and the ice model was a simplified version of the Hibler (1979) ice model similar to the cavitating fluid model of Flato and Hibler (1992). His coupled ice-ocean model had 110 km horizontal resolution and was fully prognostic in that the climatology of the model was predicted within the constraints of boundary forcing. The long-term prognostic integration also produced a realistic seasonal cycle of the ice cover, indicating the importance of coupling ocean circulation with ice circulation.
Subsequent to these studies, the Semtner (1987) model was further used by Fleming and Semtner (1991) to study the effect of interannual ocean forcing on the ice in the North Polar Region. Meanwhile the Hibler and Bryan (1987) model was modified to an 80 km resolution while still retaining 15 vertical levels by Ranelli (1991) and Ries and Hibler (1991). The study by Ranelli focused on the prognostic characteristics of this model while Ries examined the seasonal and interannual characteristics of a diagnostic version of this model over a several year period. In a diagnostic simulation, the ocean temperature and salinity are relaxed to the climatological temperature and salinity while in a prognostic simulation, they are not (see Section 3.2 for more discussion on this subject). The Ries study closely followed the Hibler and Bryan investigation and was able to demonstrate the importance of ocean circulation on the interannual variability of the ice margin. However, the results of both the Ranelli and Ries studies indicated that problems remained in the vertical resolution in shallow regions, in treating penetrative convection, and in the simulation of inflow into the Arctic basin through the Fram Strait. Another Arctic study by Riedlinger and Preller (1991) also found deficiency with their model's coarse resolution (127 km x 127 km). Their coupled ice-ocean model was based on the Hibler (1979) ice model and Bryan ocean circulation model. The ice-ocean coupling, however, was slightly different from that of Hibler and Bryan (1987). The model performed well on simulation of seasonal trends in ice growth and decay in the North Polar Region. However, excess ice growth in the Greenland and Barents Seas in fall and winter appeared to be due to the model's poor resolution of narrow currents, such as West Spitsbergen Current. The inflow at the Fram Strait is also poorly simulated by Piacsek et al (1991) and in their 127 km resolution model, the Atlantic water does not go far enough north.

It is clear from the aforementioned studies that the oceanic heat flux distribution plays an important role in determining the location of the ice margin, the ice thickness and ice concentration. Ostensibly the oceanic heat flux is closely related to the ocean vertical
mixing process which is an important aspect of oceanic circulation in the layers immediately below the ocean surface. These upper ocean layers are almost always in a state of turbulence. They are stirred by wind and surface cooling or salt rejection, resulting in active turbulent vertical mixing within them. In a three-dimensional large-scale ocean circulation model, like those adopted by Hibler and Bryan (1987) and Semtner (1987) for example, vertical mixing is usually parameterized by constant vertical diffusivity coefficients in conjunction with a large scale convective overturning when the ocean becomes vertically unstable. These procedures are easy to implement but do not capture the variations of mixing effects, particularly in the upper ocean mixed layer. Consequently, the incorporation of appropriate mixed layer physics into a coupled large-scale general ice-ocean model is an important area of research.

One approach to improve the vertical mixing processes is to introduce a local turbulence closure model into an ocean model. A system of parameterized second-order equations has been given and systematically simplified by Mellor and Yamada (1974) in a hierarchy of turbulence closure models. Some researchers designed closure models of even higher order, Andre and Lacarrere (1985), for example. The Mellor and Yamada (1974) schemes of different simplification levels have been applied to ocean or coupled ice-ocean models (Klein and Coantic, 1981; Blumberg and Mellor, 1983, 1987; Mellor and Kantha, 1989). The high-level schemes of Mellor and Yamada (1974) demand considerable computer time while the low-level schemes are related to the classic representation by local eddy coefficients as functions of mixing length and turbulence strength. The concept of non-constant parameterizations for local eddy coefficients was explored in a different way by some other studies (e.g. Pacanowski and Philander, 1981) where the eddy coefficients were fit as functions of local Richardson number through laboratory measurements and studies. This procedure was also investigated by Ries and Hibler (1991) with only limited success.
In contrast to the above mentioned turbulence closure model, which is limited essentially to numerical interactions between variables at neighboring grid points, the so-called transilient turbulence models (see Kraus, 1988; Gaspar et al, 1988) present a non-local representation of turbulence exchange processes. In this kind of model, the mixing at one point is affected by the whole water column. Therefore more coefficients than the classic eddy coefficients are needed depending on how many vertical levels the water column is discretized into.

Another approach for ocean mixed-layer modeling is to use vertically integrated mixed layer models. The integral mixed layer models involve a priori assumptions about the vertical structure of the upper ocean which is approximated to be a well-mixed quasi-uniform layer in terms of the mean physical quantities. The first and widely used one-dimensional mixed layer model for oceanic use was the one developed by Kraus and Turner (1967), and was described in detail by Niiler and Kraus (1977). Another type of mixed layer model was proposed by Garwood (1977). In that model an entrainment hypothesis dependent upon the relative distribution of turbulent energy between horizontal and vertical components was offered as a plausible mechanism for governing both entrainment and mixed layer retreat.

Most of the mixed layer models have been used for modeling the oceans in low or middle latitudes with one of the attempts being to obtain a better ocean surface temperature. Special care and corresponding modifications are need, however, to deal with the oceanic mixed layer under polar sea ice cover. In contrast to the rest of the world's oceans, the changes of the water properties in the upper ocean in the polar regions are dominated by the changes of salinity instead of temperature. This is because the temperature of the water immediately under the ice is close to the freezing point while the constant freezing or
melting of sea ice results in an active salt exchange between the ice and the ocean. Therefore the amount of mixing caused by the buoyancy change is largely controlled by surface salt fluxes while the mixing that can be due to the surface mechanical energy flux is suppressed by the ice cover that prevents winds from directly reaching the ocean. Another feature of the ice-covered oceans is that the existence of ice leads makes the surface cooling or salt rejection complicated because of the localization of the leads. Some studies have targeted this situation by constructing models to simulate local mixed layer circulation under leads. An example is the work by Kozo (1983) where realistic mixed layer circulation patterns under different conditions of geostrophic currents beneath an idealized two-dimensional ice lead were generated. Mixed layer models like the one by Kozo seem suitable to be exclusively used in an ice-covered region. However, it is difficult to extend them to ice free regions that make up the major portion of the Greenland and Norwegian Seas.

The Kraus and Turner model was later extended by Lemke and Manley (1984) to include both a mixed layer and pycnocline for the ocean under polar sea ice. The Kraus and Turner model is included in Lemke and Manley’s model as a special case. Lemke’s mixed layer-pycnoclinic model has been coupled with Semtner’s thermodynamic ice model and Hibler’s dynamic-thermodynamic ice model respectively for the studies of ice distribution in the Arctic and Antarctic (Lemke, 1987; Lemke et al, 1990), while the Kraus and Turner model has been coupled by Houssais (1988) with, basically, Roed’s (1984) ice model under subarctic conditions.

The recently developed vertically integrated mixed layer models have been embedded into general ocean circulation models. The main advantages of embedding a mixed layer model into an ocean model are the simplicity and economy of implementation (Kraus, 1988). Another superiority of using mixed layer models over others is that they
can use relatively large time steps without causing computational instabilities which is important for long-term simulations. Although the basic assumptions for mixed layer models may oversimplify the conditions in the real upper ocean, they conform to the structure of most general ocean circulation models which adopt a finite difference method. The leading example of the embedding efforts is the work of Adamec et al (1981) in which conservation of conservative properties such as temperature, salinity and momentum is maintained in the coupling of the mixed layer with the fixed finite difference grid in the ocean model. Recently, Resnyanskiy and Zelen'ko (1991) adopted a different embedding approach in which the mixed layer depth increase during entrainment is determined by comparison of total buoyancy of the water column between the ocean circulation model and mixed layer model. The expression in Resnyanskiy and Zelen'ko for vertical mixing processes (Eq.s (3.4.1) and (3.4.2)) within and immediately below the mixed layer seems more straightforward than that in Adamec et al (1981) although there is no fundamental difference between both approaches in terms of expressions for vertical mixing.

The technique of Adamec et al has been used by Houssais and Hibler (1993) to embed the Kraus and Turner mixed layer model into a two-dimensional ice-ocean model for the subarctic marginal ice zones. The two-dimensional ice-ocean model was converted from the Hibler and Bryan ice-ocean model. Its simulation results for a transect across the Greenland Sea along a fixed latitude show that the penetrative convection delays and reduces ice edge expansion toward open water and, therefore, yields a better agreement with observation.

1.2. Thesis Work

Considering the previous studies briefly reviewed above, it is felt that there are two issues that particularly need to be addressed for the Arctic ice-ocean system. One is the
relatively crude resolution used in previous Arctic model studies has certain limitations in resolving the complex topography, coastlines and narrow passages of the oceans and in simulating the complicated lateral and vertical processes during the cycle of ice melting and growing. The other is the degree to which inclusion of robust oceanic mixing physics into a three-dimensional general ice-ocean circulation model can produce a more realistic simulation of the whole ice and ocean system. Since vertical penetrative convection is important in the Greenland and Norwegian Seas (Houssais and Hibler, 1993), it is important to assess its role in a full three dimensional model including other areas.

The research work of this thesis will focus on the above problems. First, an improved vertical and horizontal resolution version of the Hibler and Bryan model is developed. In the development of this high resolution model (40 km horizontal resolution and 21 vertical levels of different thicknesses), the inflow and outflow conditions for the Bering and Denmark Straits and the Faroe-Shetland passage are established from experimental data. A time-varying precipitation has been introduced into the model in an attempt to have a better salt budget as well as ocean stratification. Since a mixed layer is not embedded into the ice-ocean model at this point, the ice-ocean model has a conventional fixed mixed layer. That is, the first ocean level will serve as a de facto mixed layer. This model (hereafter referred as ice-ocean-only model) is then used to examine the predicted mean fields and seasonal and interannual variability of the Arctic, Barents, Greenland and Norwegian Sea ice-ocean system over a several year period (1979-85). The effects of the high resolution, the ice interaction, the lateral open boundary conditions and the atmospheric forcing fields on the oceanic circulation and ice properties in that region are systematically investigated.

Secondly, once this more conventional investigation is complete, a mixed layer model of Kraus and Turner type is imbedded into the ice-ocean model. The reason for
selecting the Kraus and Turner mixed layer model is that it is simple to implement and it has been well tested for ice-covered oceans by previous studies (Lemke et al and Houssais et al as cited before). Embedding the mixed layer model into a general three-dimensional ice-ocean model is therefore a natural extension of Houssais and Hibler's (1993) work. Before embedding the full mixed layer model, a corresponding one-dimensional salinity mixed layer model is created and tested in order to ensure that the model is functioning normally and, particularly, that the conservative properties are properly modeled. After the fully coupled model (hereafter referred as ice-ocean-mixed-layer model) has been constructed, tests were made to examine the performance of the model. The effects of the embedded mixed layer on different processes is assessed by comparing the ice-ocean-mixed-layer model to the previous ice-ocean-only model through a parallel simulation of the ice-ocean system in the North Polar Region. Both the ice-ocean-only and the ice-ocean-mixed-layer simulations are diagnostic simulations.

The study of this thesis is presented in several chapters. Chapter 2 briefly gives some features of descriptive oceanography and climatology of the North Polar Region, so that observed information will be available for comparison with simulation results. Chapter 3 systematically gives the formulation of all the related models (the ice, ocean and mixed layer models), the model's numerical aspects and the coupling procedure. Chapter 4 presents the main simulated results from the ice-ocean-only model and some output of the related sensitivity studies. Chapter 5 presents the experimental results from the ice-ocean-mixed-layer model and comparisons are often made between the two major models. Chapter 6 addresses the characteristics of interannual variability in order to identify the essential physical processes and conditions. Finally Chapter 7 summarizes the main conclusions of the study.
Chapter 2. Description of the North Polar Region

This chapter briefly presents some information regarding the basic geophysical features of the North Polar Region in the hope that the information will help for further presentation of the thesis research and lay a basis for comparison with the numerical results as well. For more information about the physical oceanography and climatology of the region, one can find excellent reviews by Coachman and Aagaard (1974), Lewis (1982), Carmack (1986), Hopkins (1988), Rudels (1987) and Barry (1989). Whereas for more information about the general characteristics of the North Polar ice cover, one can read comprehensive descriptions given by Hibler (1980b, 1989) and Gow and Tucker (1990). The theses of Semtner (1973) and Ranelli (1991) also serve as good sources of information for understanding the North Polar Region. A majority of the following review comes from these authors' presentations.

2.1 Physical Features

The numerical model used in this study covers the whole North Polar Region (North Polar Oceans) defined here as the oceans bordered by Europe, Siberia, Alaska, Canada and Greenland. This region consists of the Arctic Ocean (or Arctic Basin as called later in this thesis), the Barents Sea and the GIN Sea. Fig.(2.1.1) gives the names of major geographical features in this region in order to provide a reference frame for subsequent description of the model and the numerical results.

As shown in Fig.(2.1.1) the Arctic Ocean consists of the central polar basin and its five major surrounding seas: the Kara, Laptev, East Siberian, Chukchi and Beaufort Seas. The Arctic Ocean is the largest mediterranean (enclosed) sea in the world with an area of approximately $10^7 \text{ km}^2$. Its main communication with the rest of the world oceans takes
place through four restricted passages: the Canadian Arctic Archipelago, the Fram Strait, the Barents Sea and the Bering Strait. The Fram Strait between Greenland and Spitzbergen is the only deep and wide outlet (about 2700 m deep and 600 km wide) and, hence, the most important passage that connects the Arctic Ocean to southern oceans. In contrast to the Fram Strait, the Bering Strait is very shallow and narrow, only about 45 m deep and 85 km wide, and is the only connection between the Arctic system and the Pacific. The Canadian Arctic Archipelago consists of numerous passages connecting the Arctic and Atlantic Oceans, while the connections between the Arctic Ocean and the Barents Sea are mainly through two openings. One is the opening between Spitzbergen and Franz Josef Land and the other Franz Josef Land and Novaya Zemlya.

Some physical features of the North Polar Region are effectively illustrated in Fig. (2.1.2) with a contour map of the ocean bottom topography. As shown in this figure, the Arctic Ocean has two deep ocean basins (about 4000 m deep): the Canadian Basin and the Eurasian Basin. These two basins are divided by the 1600 m deep Lomonosov ridge. In addition to the deep basins, the Arctic Ocean has large areas of continental shelf associated with the five major bordering seas as cited before. The shelf occupies a significant percentage of the area of the whole Arctic Ocean but contains a small fraction of the whole volume of water. The continental shelf typically breaks at 200 m depth. North of North America the shelf is generally less than 100 km in width while north of Siberia it is up to 800 km wide, the widest shelf in the world. As pointed out by Coachman and Aagaard (1974), all the major continental rivers reaching the Arctic Ocean, with the exception of the Mackenzie and Yukon, flow into these marginal seas. Thus these shallow seas, with a high ratio of surface area to total volume and a large input of fresh water in summer, greatly affect the surface water conditions in the Arctic Ocean.
As shown in Fig.(2.1.2), the Barents Sea is a deep shelf sea with quite an irregular bottom of depths ranging from generally less than 200 m to 400 m in some parts. The Barents Sea is bounded by Europe on the South, Novaya Zenyla on the east and Franz Josef Land and Spitzbergen on the north with an area of approximately $1.5 \times 10^6$ km$^2$. Its main connections with the Arctic Ocean are through the two openings between Spitzbergen and Franz Josef Land and between Franz Josef Land and Novaya Zemlya as mentioned earlier. Its connection with the GIN Sea is through the Barents Sea Opening (Hopkins, 1988) between Norway and Spitzbergen.

The GIN Sea is defined here as the collective waters of Greenland, Iceland and Norwegian Seas, covering an area of approximately $2 \times 10^6$ km$^2$. The GIN Sea plays a principal role in connecting the waters of the Arctic Basin and the rest of the world oceans. The GIN Sea comprises three major basins: the Greenland, Norwegian and Lofoten Basins, separated by mid-ocean ridges. It is connected to the Arctic Ocean via the Fram Strait, to the Barents Sea via the Barents Sea Opening and the North Atlantic through the Denmark Strait and the Faroe-Shetland Passage.

2.2 Ice Characteristics

A feature contributing to the uniqueness of the Arctic oceanography is that large areas of the Arctic Ocean and its adjacent seas are covered with a layer of ice formed from freezing seawater. The ice is a mixture of young and old floes that are highly variable in thickness due to the complex annual cycle of growth, decay and ice drift. The obvious manifestation of the ice growth and decay and ice motion is the variation in the ice concentration and extent with considerable seasonal as well as interannual variability (Fig.(2.2.1)). In the winter around February, the ice reaches its maximum with almost all the Arctic Ocean and a significant part of the Barents and GIN Seas ice covered. In the
summer at late August or early September the ice retreats to its minimum, yielding about half of its winter extent to open water with an ice coverage of approximately 8 million km² (Barry, 1989). The Barents Sea is subject to the largest seasonal variations: from considerably ice-covered winters to almost ice-free summers. In the GIN Sea, the ice cover is normally confined to a narrow (200km) band of ice along the Greenland coast with an interesting feature that there is almost continuous existence of ice off the coast even during the summer. This is credited to the continual ice flow from the Arctic Basin into the North Atlantic. However, the extent of the ice can vary substantially from year to year. Many studies, according to Barry (1989), have demonstrated that the interannual variations in ice extent are primarily related to interannual changes in atmospheric and oceanic circulation. The characteristics of the interannual variability in ice extent are well illustrated in Fig.s (2.2.2) and (2.2.3), where the changes of ice extent at the Greenland and Beaufort Seas over many years are shown.

A prominent characteristic of sea ice is its constant drift. Being an enclosed ocean, the Arctic Ocean ice motion is due to its relatively landlocked character: most of the ice is contained in the Arctic with a main outlet at the Fram Strait. While the main driving forces that move the ice cover are from wind and ocean currents, this landlocked character lays a basis for a general pattern of the ice circulation in the Arctic Basin. This general ice motion pattern is known to consist of an anticyclonic (clockwise) gyre in the Beaufort Sea and a transpolar drift stream (Fig.(2.2.4)). The transpolar drift stream carries ice from the Siberian Sea or even Chukchi Sea, passing it across the North Pole and the Fram Strait and down the east coast of Greenland with a large volume. Recently Colony (1990), based on observation, estimated a $0.9 \times 10^6$ km²/year areal ice transport out of the Fram Strait. As for the Bering Strait, observations show alternating south and north ice motion closely related to local geostrophic wind (Kozo et al, 1988) without significant net ice transport through it
To obtain an idea how the ice moves, some buoy drift tracklines (Thorndike et al, 1983) are shown in Fig.(2.2.5).

The almost constant motion, deformation, growth and decay of the ice result in a high spatial variability of the ice thickness distribution. The main characteristic of observed variation of ice cover in the Arctic is thick area off the Canadian Archipelago, averaging over 6 m and less thick areas off the Siberian Coast and North Slope of Alaska. The thickest ice areas result from ice converging and ridging due to the Beaufort ice circulation and the transpolar ice stream, while the less thick areas come from ice diverging due to the same ice motion pattern. In the Barents and GIN Seas the ice is normally less than 2 m.

The ice cover does not maintain a passive existence; it's growth, decay and drift are closely related to both the dynamic and thermodynamic variations in the atmosphere and ocean. In particular, as pointed out in Hibler (1980b), rates of growth and decay of the ice depend on the distribution of ice thickness as well as on the thermal characteristics of the atmosphere and ocean. The thicknesses are, in turn, modified by the ice transport and deformation patterns which are driven by atmospheric and oceanic circulation. In addition the presence of the dynamic ice cover can substantially alter heat, salt and momentum transfers between the atmosphere and ocean, and hence has the potential to alter atmospheric and oceanic circulation (Hibler, 1989). According to Coachman and Aagaard (1974) the manifestations of the ice cover's influence on the ocean are:

1. The water temperature of the near-surface layer in the presence of ice is always close to the freezing point for its salinity by the change-of-phase process.

2. Salt is excluded from the ice to varying extents, but the water under the ice is always enriched in salt by ice growth. The dependence of water density on temperature and salinity is such that close to the freezing point density is almost solely a function of salinity.
Therefore, ice formation can increase the density locally and some vertical convection may result.

(3) In the transfer of momentum from the atmosphere to the ocean, the wind must act on the sea through the ice.
The above three points are worth citing because they are particularly important for the understanding of the oceanic mixed layer under the ice cover.

2.3 Ocean Features

(i) Water Masses

The North Polar oceans consist of several distinct water masses according to their origin or temperature and salinity characteristics. These water masses exhibit large spatial variations, reflecting their different mixing histories or formation processes.

From the review by Coachman & Aagaard (1974) the Arctic Ocean is composed of four different water masses, the surface water, the subsurface layers, the Atlantic water and the bottom water, with their vertical profiles of temperature and salinity shown in Fig.(2.3.1). The surface water, or the mixed layer, is a 30-50 m deep, low saline (about 32 ppt) layer. The low surface salinity (Fig.(2.3.2)) is believed to be mainly due to the large volume of river runoff, low-salinity inflow from the Bering Strait and, probably, precipitation. Under ice cover, the temperature is close to the freezing point while in ice-free regions the temperature may be above freezing. During the freezing seasons, this layer is made relatively homogeneous locally by the mixing process due to haline convection and wind or keel stirring. In the summer, the mixed layer retreats into a shallow (10-20 m) fresher surface layer by ice melt. The properties of the mixed layer vary over the Arctic Basin. It is deeper (about 100 m) and more saline (about 33 ppt) close to the Fram Strait, while its salinity may be less than 30 ppt in the Beaufort Sea. This mixed layer is important
in heat, mass and momentum exchange between the ice, ocean and atmosphere, and should be represented as well as possible by a mixed layer model.

Beneath the relatively uniform mixed layer is a 100-200 m thick pycnocline, where water density increases rapidly. The increased density is due to a strong halocline with salinity increasing with depth to 35.0 ppt at 200 m, while the temperature remains at the freezing point down to 100 m. This feature excludes the possibility of forming a pycnocline through mixing between the mixed layer and the underlying warm Atlantic water. However, the pycnocline at the area close to the Bering Strait may be affected by the incoming Pacific water. The pycnocline is somewhat different in the major basins. For example, the salinity in the Eurasian basin increases with depth to a value of that of Atlantic water by 200 m, while the salinity in the Beaufort Sea does not reach this value until 300 m.

Further down lies a 600 m thick layer of Atlantic water commonly defined as water with temperature greater than 0°C. This water, of Atlantic Ocean origin, flows northward through the GIN Sea and enters the Arctic Ocean through the Fram Strait and via the Barents Sea. The maximum temperature of the water is above 2°C near the Fram Strait. As the water spreads over the Arctic Basin, the temperature gradually decreases, down to 0.5°C or below in the Beaufort Sea because of cooling. The Atlantic water is a major source of heat to the Arctic, probably storing enough heat to melt the ice cover. Fortunately upward transfer of heat is prevented by the strong vertical stratification resulting from the layer of fresh surface water (Ranelli, 1991).

The bottom water, or deep water as it is often called, contains about 60 percent of the water in the Arctic Basin. The deep water has an almost constant salinity of about 34.94 ppt, while the temperature is usually below 0°C and decreases slightly with depth, reaching
-0.9°C at the Eurasian basin. Because of the presence of the Lomonosov Ridge, there is a slight difference in temperature between the Eurasian basin and Canadian basin.

The water masses in the GIN Sea are normally classified (Carmack & Aagaard, 1973) to consist of three different types, Polar water, Atlantic water and deep water, basically according to their origins. The Polar water, originating from the Arctic Basin, extends from the surface down to a mean depth somewhat greater than 150 m. The temperature varies between the freezing point (near the surface) and 0°C (at the bottom of the water). This water has a 30 ppt salinity at the surface and a 34 ppt or more at its bottom. Polar water is basically Arctic surface water that flows southward into the GIN Sea through the Fram Strait and is generally confined to be along the east Greenland coast in a relatively narrow region.

The Atlantic water enters the GIN Sea via the Faroe-Shetland Passage. In the east of the GIN Sea, the Atlantic water forms a water column from the surface to a depth of about 800 m. However, a part of this water meets the Polar water in the west GIN Sea and forms an intermediate layer beneath the Polar water. The temperature of the Atlantic water under the Polar water is greater than 0°C while its salinity increases from the Polar water downward until reaching a value between 34.88 and 35.0 ppt. As for the Atlantic water in the east GIN Sea, its temperature is frequently above 5°C in the upper 100 m and the salinity increases rapidly in the first 200 m until reaching a maximum of 35.0 to 35.2 ppt. A consequence of this warm current is permanent open water over much of this region.

(ii) Ocean Currents

The surface circulation in the North Polar Region has been deduced from the observed drift of various manned ice islands, floe stations and ships. A chart of surface circulation taken from the U.S. Navy Arctic Atlas, H.O. Publication 705 (1963) is shown
in Fig.(2.3.3). This figure gives the whole circulation pattern as well some local characteristics. The averaged circulation pattern is quite similar to that of ice motion: a clockwise movement in the Beaufort Sea (Beaufort gyre) and a stream moving across the North Pole (transpolar stream). The transpolar stream, with average speed of 2-3 cm/s, converges toward the Fram Strait after passing the pole and exits there from the Arctic Basin as part of East Greenland Current. As shown in Fig.(2.3.3), the Pacific water flow entering the Arctic via the Bering Strait also contributes to the surface circulation. One part of the Pacific water seems to join the transpolar stream while another part moves along the Alaskan coast towards the Beaufort Sea. Although there are water transport fluctuations across the Bering Strait with water flowing southward, the average is estimated as an inflow of 0.8 (Aagaard & Carmack, 1989) to 1.5 (Hopkins, 1988) Sv (1Sv = 10^6 m^3/s).

The Fram Strait serves as both an entry and exit port. There are two different surface currents flowing through the Fram Strait. One is the West Spitzbergen Current flowing into the Arctic from the GIN Sea, and the other is the Arctic cold water flowing into the GIN Sea from the basin as the East Greenland Current. The East Greenland Current is estimated to transport about 5 Sv of water southward (as cited by Ranelli, 1991) while the net inflow into the Arctic through the Fram Strait may be in the neighborhood of 1 Sv (Aagaard, et al, 1987). The West Spitsbergen Current is the Arctic's major Atlantic water supplier via the Fram Strait. Note from Fig.(2.3.3) that the warm Atlantic water sinks near the west of Spitsbergen when it encounters the cold Arctic water and becomes denser. There is a great disparity among the estimates of the water volume carried by the West Spitsbergen Current, ranging from 1.4 Sv to 7.1 Sv (reviewed by Rudels, 1987). The study by Rudels (1987) found that a mass transport of 1.92 x 10^9 kg/s is carried with the current; half of the water flows into the Arctic Basin and the other half returns as warm Atlantic water in the East Greenland Current. That is, according to Rudels, about 0.9 Sv Atlantic water actually enters the Arctic Basin. At the east of Spitsbergen there is also a
smaller current called East Spitsbergen Current which seems to split from the transpolar stream and flow out into the Barents Sea.

The Faroe-Shetland passage is the major entry port for Atlantic water to enter the GIN Sea. The water comes with a relatively high temperature and salinity (~ 12°C and 35.0-35.6 ppt) and an estimated average transport of 8 Sv. The majority of the water flows north at the eastern side of the GIN Sea as the Norwegian Current (Fig.(2.3.3)), however, a portion of it recirculates in the middle of the GIN Sea, joins the East Greenland Current and exits at the Denmark Strait. The northward flow of this water will split again. One part of it goes into the Barents Sea through the Barents Sea Opening as the North Cape Current carrying about 3 Sv of water (cited in Aagaard and Carmack, 1989). Some of this flow will eventually enter the Arctic Basin (~ 1 Sv) through the openings connecting the Arctic Ocean and the Barents Sea. The other part meets the outflowing Arctic cold surface water and sinks to a level of 200-400 m, and then flows into the Arctic in the intermediate layers as a part of West Spitzbergen Current. This part of Atlantic water seems to form a cyclonic circulation around the Eurasian and an anti-cyclonic circulation in the Canadian Basin in the subsurface layers before returning to the Greenland Sea (Fig.(2.3.4)).

The Denmark Strait serves as a main exit port of the East Greenland Current. The majority of the Arctic water carried by the current flows out of Greenland Sea at this strait except for the water that flows along the east of Iceland as the East Icelandic Current. There is a small current, however, flowing into the Greenland Sea at the west of Iceland.

The bottom water has shown extreme uniformity of conditions down the deep basins. Measurements of such bottom water often are overwhelmed by uncertainties. Consequently the knowledge of the bottom-water circulation is very limited. It is assumed
to be a slow flow of 1 cm/s or less, mostly moving as a unit without significant shear due to essentially zero density gradient (Coachman and Aagaard, 1974).
Figure (2.1.1) Map of the North Polar Region identifying major geographical features (from Ranelli, 1991).
Figure (2.1.2) Contour map of the bathymetry of the North Polar Oceans (from Coachman and Aagaard, 1974).
Figure (2.2.1) The average and extreme seasonal limits of the Arctic sea ice extent for ice concentrations ≥ 1/8 (from Barry, 1989).
Figure (2.2.2) Extreme sea ice conditions at the end of (a) February, and (b) August, from 1966 to 1975 (adopted from Vinje, 1976 by Hibler, 1980b).
Figure (2.2.3) Record of ice conditions off the North Slope of Alaska. The ordinate is the distance in nautical miles from Point Barrow northward to the boundary of 50% ice concentration on September 15 (Hibler, 1980b).

Figure (2.2.4) The pattern of mean ice drift and surface water flow in the Arctic Ocean (from Hibler, 1980b).
Figure (2.2.5) Recorded tracklines of buoys deployed in the Arctic in 1981 and 82 (from Thorndike et al, 1983).
Figure (2.3.1) Vertical profiles of temperature and salinity at selected locations in the North Polar Region (from Coachman and Aagaard, 1974).
Figure (2.3.2) Summer surface (5 m) salinity of the North Polar Oceans (from Coachman and Aagaard, 1974).
Figure (2.3.3) Pattern of surface circulation in the North Polar Oceans
(from U.S. Navy H.O. Publication 705).
Figure (2.3.4) Pattern of the Atlantic water circulation in the North Polar Oceans (from U.S. Navy H.O. Publication 705).
Chapter 3. Model Description

The model is developed by first generating an ice-ocean model of high resolution for the Arctic Basin, Barents and GIN Seas based on the Hibler and Bryan (1987) model and then embedding the Kraus and Turner (1967) mixed layer model into the ice-ocean model. The Hibler and Bryan model, an ice-ocean model for the Arctic Ocean with 160 km horizontal resolution and 14 vertical levels, is a combination of the dynamic-thermodynamic ice model of Hibler (1979, 1980a) and the multilevel baroclinic ocean model of Bryan (1969). Additional details of the ocean model are available in Semtner (1974, 1986) and Cox (1984). The mixed layer model of Kraus and Turner is a two-level model which is to be embedded into the multilevel finite difference grids of the ice-ocean model with appropriate measures to insure that the heat, salt, momentum and other conservative properties of the ice-ocean system are conserved.

The basic setting for all these models working together is as follows: First, the sea ice model is driven by atmospheric forcing, consisting of geostrophic winds, surface air temperature, humidity, and longwave and shortwave radiations. The ice model then supplies surface heat, salt, and momentum fluxes into the model ocean as surface boundary conditions. The ocean model in turn, via a mixed layer, supplies current and heat exchange information to the ice model. For the ice-ocean-only model defined in Section 1.1, the first ocean level is taken as a *de facto* mixed layer of constant depth and it has more exchange with the second level by using larger vertical viscosity and diffusivity between the top two levels to approximate the effect of vertical mixing. For the ice-ocean-mixed-layer model, on the other hand, the mixed layer is of variable depth determined by the mixed layer model. Once the mixed layer model is embedded, it monitors the evolution of the mixed layer and modifies the whole temperature and salinity structure of the upper ocean. Some details of each model and the coupling and embedding procedure are given below.
3.1 Ice Model

The Hibler (1979) ice model treats the sea ice cover as a two dimensional dynamic and thermodynamic continuum. As far as the representation of ice mass is concerned, the ice model is a two-level model, having two thickness categories: thin ice and thick ice. The thin ice category, more sensitive to heat transfer, represents both open water and an ice thickness up to a small cutoff thickness $h_0$. The thick ice category represents the remaining ice mass in a grid cell. The ice cover is represented by two quantities: mean ice thickness, $h$, of a grid cell and ice compactness, $A$, defined as the fraction of a grid cell covered by thick ice. The dynamic and thermodynamic processes of the ice cover are described by momentum equations and conservation equations for thickness and compactness respectively. The momentum equations require a constitutive law for the relationship between the ice stress and ice displacement or strain rate, while the conservation equations reflect the ice thermodynamic process by incorporating thermodynamic source or sink terms which determine the growth and decay of ice.

The sea ice is considered to move in a two-dimensional plane with the momentum balance being

$$m\frac{Du}{Dt} = -mk\times u + \tau_a + \tau_w - mgV_Hp(0) + F,$$  \hspace{1cm} (3.2.1)

where $u$ is the ice velocity, $m$ the ice mass per unit area, $k$ a unit vector normal to the sea surface, $f$ the Coriolis parameter, $g$ the gravity acceleration, $p(0)$ the sea surface dynamic height, $\tau_a$ and $\tau_w$ the forces due to air and water stresses, and $F$ the force due to internal ice stress. The air and water stress terms are determined from McPhee (1978).

$$\tau_a = \rho_a c_a |U_g|(U_g\cos\phi + k\times U_g \sin\phi),$$  \hspace{1cm} (3.1.2)

$$\tau_w = \rho_w c_w |U_w - u|[(U_w - u)\cos\theta + k\times(U_w - u) \sin\theta],$$  \hspace{1cm} (3.1.3)
where \( \rho_a \) and \( \rho_w \) are the densities of air and water, \( c_a \) and \( c_w \) the air and water drag coefficients, \( \phi \) and \( \theta \) the air and water turning angles (both 25°), \( U_g \) and \( U_w \) the geostrophic wind velocity and the ocean surface current respectively. \( U_w \) is approximated by the velocity in the third level of the model ocean. The force due to internal ice stress \( F \) is given by

\[
F = \nabla \cdot \sigma,
\]  

where \( \sigma \) is ice stress tensor (\( \sigma_{ij} \)) which is related to ice strain rate and ice strength via a nonlinear viscous-plastic constitutive law:

\[
\sigma_{ij} = 2\eta(\dot{e}_{ij}, P)\dot{e}_{ij} + [\zeta(\dot{e}_{ij}, P) - \eta(\dot{e}_{ij}, P)]\delta_{ij} + \frac{P}{2}\delta_{ij}.
\]  

In the above equation, \( \eta \) and \( \zeta \) are the nonlinear bulk and shear viscosities (functions of \( \dot{e}_{ij} \) and \( P \)), \( \dot{e}_{ij} \) the ice strain rate, \( P \) the ice compressive strength. The dependence of \( \eta \) and \( \zeta \) on \( \dot{e}_{ij} \) and \( P \) is normally taken so that the stress state lies on an elliptical yield curve passing through the origin, but for very small strain rates the stress state stays inside the yield curve (Hibler, 1979). The treatment of a variety of other nonlinear rheologies is described by Ip (1993). The ice strength is taken to be a function of ice compactness and thickness according to

\[
P = P^* h \exp\left[-C(1 - A)\right],
\]  

where \( P^* \) (=2.75 N-m^2) and \( C \) (=20) are fixed empirical constants (Hibler, 1979; Hibler and Walsh, 1982).

For ice thickness and compactness, the following continuity equations are used:

\[
\frac{\partial h}{\partial t} = - \frac{\partial (uh)}{\partial x} - \frac{\partial (vh)}{\partial y} + S_h + \text{diffusion},
\]
\[
\frac{\partial A}{\partial t} = -\frac{\partial (uA)}{\partial x} - \frac{\partial (vA)}{\partial y} + S_A + \text{diffusion,}
\]

(3.1.8)

where the diffusion terms, which are small and are described in detail in Hibler (1979), are artificial and added only for numerical stability. \(S_h\) and \(S_A\) are thermodynamic terms given by

\[
S_h = f(h/A)A + (1 - A)f(0),
\]

(3.1.9)

\[
S_A = \begin{cases} 
(f(0)/h_0)(1 - A), & \text{if } f(0) > 0, \\
0, & \text{if } f(0) < 0, \\
(A/2h)S_h, & \text{if } S_h > 0, \\
0, & \text{if } S_h < 0,
\end{cases}
\]

(3.1.10)

with \(f(h)\) the growth rate of ice of average thickness \(h\) due to the atmospheric forcing, and \(h_0 (= 0.5 \text{ m})\) a fixed demarcation thickness between the idealized two ice categories (thick and thin) that are used in the ice model. The ice growth rate \(f(h)\) is calculated following Semtner (1976a) and Manabe et al (1979) with surface heat budget computations similar to Parkinson and Washington (1979), as described in detail in Hibler (1980). In calculating ice thermodynamic growth rate of thick ice \((f(h/A))\), a seven-category ice growth rate calculation (Walsh et al, 1985) is used. In this method, the thick ice is divided equally into seven categories from 0 to twice the mean ice thickness, \(2h\). Each category's growth rate is calculated separately and the whole growth rate of the thick ice is the summation of the seven categories.

Following Hibler (1980), the surface heat budget is given by

\[
(1 - a)F_s + F_L + D_1U_g(T_a - T_0) + D_2U_g[q_a(T_a) - q_s(T_0)] - D_3T_0^4 + (K/H)(T_w - T_0) = 0
\]

(3.1.11)
where $a$ is the surface albedo, $T_0$ the surface temperature of ice, $T_a$ the air temperature, $T_w$ the water temperature, $q_a$ the specific humidity of air, $q_s$ the specific humidity immediately above the ice or snow surface, $F_S$ the incoming shortwave radiation, $F_L$ the incoming longwave radiation, $D_1$ the bulk sensible heat transfer coefficient, $D_2$ the bulk latent heat transfer coefficient (water or ice), $D_3$ the Stephan-Boltzmann constant times the surface emissivity and $K$ the ice conductivity.

The original ice model used by Hibler and Bryan (1987) does not include the effect of precipitation in terms of salt and water budgets, but the insulating effect of snow was approximated by allowing the ice surface albedo to be that of snow (see e.g. Manabe et al, 1979) whenever the surface temperature is found below freezing. In order to include the precipitation effect on the salt and water balance and, hence, the ocean circulation, a certain amount of monthly-varying spatially-changing precipitation (Vorwinkel and Orvig, 1970) is introduced into the models used in this research. A thermodynamically simple approach is adopted, which treats the precipitation either as ice or as water. If the local condition is freezing, the precipitation is taken as an equivalent ice, with the latent heat ignored, which is incorporated into other existing ice if any, whereas if it is melting or the precipitation encounters open water, the precipitation is taken to be water going directly into the ocean. The direct freshwater addition to the ocean is treated as a negative surface salt flux. Whereas the precipitation as equivalent ice will not provide an immediate salt flux. It remains to be melted at some point later in order to provide such a negative salt flux.

3.2 General Ocean Circulation Model

(i) Governing equations

The ocean model is a modified version of Hibler and Bryan (1987) which is based on the models of Bryan (1969) and Cox (1984). It is a finite difference formulation
of a three-dimensional multilevel baroclinic model specially designed for the Arctic Ocean and Barents and GIN Seas. The governing equations describing the general ocean circulation model consist of two horizontal momentum equations, the hydrostatic equation in the vertical dimension, the mass continuity equation for an incompressible fluid, the conservation equations for temperature and salinity and an equation of state. This system of equations is derived through a series of assumptions and approximations. Some basic simplifications are listed here (Ranelli, 1991):

1. The hydrostatic assumption which allows the vertical momentum component equation to be represented by a balance between vertical pressure change and gravity force;

2. The Boussinesq approximation which neglects the effect of water's density variations on mass (inertial) of the water but keeps their effect on the weight (buoyancy). Therefore a mean density can be used in the horizontal momentum equations while the actual density has to be used in the hydrostatic equation;

3. Ocean incompressibility is assumed which leads to a simple continuity equation.

4. The rigid-lid approximation is used which specifies the vertical velocity to be zero at the ocean surface. This approximation results in high-frequency surface gravity waves being filtered out which allows a longer numerical timestep. However, it retains the effects of the waves through a horizontally varying surface pressure.

The horizontal momentum equations are derived from Newton's second law of motion. After the above-mentioned simplifications are made, the momentum equations in a general coordinate system can be given in a vector expression:

\[
\frac{\partial \mathbf{U}}{\partial t} + \mathbf{U} \cdot \nabla \mathbf{U} + \frac{\partial \mathbf{U}}{\partial z} = f(k \times \mathbf{U}) - \frac{1}{\rho_0} \nabla H_P + \kappa_M \frac{\partial^2 \mathbf{U}}{\partial z^2} + A_M \nabla^2 \mathbf{U} + \delta(z)(\mathbf{r}_a + F). \tag{3.2.1}
\]
where $U$ is the horizontal ocean velocity, $w$ the vertical ocean velocity, $\rho_0=\rho_w$ the reference water density and $p$ the pressure. The last term in Eq.(3.2.1) represents the surface momentum flux into the ocean calculated from the vector combination of air stress ($\tau_a$) and ice internal force ($F$). $K_M$ and $A_M$ are the vertical and horizontal eddy viscosities respectively. Here the turbulence processes are represented by conventional constant kinematic eddy viscosities. Considering that the top two ocean levels are quite thin (see Section 3.6), normally within the depth of the actual mixed layer, the two levels in the ice-ocean-only model are more or less bound together by allowing an increased vertical eddy viscosity and diffusivity in order to approximate the effects of the mixed layer. Here the value for the vertical viscosity in the top two levels is $K_M = 500 \text{ cm}^2/\text{s}$. While everywhere else, the vertical viscosity is $K_M = 1 \text{ cm}^2/\text{s}$. The horizontal viscosity, however, is uniquely chosen to be $A_M = 1.25 \times 10^8 \text{ cm}^2/\text{s}$ which is much smaller than that used in Hibler and Bryan (1987). Since higher resolution is used here, this smaller value of $A_M$ does not result in numerical instability. Note that this arrangement for viscosities is retained in the ice-ocean-mixed-layer model also.

The conservation equations for temperature and salinity are

$$\frac{\partial T}{\partial t} + \nabla_h (TU) + \frac{\partial}{\partial z} (Tw) = -\frac{\partial}{\partial z} \overline{(wT)} + A_h \nabla_h^2 T - R_e(T - T_0), \quad (3.2.2)$$

$$\frac{\partial S}{\partial t} + \nabla_h (SU) + \frac{\partial}{\partial z} (Sw) = -\frac{\partial}{\partial z} \overline{(wS)} + A_h \nabla_h^2 S - R_e(S - S_0). \quad (3.2.3)$$

In Eqs. (3.2.2) and (3.2.3), $T$ and $S$ are temperature and salinity respectively, $\overline{(wT)}$ and $\overline{(wS)}$ are the vertical turbulent fluxes of heat and salt, and $A_h$ is the horizontal diffusion coefficient (also uniquely chosen to be $A_h = 1.5 \times 10^6 \text{ cm}^2/\text{s}$ for both the ice-ocean-only model and the ice-ocean-mixed-layer model). The last terms in Eqs. (3.2.2) and (3.2.3) are so called diagnostic terms. In these two terms, $T_0$ and $S_0$ are the climatological
temperature and salinity from Levitus (1982), and \( R_t \) is the relaxation constant taken as 0.2/year used for a robust diagnostic ocean simulation and zero for a fully prognostic simulation. The inclusion of these two source terms will always place the ocean under the influence of observation, via the climatological temperature and salinity. How much the influence is, however, depends on how large \( R_t \) is. The value used here, 0.2/year, is rather small and represents a weak relaxation to the observation allowing interannual variability to be simulated. However, \( R_t \) is always set to zero in the first two levels of both the diagnostic model and the prognostic model so that the ocean's surface layer or mixed layer will not be subject to the restraints of observation and free interaction between ocean surface layer and ice or air is ensured.

The continuity equation is:

\[
\nabla \cdot \mathbf{U} + \frac{\partial w}{\partial z} = 0; \quad (3.2.4)
\]

The hydrostatic equation is made by simplifying vertical momentum equation to be

\[
\frac{\partial p}{\partial z} = -\rho z, \quad \text{or} \quad p = p_s + \int g\rho \mathrm{d}z',
\]

where \( \rho \) is sea water density determined by an equation of state of the form:

\[
\rho = \rho(T,S,p). \quad (3.2.6)
\]

The Bryan and Cox (1972) formulation for the equation of state is used.

For the ice-ocean-only model, the formulation for the vertical turbulent flux terms in Eqs (3.2.2) and (3.2.3), \( \langle w'T \rangle \) and \( \langle w'S \rangle \), is conventionally parameterized everywhere by eddy diffusivity such that

\[
-\langle w'T \rangle = K_H \frac{\partial T}{\partial z}, \quad -\langle w'S \rangle = K_H \frac{\partial S}{\partial z}. \quad (3.2.7)
\]
where the vertical diffusivity coefficient $K_H$ is more or less constant. This coefficient is also of a relatively larger value, $K_H = 5 \text{ cm}^2/\text{s}$, in the top two levels due to the same reason as that using larger vertical viscosity. Everywhere else, $K_H = 0.2 \text{ cm}^2/\text{s}$ is used.

For the ice-ocean-mixed-layer model, $(\bar{w'T})$ and $(\bar{w'S})$ are differently parameterized for the upper layer of the ocean and below. For the upper layer of the ocean, where the active turbulence, due to the stirring action of wind or ice and convective penetration, enhances vertical mixing, quasi-homogeneous temperature and salinity structures are often maintained down to the depth defined by sharp vertical tracer gradients (pycnocline) and a drastic decrease of the turbulence energy. Therefore, within the well mixed layer, these vertical flux terms are determined with the help of the mixed layer model. For the deeper layers below the pycnocline where the turbulence level is low due to the insulation from the surface energy sources, the classical parameterization of these terms, Eq.(3.2.7), remains to be used. Specifically, at the top and bottom of every level below the mixed layer base, Eq.(3.2.7) is used, with $K_H = 0.2 \text{ cm}^2/\text{s}$. While at the top and bottom of every level inside the mixed layer as well as at the mixed layer base, $(\bar{w'T})$ and $(\bar{w'S})$ are determined by the mixed layer model. At the sea surface, $(\bar{w'T})$ and $(\bar{w'S})$ are determined by surface flux conditions.

(ii) Boundary conditions

The sea surface stress comes from the interaction between the atmosphere and ocean or ice and ocean. For surface stress,

$$\rho_0 K_M \frac{\partial U}{\partial z} = (\tau_a + F)/\rho_w, \quad \text{at} \quad z = 0$$

(3.2.8)
For the ice-ocean-mixed-layer model the sea surface conditions for heat and salt fluxes are introduced into the mixed layer model which will be described in the following section. For the ice-ocean-only model, the heat and salt fluxes are introduced into the first level of the ocean model and are more rapidly distributed to the second level via an increased vertical viscosity and diffusivity in between the first level and the second one to approximate the effects of a mixed layer. At the ocean bottom, fluxes of momentum, heat, and salt are taken to be zero:

\[- \{ w'u', w'v' \} = 0 \quad \text{at} \quad z = -H \quad (3.2.9)\]

\[- \{ w'T', w'S' \} = 0 \]

where \( H = H(x,y) \) is ocean depth. The rigid lid condition is enforced at the ocean surface:

\[ w = 0, \quad \text{at} \quad z = 0 \quad (3.2.10) \]

The parallel flow condition is assumed at the ocean bottom:

\[ w = -u \frac{\partial H}{\partial x} - v \frac{\partial H}{\partial y} \quad z = -H(x,y) \quad (3.2.11) \]

At ocean's lateral walls, a non-slip condition \((u, v = 0)\) is used, and no flux of salt or heat is allowed. At open boundaries, the stream function is specified with geostrophic velocity calculated as an adjustment to the mean velocity. The temperature and salinity, at and near open boundaries, are strongly relaxed at every level to the climatological average to keep them from drifting too far from observations. Furthermore, the heat and salt gradients are set to be zero across the open boundaries while the deviation is compensated by the geostrophic velocity adjustment. The specifications of open
boundaries will be given in Section 3.5, while detailed description about the numerical treatment of open boundaries is referred to Appendix 1.

3.3 Mixed Layer Formulation

For the ice-ocean-mixed-layer model, the mixed layer model serves as an important connection between the ice model and the ocean. The Kraus and Turner mixed layer model, adopted in this research, was systematically presented in Niiler and Kraus (1977) and additional information can be found in Houssais (1988) and Lemke and Manley (1984). For the one-dimensional mixed layer model, the conservation equations for temperature and salinity are:

\[
\frac{\partial T}{\partial t} + \frac{\partial}{\partial z} \left( wT \right) = - \frac{1}{\rho_w c} \frac{\partial I_0}{\partial z}, \tag{3.3.1}
\]

\[
\frac{\partial S}{\partial t} + \frac{\partial}{\partial z} \left( wS \right) = 0. \tag{3.3.2}
\]

where, \( \rho_w \) is sea water density, \( c \) is sea water heat capacity, \( I = I_0 \exp(rz) \) is the penetrating component of solar radiation in the open ocean, and \( r^{-1} \) is the e-folding scale. Integration over the mixed layer depth gives:

\[
h_m \frac{\partial T_m}{\partial t} = - (w'T)_0 + (w'T)_h - \int_{h_m}^0 r I_0 \exp(rz) dz, \tag{3.3.3}
\]

\[
h_m \frac{\partial S_m}{\partial t} = - (w'S)_0 + (w'S)_h. \tag{3.3.4}
\]

where \( T_m \) and \( S_m \) are temperature and salinity in the mixed layer respectively, \( T(-h_m) \) and \( S(-h_m) \) are temperature and salinity immediately below the mixed layer base, and \( h_m \) is
mixed layer depth. It is known that the combined effects of the solar penetration and of wind stirring or ice stirring are important in preventing the mixed layer from becoming extremely shallow. For penetrating solar radiation, however, it is difficult to determine its behavior through ice of variable thicknesses, therefore one way to parameterizing its effect is to require that \( h_m \geq h_{\text{min}} \) (Schopf and Cane, 1983). In the mixed layer model, \( h_{\text{min}} \) is chosen to be as deep as the first ocean level thickness, 10 m, which may be reasonable because the shallow mixed layer depth is usually around 10-20 m in the Arctic as cited before.

Surface heat and salt fluxes are from the forcings at the atmosphere - ocean or ice - ocean interfaces. The total surface heat and salt fluxes in a differential time \( dt \) can be written as

\[
- (w' T')_0 \, dt = \max \{0, \left[ (1 - A)F_{\text{AO}} + AF_{\text{AI}} \right] dt - hQ_l/\rho_w c \}, \quad \text{for} \quad h > 0; \quad (3.3.5a)
\]

\[
- (w' T')_0 \, dt = \left[ (1 - A)F_{\text{AO}} + AF_{\text{AI}} \right] dt/\rho_w c, \quad \text{for} \quad h = 0. \quad (3.3.5b)
\]

\[
- (w' S')_0 \, dt = (1 - A)S_m(E - P)dt + S'_h(h,A)dt(S_m - S_l)\rho_l/\rho_w. \quad (3.3.6)
\]

where \( F_{\text{AO}} \) and \( F_{\text{AI}} \) are the net atmospheric radiative and heat fluxes reaching the open ocean surface, or thin ice, and thick ice respectively which are calculated using the method described in detail by Hibler (1980). In these equations, \( h \) is ice thickness, \( \rho_l \) is ice density, \( Q_l \) is volumetric heat of fusion of sea ice, \( P \) is precipitation, \( E \) is the evaporation at the surface; \( S_l \) is the salinity of ice taken here to be zero and \( S'_h \) is the actual ice growth rate which includes the effects of oceanic heat flux in the area covered by ice. The term \( S'_h \, dt \) can be written as
\begin{equation}
S_h \, \text{dt} = -\min\{h, \left( (1 - A)F_{Ao} + AF_{Ai} \right) dt + h_m(T_m - Tf)\rho_w c/Qt \} 
\tag{3.3.7}
\end{equation}

where \( T_f \) is the sea ice freezing point, \( h_m \) is the predicted variable mixed layer depth in the ice-ocean-mixed-layer model and the fixed first ocean level thickness in the ice-ocean-only model. It is assumed that the mixed layer temperature \( T_m \) at a grid cell is always at the freezing point whenever ice exists in that cell, and it would be above freezing when the grid cell is ice free and heat is added to the mixed layer that serves as a heat reservoir. Note that the fresh water input due to river runoff is not expressed in Eq.(3.3.6). The way of including fresh water input will be discussed in Section 3.5.

The mixed layer bottom fluxes are expressed as follows:

\begin{equation}
(w'T)_{h_m} = w_e(T - h_m - T_m) \tag{3.3.8}
\end{equation}

\begin{equation}
(w'S)_{h_m} = w_e(S - h_m - S_m) \tag{3.3.9}
\end{equation}

The complete expression for the above entrainment velocity \( w_e \) is:

\begin{equation}
w_e(q^2 + \Delta b h_m - s l \Delta u^2) = 2mu^3 + 0.5h_m((1 + n)(b'w')_0 - (1 - n)(b'w')_0) + \\
\begin{array}{lllll}
(A) & (B) & (C) & (D) & (E) \\
\end{array}
\end{equation}

\begin{equation}
\frac{g\alpha}{\rho_w c} I_s \left[ h_m \left[ 1 + \exp(- rh_m) \right] - (2/r) \left[ 1 - \exp(- rh_m) \right] \right] \tag{3.3.10}
\end{equation}

where \( b = -g(\rho - \rho_r)/\rho_r = g[\alpha(T - T_r) - \beta(S - S_r)] \) is the buoyancy with \( \rho_r, T_r \) and \( S_r \) being reference density, temperature and salinity respectively, \( \alpha = -1/\rho(\partial \rho/\partial T) \) the thermal expansion coefficient, and \( \beta = 1/\rho(\partial \rho/\partial S) \) the expansion coefficient for salinity; \( \Delta b \) and \( \Delta u \)
are the buoyancy and velocity jumps at the mixed layer base; \( u^* = (\tau_a + F/\rho)^{1/2} \) is the friction velocity; \( \langle w'b' \rangle_0 = g[\alpha(w'T)_0 - \beta(w'S')_0] \) is the surface buoyancy flux; \( s, m \) and \( n \) are coefficients of decay of the energy production; and \( q^2 \) is the rate of energy needed to agitate the entrained water.

In the entrainment regime the underlying otherwise undisturbed water is agitated and entrained into the mixed layer so that the mixed layer depth has a tendency to increase, which is represented by \( w_e > 0 \). When there is inadequate vertical turbulence energy to reach the base of the existing mixed layer, no entrainment would occur and the existing mixed layer can not maintain itself and tends to retreat to a new and shallower position if water does not converge locally. This phenomenon is sometimes called detrainment with \( w_e \leq 0 \) (Cushman-Roisin, 1987).

For a one-dimensional mixed layer model, Eq.(3.3.11) can be used to determine the evolution of mixed layer depth. For entrainment, \( w_e > 0 \), the mixed layer depth can be predicted by

\[
\frac{dh_m}{dt} = w_e, \quad (3.3.11)
\]

which we calculated from Eq.(3.3.11). For detrainment, \( w_e \leq 0 \), the new equilibrium mixed layer depth is determined diagnostically by solving the following equation:

\[
2mu^*^3 + 0.5h_m[(1 + n)(\langle b'w' \rangle_0 - (1 - n)\langle b'w' \rangle_0)] \\
+ \frac{g \gamma}{\rho_w c_s} h_m [1 + \exp(- rh_m)] - (2/r)[1 - \exp(- rh_m)] = 0 \quad (3.3.12)
\]

For a three-dimensional mixed layer model, the horizontal effects such as advection have to be taken into consideration for the determination of the actual mixed layer depth. In order to take the contribution of advection into account within the mixed layer, a
mass balance is made over a stationary volume element $h_m \Delta x \Delta y$ and a relationship can be obtained among the mixed layer thickness, the advection term and the entrainment velocity (Cushman-Roisin, 1987):

$$ \frac{\partial h_m}{\partial t} = w_e - w(-h_m) = w_e - \left[ \frac{\partial}{\partial x}(u_m h_m) + \frac{\partial}{\partial y}(v_m h_m) \right] $$ (3.3.13)

where $w(-h_m) = \frac{\partial}{\partial x}(u_m h_m) + \frac{\partial}{\partial y}(v_m h_m)$ can also be derived from the continuity equation by integrating it vertically over the mixed layer thickness, $u_m$ and $v_m$ are the horizontal velocity components in the mixed layer, and $w$ is upwelling velocity. It is obvious that the mixed layer does not necessarily become thicker in entrainment and, on the other hand, it does not necessarily become shallower in detrainment because of the existence of the advection term which can be determined by the ocean model.

In the Kraus and Turner mixed layer model, Eq.(3.3.10) plays a central role in that it has to be used to determine the shape of the mixed layer, in conjunction with the advection term, during both entrainment and detrainment. Eq.(3.3.10) looks cumbersome and has several parameters that have to be set before putting the equation into use. Therefore, a discussion about the equation and its application to the present ice-ocean modeling is needed. Although the significance of those terms in Eq.(3.3.10) have been pointed out in Niiler and Kraus, it may be useful to cite their meaning here.

- Term (A) in Eq.(3.3.10) is the rate of energy needed to agitate the entrained water;
- Term (B) is the work per unit time needed to lift the dense entrained water and to mix it through the layer;
- Term (C) is the rate at which energy of the mean velocity field is reduced by mixing across the layer base;
- Term (D) is the rate of working by the wind or ice;
• Term (E) is the rate of potential energy change, associated with buoyancy change, produced by fluxes across the sea surface. This term can result in either agitating underlying water due to penetrating convection or reducing the turbulence level due to buoyancy increase;
• Term (F) is the rate of potential energy change produced by penetrating solar radiation.

Term (D) parameterizes the mechanical energy input to the mixed layer due to wind or ice induced stress and is always a source term of turbulence that tends to deepen the mixed layer. In most of the ice covered ocean, the ice cover generally reduces the intensity of ocean surface stress by preventing wind from directly acting on the ocean surface. The mechanical energy is mainly induced by the ice keel stirring which is usually small. In ice free region such as part of the GIN Sea where the wind is frequently strong, however, the wind working rate can be a dominant factor. (E), on the other hand, is assumed to play a dominant role in both entrainment and detrainment regimes in the ice covered area. In the ice covered ocean active ice freezing and melting result in positive or negative surface buoyancy flux which causes buoyancy change in the water column. The buoyancy change can result in either agitating the underlying water or reducing turbulence level. That is, (E) can, to a large extent, deepen the mixed layer by penetrating convection as well as make it shallower by stabilizing the water column. When positive, these two production terms, the kinetic energy production (D) and the potential energy production (E), are subject to dissipation in energy. Following Lemke and Manley (1984) the dissipation process is parameterized by introducing the following coefficients (Eq.(3.3.10)), that insure an exponential decay of the energy production as $h_m$ increases:

$$n = \exp\left(- \frac{h_m}{h_C}\right)$$
$$m = \exp\left(- \frac{h_m}{h_W}\right)$$
$$s = \exp\left(- \frac{h_m}{h_W}\right)$$

(3.3.14)
where coefficients $h_c$ and $h_w$, controlling the rate of dissipation, cannot be determined theoretically and have to be chosen empirically. A wide range of values for these constants have been seen in a number of studies (Lemke and Manley, 1984; Lemke, 1987 and Houssais, 1988). Particularly for the Arctic Basin, Lemke and Manley tested a number of values for $h_w$ and $h_c$, ranging from about 9 m to less than 30 m, while for the Greenland Sea, Houssais used $h_w = 20$ m and $h_c = 100$ m. Obviously, there are no universal decay coefficients that would fit each of the regions under consideration because these regions are very different in environment. It would be a tedious effort to obtain individually useful coefficients based on a series of tests in every individual region. For simplicity, therefore, $h_c$ and $h_w$ are uniquely expressed in this research as linear functions of $h_m$ regardless of region:

$$h_w = (h_m - 10) / 10 + 4, \quad h_c = (h_m - 10) / 7.5 + 4.$$  \hspace{2cm} (3.3.15)

Compared to the corresponding constant coefficients these coefficients will dissipate a little more energy when $h_m$ is relatively small and a little less when $h_m$ becomes large.

Term (C) is generally small for a convective mixed layer penetrating deep into the underlying weakly stratified water mass (Houssais, 1988) and, hence, is neglected. Also dropped is term (F) for it is also small in the ice covered areas. As for term (A), it is determined by setting $q^2 = 9 \times \max(10^{-4} \text{ m}^2/\text{s}^2, u^*^2)$ according to Kim (1976). Term (B) includes a buoyancy jump $\Delta b$. This buoyancy jump can be obtained by calculating the temperature and salinity differences across the mixed layer base which is approximated as a step-like profile of the vertical structure within the pycnocline. According to Lemke (1987), there is a correction to the buoyancy jump for an ice covered ocean in taking into account the effect of ocean heat flux so that the complete buoyancy jump is expressed as:

$$\Delta b' = \Delta b + B',$$

$$B' = g[b \rho c T(S_m - S_i)/L_1 - \alpha](T(- h_m) - T_m).$$  \hspace{2cm} (3.3.16)
where $L_I$ is the latent heat of melting of ice. Since $B'$ is normally positive, it leads to a small reduction of entrainment rate.

3.4 Coupling of the Mixed Layer Model with the Ocean Model

The technique of embedding the mixed layer model into the ocean model is based on that of Adamec et al (1981) and Resnyanskiy and Zelen'ko (1991) with some modifications. The basic objective for the embedding procedure is to couple the vertical mixing simulated by the two-level mixed layer model with the advective and diffusive processes calculated in the multi-level primitive equation ocean model. As described before, the embedded mixed layer model should be able to predict, in conjunction with the advective changes supplied by the ocean model, the depth of the mixed layer, calculate the temperature and salinity jumps at the base, and mix the related quantities to make the mixed layer vertically quasi-uniform (well mixed). Since the depth of the mixed layer does not necessarily coincide with any of the fixed model levels used in the primitive equations calculations, work has to be done to insure that the incorporation of the step-like vertical profile of the mixed layer density will not lose the conservation of heat, salt and momentum both in the mixed layer model and the ocean model. When the mixed layer reforms at a shallower depth due to detrainment, Adamec et al's technique would require an additional potential energy conservation to adjust the shallower layer to fit the fixed-level structure.

The coupling procedure is outlined below:

(i) Calculate the ocean circulation with the ocean model, that is, solve Eq.s (3.2.1) - (3.2.5) with the help of (3.2.6) - (3.2.11) to obtain the $u$, $v$, $T$ & $S$ due to advective and diffusive processes at every ocean model's level.
(ii) The contribution of the advective changes to the mixed layer depth is obtained by calculating $u_m$ and $v_m$ (Eq.(3.3.13)) using the velocity profile solved at step (i) so that a preliminary mixed layer depth is obtained by solving:

$$\frac{\partial h_m}{\partial t} = - \left[ \frac{\partial}{\partial x}(u_m h_m) + \frac{\partial}{\partial y}(v_m h_m) \right].$$

(iii) The temperatures and salinities in the mixed layer, $T_m$, $T(-h_m)$, $S_m$ and $S(-h_m)$, after advective and diffusive calculations have to be adjusted using the newly obtained $T$ and $S$ profiles at step (i) (Resnyanskiy and Zelen'ko, 1991; Adamec, et al, 1981):

$$T_m = \sum_{i=1}^{k} T_i \Delta z_i - (z_k - h_m)T_k/h_m, \quad S_m = \sum_{i=1}^{k} S_i \Delta z_i - (z_k - h_m)S_k/h_m,$$

$$T(-h_m) = T_k, \quad S(-h_m) = S_k. \quad (3.4.1)$$

where level $k$ is containing the mixed layer base as shown in Fig.(3.4.1), $T_l$ and $S_l$ are temperature and salinity at level $l$ in the ocean model. The $T$ and $S$ profiles obtained from the ocean model are to be corrected with the above quantities in the mixed layer:

$$T_l = T_m, \quad S_l = S_m, \quad \text{for } l = 1, 2, ..., k-1. \quad (3.4.3)$$

In order to ensure the conservation of the integral quantities in the water column, such values are corrected in the $k$-th level taking into account the actual fractions occupied in this layer by mixed layer water and underlying water:

$$T_k = \frac{[(z_k - h_m)T(-h_m) + (h_m - z_k-1)T_m] \Delta z_k}{\Delta z_k},$$

$$S_k = \frac{[(z_k - h_m)S(-h_m) + (h_m - z_k-1)S_m] \Delta z_k}{\Delta z_k}. \quad (3.4.4)$$
(iv) Computation of mixing process:

First of all the related quantities in Eq.(3.3.10) are calculated; The surface buoyancy flux and mechanical energy are supplied by both the ice and the ocean models (Eq.(3.2.8), (3.3.5) and (3.3.6)); The buoyancy jump $\Delta b$ is determined by $T_m$, $T(-h_m)$, $S_m$ and $S(-h_m)$ from step (iii).

If $w_e > 0$, the new mixed layer with an increased depth is obtained using Eq.(3.3.11) based on the mixed layer depth after the advective adjustment. Then by solving Eq.(3.3.3) and (3.3.4) with the help of (3.3.5) - (3.3.8) and doing vertical mixing following the same procedure as step (iii), we can obtain new values for $T_m$ and $S_m$.

If $w_e \leq 0$, the mixed layer normally retreats to a shallower depth which can be determined by solving Eq.(3.3.12). In this case the embedding technique of Adamec et al requires an adjustment for $T_m$, $S_m$ in order to insure potential energy conservation in
addition to heat and salt conservation so that the new mixed layer and pycnocline temperatures (similarly for salinities) are given by (Fig.(3.4.2)):

\[
T'_m = T(- h_m) + \frac{(T_m - T(- h_m))h_m(D_k - h_m + D_m)}{D_k D_m},
\]

\[
T'(- h_m) = T(- h_m) + \frac{(T_m - T(- h_m))h_m(h_m - D_m)}{D_k \Delta Z_k}
\]

where

\[
D_k = \sum_{l=1}^{k} \Delta Z_l, \quad D_m = \max(h_m', D_k - \Delta Z_k).
\]

In Eq.(3.4.5), \( T'_m, T'(- h_m) \) and \( h'_m \) are the new mixed layer temperature, mixed layer base temperature and mixed layer depth respectively.

---

**Fig. (3.4.2)** Illustration of mixed layer shallowing. The solid line is the temperature profile before shallowing and the dashed line is the profile after shallowing.
Based on potential energy conservation, Eq. (3.4.5) has a firm foundation in a physical sense, and it may work well in reforming a new shallower mixed layer in the equatorial or temperate oceans where the upper ocean temperature is normally well above freezing point. However, when the new mixed layer becomes much shallower than before, say, compared to a previous time step, a problem arises with the use of Eq. (3.4.5) for ice covered oceans or the areas that have a mixed layer temperature close to freezing. That is, Eq. (3.4.5) may result in such a temperature decrease that the mixed layer temperature falls below freezing, causing phase change and possible formation of a large volume of ice that is not realistic. This situation may also create numerical instability. Another problem with the implementation of Eq. (3.4.5) is that when the mixed layer base is slightly above a fixed-level's base in the ocean model, it seldom makes any difference. In order to avoid these problems, Houssais and Hibler (1993) dropped the requirement of potential energy conservation, with only heat and salt conservation retained, which seems to work well in the subarctic ocean.

In this research, however, a slightly different approach is adopted. Eq. (3.4.4) is still used but every time the diagnostic mixed layer depth is obtained from solving Eq. (3.3.12), the average of the depth and the one of the previous time step is used as the new depth. This was found to be able to greatly limit the possibly dramatic temperature decrease in the mixed layer while only delaying the mixed layer approaching its shallowest position by a few time steps.

Once the new $h_m$ is calculated, all related integral quantities are solved using Eq.s (3.3.3) - (3.3.6), (3.3.8) and (3.3.9). And as in the case of $w_e > 0$ the mixing procedure of step (iii) is followed again for profile matching between the ocean model's multi-levels and the mixed layer model's two levels.
For vertical velocity profiles calculated from the ocean model, one can choose either to keep them unchanged or to mix them over the mixed layer thickness the same as the temperature and salinity profiles are adjusted in step (iii). If the latter is chosen, the following group of equations is used:

\[
\begin{align*}
    u_m &= \left[ \sum_{l=1}^{k} u_l \Delta z_l - (z_k - h_m)u_k \right]/h_m, \\
    v_m &= \left[ \sum_{l=1}^{k} v_l \Delta z_l - (z_k - h_m)v_k \right]/h_m, \\
    u(-h_m) &= u_k, \\
    v(-h_m) &= v_k, \\
    u_l &= u_m, \quad v_l = v_m, \quad l = 1, 2, \ldots, k-1 \\
    u_k &= \left[ (z_k - h_m)u(-h_m) + (h_m - z_{k-1})u_m \right]/\Delta z_k, \\
    v_k &= \left[ (z_k - h_m)v(-h_m) + (h_m - z_{k-1})v_m \right]/\Delta z_k.
\end{align*}
\]

However, the standard ice-ocean-mixed-layer model used in this study does not mix the velocity within the mixed layer for simplicity. This is because it is felt that vertical mixing of temperature and salinity is more important in modifying the ocean's surface density structure and ice conditions.

In addition to the above-mentioned mixing calculations carried out in the mixed layer model, there are still instantaneous overturning calculations at any level of the ocean model where hydrostatic instability occurs. This convective adjustment procedure is done in the same manner as in the ice-ocean model without variable depth mixed layer.

In order to examine the behavior of the mixed layer model and the embedding procedure, a simple one-dimensional salinity mixed layer model is constructed and tested (see Appendix 2 for detail). This salinity mixed layer model is driven by periodic seasonal ice melting and constant wind (following Lemke and Manley, 1984). It is found that the model behaves well in that cyclic change of the mixed layer salinity over many years,
resulting from the periodic forcing, is generated and the total salt is always conserved. Therefore long-term simulations are possible using such a mixed layer model and embedding technique. Also the variations in depth are commensurate with observations.

3.5 Coupling of the Ice Model and the Ocean Model

A successful numerical ice-ocean model should be able to realistically simulate the complex process of ice-sea interaction. The modeling of the surface fluxes of heat, salt and momentum is important in properly determining the interaction between the ice and ocean models and correctly coupling them together. The surface fluxes are closely related to the atmospheric forcing, consisting of geostrophic surface winds, surface air temperature, humidity and shortwave and longwave radiations. The atmospheric forcing is pre-calculated (see section 3.7) and specified in this model which will act on the ocean surface either directly, if there is no ice, or indirectly with ice effects included. The feedback effects of ice cover on most of the atmospheric properties are not considered except that the surface air temperature is subject to a modification due to the influence of ice (section 3.7).

The atmospheric forcing is used by the ice model where all the surface fluxes are calculated and then transmitted into the ocean model. Momentum flux is transferred as a stress acting on the ocean surface which consists of wind induced stress modified by an ice interaction force (Eq.(3.2.1)). If there is no ice existing, the ice interaction is zero and the stress acting on the water is the wind stress only. On the other hand, the geostrophic velocity of ocean is used, together with ice velocity, to determine water stress acting on the ice according to a quadratic drag law (Eq.(3.1.3)). In the model, the velocity at the third level in the ocean model is taken to approximate the geostrophic ocean velocity.
Atmospheric thermodynamic forcing is used in the ice model for heat budget calculations to determine the amount of heat transferred at the air-ice or air-ocean interfaces. If no ice exists, the downward heat transfer calculated in the ice model will be directly put into the ocean to increase the temperature in the mixed layer while loss of heat to the atmosphere results in temperature decrease in the mixed layer and ice growth when the temperature drops to the freezing point (-1.96°C). If ice exists, the mixed layer temperature is always maintained at the freezing point due to ice growth or melting. The atmospheric heat flux acts only on the ice and no heat is transferred into the ocean. However, if the downward heat flux is large enough to melt the ice at one timestep, the surplus of the heat will be forwarded into the ocean.

The ice growth or decay is also affected by the oceanic heat flux calculated in the ocean model, in conjunction with the mixed layer model if incorporated. In the coupled model the oceanic heat flux in a time step is computed from the following equation:

\[ F_{w \Delta t} = (T_m - T_f) \rho_w c_h m / Q_i \]  

(3.5.1)

where the unit of \( F_{w \Delta t} \) is in 'meters of ice' that is the amount of heat needed to melt a meter of ice, \( T_m \) and \( h_m \) are either the mixed layer temperature and depth for the ice-ocean-mixed-layer model or the first-level temperature and thickness for the ice-ocean-only model. As described before, \( T_m \) is obtained by simulating the advective, diffusive and turbulent mixing processes in the ocean and mixed layer models. Once \( T_m \) is found to be above the freezing point, the positive heat flux will be used to melt a corresponding portion of ice if enough ice exists and \( T_m \) will be corrected to be at the freezing point again. If there is not enough ice to be melted by the oceanic heat flux, the surplus of the heat flux will be stored in the mixed layer, just like the surplus of atmospheric heat, resulting in a \( T_m \) above the freezing point. In contrast to this, if \( T_m \) is below the freezing point, negative \( F_w \) will
cause ice growth, and the released heat of fusion of ice will bring $T_m$ up to the freezing point again.

The surface salt flux is determined by net precipitation (actual precipitation minus evaporation) which is specified in the model, and the amount of ice that actually grows or melts (Eq.(3.3.6)). Eq.(3.3.6) does not include the river runoff which brings in significant fresh water into the ocean like net precipitation. The contribution of the rivers' fresh water is evenly distributed in the grid cells where the rivers outlet is located. Fresh water is treated as negative salt flux of 35 ppt. Actually in the model the net precipitation is treated in the same way, this means that the $S_m$'s at the right hand side of Eq.(3.3.6) are also of a value of 35 ppt instead of the value of the mixed layer salinity. This procedure, adopted by Hibler and Bryan (1987), is used to insure global salt conservation.

3.6 Numerical Framework

The model ice-ocean system for the Arctic Ocean and Barents and GIN Seas is discretized and solved using finite difference techniques. The model's grid configuration is shown in Fig.(3.6.1). The model has a 40 km x 40 km horizontal resolution with grid size of 130 x 102 in rectangular coordinates. In creating this model grid, the Lambert equivalent projection is used which conserves area. The ocean model's corresponding rectangular geometry is converted from a spherical coordinate system whose equator is aligned with 35° E and passes through the north pole to obtain most uniform space differencing over the Arctic Ocean. The ocean's vertical dimension has 21 levels of different thicknesses shown in Table (3.6.1). As listed in Table (3.6.1), the first level thickness is as thin as 10 m, and the remaining level thicknesses gradually increase with depth.
Open boundary conditions are specified across the Bering and Denmark Straits and Faroe-Shetland Passage where the open boundary cells are marked in Fig.(3.6.1) The hatched areas adjacent to the open boundaries are strongly relaxed to Levitus (1982) climatological data, as a part of open-boundary treatment (see section 3.8 and Appendix 1). In Fig.(3.6.1), some blackened grid cells are shown, which are small islands. The grid is divided into 3 geographical areas, Arctic Basin, Barents Sea and Greenland Sea (more often called GIN Sea), as shown in the figure, for the purpose of result analyses.

Such a high horizontal as well as vertical resolution adopted in the model can very well resolve the coastline and the bottom topography with many ridges that separate different basins in the whole region. The narrow passages like the Fram Strait and Bering Strait can also be well represented with a reasonable number of grid cells across them. Another advantage is that higher resolution makes it possible to reduce the horizontal viscosity and diffusivity, and still maintain numerical stability so that more robust ocean circulation may be generated which can enhance water transport and ice drift in general. Furthermore the high vertical resolution in the upper ocean is beneficial to the mixed layer modeling.

The model ocean's bottom topography is resolved by the 21 vertical levels, using differing numbers of levels at different locations. Fig.(3.6.2) shows the model bottom topography contours represented by the 21-level ice-ocean model. The bottom topography was calculated from the ETOPO5N data base from the National Geophysical Data Center. This data base consists of 5-minute gridded world-wide bathymetry and topography. The bathymetry was interpolated using weighted averages to an average depth for a 5 km x 5 km square grid centered at the North Pole. The bathymetry for the 5 km grid was then averaged to an 40 km x 40 km grid with the requirement that at least 75% of the 5 km grid cells were ocean cells (Ranelli, 1991). The bathymetry in each 40 km grid cell was then
converted to an equivalent number of levels, with limited smoothing using 9 point Hanning, for the ocean model. For very shallow areas, 4 levels are specified. Fig.(3.6.3) shows an equivalent bottom topography represented by numbers of levels.

Table (3.6.1). Vertical level thicknesses (in meters) for the ocean model

<table>
<thead>
<tr>
<th>Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10.00</td>
</tr>
<tr>
<td>15.43</td>
</tr>
<tr>
<td>22.50</td>
</tr>
<tr>
<td>31.47</td>
</tr>
<tr>
<td>42.83</td>
</tr>
<tr>
<td>56.83</td>
</tr>
<tr>
<td>74.00</td>
</tr>
<tr>
<td>94.50</td>
</tr>
<tr>
<td>118.7</td>
</tr>
<tr>
<td>214.9</td>
</tr>
<tr>
<td>254.3</td>
</tr>
<tr>
<td>296.9</td>
</tr>
<tr>
<td>345.8</td>
</tr>
<tr>
<td>405.8</td>
</tr>
<tr>
<td>469.2</td>
</tr>
<tr>
<td>537.5</td>
</tr>
<tr>
<td>613.3</td>
</tr>
<tr>
<td>694.2</td>
</tr>
<tr>
<td>792.5</td>
</tr>
</tbody>
</table>

The grid is an Arakawa 'B' grid type (Mesinger and Arakawa, 1976) with spatially staggered variables both for the ice and ocean models (Bryan, 1969; Hibler, 1979). The differencing schemes are designed for global conservation of mass, heat, salt and total energy in the ocean model and the mass and momentum in the ice model. A one-quarter day time step is used in the integration of ocean conservation equations while 1/80 day time step is used for the ocean momentum equations. This is so called "distorted physics" (Bryan, 1984) in which the length of the time step for the conservation equations is 20 times the length as the time step for the momentum equations (Ranelli, 1991). The one-quarter day time step is also used for ice conservation equations as well as momentum equations. The finite-difference equations are solved using a leapfrog timestep. However, a modified Euler time step is used after every 17 leapfrog time steps to eliminate one branch
of the split solution caused by leapfrog differencing. For the mixed layer model, the forward differencing scheme is used with the same one-quarter-day time step.

Given the fact that the Hibler (1979) numerical approach for solving ice velocity in such a high resolution model takes rather long CPU time, an efficient numerical method by Zhang and Hibler (1993) is used. Based on a semi-implicit treatment and a line successive relaxation technique, this method greatly speeds up the ice modeling and, hence makes possible long-term fine-resolution simulations with our existing computer facility. Furthermore, this new method seems to have a better behavior for solutions close to the plastic regime (see Zhang & Hibler, 1993).

3.7 Atmospheric Forcing Fields

Atmospheric forcing fields driving the model are daily varying geostrophic winds, surface air temperature, specific humidity, and shortwave and longwave radiation over a period of 7 years (1979-85). The geostrophic winds are calculated from daily surface atmospheric pressures. The pressure data were interpolated from 5° by 5° latitude-longitude pressure fields provided by John Walsh, which were obtained by merging the Arctic data-buoy pressure analysis with NCAR daily surface analysis. Also provided by John Walsh were Hansen's monthly mean surface air temperature fields, interpolated into daily temperature, of the same resolution as the pressure data. The Hansen (Hansen and Lebedeff, 1987) temperature fields were modified to take into account the effects of ice feedback on air temperature. This modification is thought reasonable from the observation that the air temperature over the ice in summer rarely goes above freezing due to the interaction between the ice and the air. This observation is inconsistent with the Hansen temperature data which show very warm temperature over regions with significant observed ice concentrations in summer because of the use of only a few data points over
the sea ice to anchor the temperature analysis. To ameliorate this problem, the Hansen temperatures are modified in such a way that they are never allowed to go above 0°C if at least 25% ice concentration exists (see Hibler and Zhang, 1993 for more information). This procedure basically modifies the Hansen summer temperature fields in marginal ice zone. The specific humidity and longwave and shortwave radiation fields were calculated using the formulation of Parkinson and Washington (1979) based on the surface air pressure and temperature fields, and a constant relative humidity of 90%.

3.8 River Runoff and Precipitation

In the North Polar Region, the river runoff and precipitation are the major source of freshwater input specified as a negative salt flux of 35 ppt. A previous study by Ranelli and Hibler (1990) demonstrated the necessity for a realistic freshwater flux to maintain vertical density stratification in the ocean which is important to obtain a realistic ocean circulation. River runoff values are from Hibler and Bryan (1987) and are shown in Table (3.8.1). The river runoff is specified to occur mainly in the summer, between day 171 and day 291 (20 June and 18 Oct). The river runoff from any of the large rivers in the region under consideration is specified at the grid cell adjacent to the river mouth, while residual river runoff is uniformly distributed around the perimeter of the model grid.

<table>
<thead>
<tr>
<th>River</th>
<th>Runoff (10^3 m^3 s^-1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mackenzie</td>
<td>7.9</td>
</tr>
<tr>
<td>Kolyma</td>
<td>3.8</td>
</tr>
<tr>
<td>Lena</td>
<td>15.4</td>
</tr>
<tr>
<td>Yenisei</td>
<td>17.3</td>
</tr>
<tr>
<td>Ob</td>
<td>12.4</td>
</tr>
<tr>
<td>Pechora</td>
<td>4.1</td>
</tr>
<tr>
<td>Mezen</td>
<td>0.8</td>
</tr>
<tr>
<td>Dvina</td>
<td>3.5</td>
</tr>
</tbody>
</table>

Additional runoff around the perimeter of the analysis region: 74.3
In addition to the freshwater supply due to river runoff, observed monthly-varying region-dependent precipitation (Vowinkel and Orvig, 1970) is specified in the model. The precipitation values listed in Table (3.8.2) are net precipitation (actual precipitation minus evaporation). These values were assigned to the corresponding regions where they belong in a 80 km resolution grid of an ice-ocean model by Ranelli (1991). The rain/snow distribution in the 80 km resolution grid was interpolated into the 40 km resolution grid used here. The different methods of introducing precipitation into the model has been described in section 3.1.

<table>
<thead>
<tr>
<th>Area</th>
<th>Jan</th>
<th>Feb</th>
<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
<th>Aug</th>
<th>Sep</th>
<th>Oct</th>
<th>Nov</th>
<th>Dec</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cent polar Ocea</td>
<td>1.26</td>
<td>1.10</td>
<td>0.90</td>
<td>0.38</td>
<td>-0.23</td>
<td>0.067</td>
<td>1.72</td>
<td>0.80</td>
<td>0.15</td>
<td>1.26</td>
<td>0.90</td>
<td>0.90</td>
</tr>
<tr>
<td>Beaufort Sea</td>
<td>1.10</td>
<td>1.10</td>
<td>0.79</td>
<td>0.42</td>
<td>0.15</td>
<td>1.22</td>
<td>0.192</td>
<td>2.76</td>
<td>0.77</td>
<td>1.66</td>
<td>0.12</td>
<td>1.21</td>
</tr>
<tr>
<td>East Sibr. Sea</td>
<td>1.00</td>
<td>0.90</td>
<td>0.90</td>
<td>0.23</td>
<td>0.13</td>
<td>1.21</td>
<td>2.68</td>
<td>1.66</td>
<td>1.64</td>
<td>0.083</td>
<td>1.81</td>
<td>1.21</td>
</tr>
<tr>
<td>Kara-Laptev Sea</td>
<td>1.93</td>
<td>1.64</td>
<td>1.42</td>
<td>1.13</td>
<td>0.49</td>
<td>-0.49</td>
<td>2.48</td>
<td>3.05</td>
<td>0.82</td>
<td>0.40</td>
<td>1.76</td>
<td>1.81</td>
</tr>
<tr>
<td>Barents-GIN Sea</td>
<td>-0.017</td>
<td>-0.52</td>
<td>-0.19</td>
<td>-0.72</td>
<td>0.32</td>
<td>1.41</td>
<td>2.34</td>
<td>1.71</td>
<td>2.13</td>
<td>0.084</td>
<td>1.02</td>
<td>0.42</td>
</tr>
</tbody>
</table>

3.9 Boundary and Initial Conditions

Strictly speaking, the atmospheric forcing and river runoff and precipitation described in Sections 3.7 and 3.8 respectively are an indisputable part of boundary conditions (surface boundary conditions). The river runoff and precipitation serve as specified ocean surface salt flux, while the atmospheric forcing dictates the surface heat, salt and momentum transfer. For example, the geostrophic winds, which are a part of the atmospheric forcing, are used to compute wind stress (Eq.(3.1.2)) that acts on the surface of the ice cover or the surface of the ocean if ice free. This wind stress is also a part of total stress that acts on the ocean surface (Eq.(3.2.8)) if ice exists. The rest of the atmospheric forcing is thermodynamic forcing that is specified on the ice surface for ice thermodynamic processes in order to obtain the information about ice growth and decay. This information
will be used to determine the ocean surface heat and salt fluxes. Nevertheless, these surface boundary conditions are not put together into this section. Instead, they were detailed in the previous two sections for descriptive purposes. In this section emphasis is laid on the lateral open boundary conditions enforced in the model.

Besides the non-slip lateral conditions along the model ocean walls, open boundary conditions are used across the Bering Strait, Denmark Strait and Faroe-Shetland Passage where water transport is specified. Based on various observational reports, 1 Sv water inflow is specified at the Bering Strait, whereas 6 Sv inflow at the Faroe-Shetland Passage, and 7 Sv outflow at the Denmark Strait are specified. These transports are close to those used by Semtner (1976). Such a water-transport specification at the open boundaries will meet the requirements for global water balance. The water transport along the Bering Strait is uniformly distributed, while those along the Faroe-Shetland Passage and Denmark Strait are not. Across the Bering Strait there are 2 grid cells, while there are 11 and 20 grid cells across the Denmark Strait and Faroe-Shetland Passage respectively. At the Denmark Strait only an outflow condition is enforced. The small inflow at the west of Iceland is not considered, but the outflow at the Faroe-Shetland Passage due to the East Icelandic Current is. The outflow condition is extended to a few grid cells at the west of the Faroe-Shetland Passage to obtain the East Icelandic Current. Other grid cells at the passage are specified with inflow conditions with the strongest inflow enforced at the area a few grid cells away from the coast of Norway to ensure that the Norwegian Current is obtained. Detailed grid cell dependent water inflow or outflow values at these grid cells are listed in Table (3.9.1). As a matter of fact, listed in Tables (3.9.1) - (3.9.3) are vertically integrated volume transport streamfunction values. Because of the rigid-lid assumption, the streamfunction has to be computed in the ocean model in order to solve for ocean velocity (Semtner, 1987). Specifying the streamfunction values along the open boundaries is a crucial part of the open boundary treatment because they are equivalent to the corresponding mean inflow
and outflow values. The streamfunction values are simultaneously specified at the grid cells at the open boundaries and the grid cells at their immediate neighborhood to ensure that the inflow/outflow crosses the boundaries in their normal direction. The possible flows along the open boundaries are not considered. Elsewhere, the streamfunction values are specified in such a way that each individual coast line must have spatially constant streamfunction value to allow zero normal velocity but different values may be allowed for different continents and islands. The treatment of islands is developed in Semtner (1974).

In addition to specifying mean flow across the open boundaries, a geostrophic adjustment is used (see Appendix 1 for details) so that vertical velocity variations associated with local density variations are achieved. In the regions near and at these open boundaries (see Fig. (3.6.1)) strong relaxation to the observed temperature and salinity, $T_0$ and $S_0$, is used at every vertical level. The strong relaxation is of a damping constant (inverse of $R_t$ in Eq.s (3.2.2) and (3.2.3)) of 30 days which is effective enough to bring the drifting of the temperature and salinity under control so that numerical stability is ensured. Applying this procedure at outflow boundaries may create certain salt and heat balance problem in long term prognostic simulation not considered here. The open boundaries are allowed to have vanishing spatial derivatives of temperature and salinity along the normal direction of the boundaries. This constraint does not seem to have an impact on the modeling since the strong relaxation at and near the open boundaries has a strong damping effect to disturbances excited at the open boundaries.

The following fields are used as initial conditions of model simulations: zero ice and ocean velocities; constant ice temperature of 273°K; constant ice compactness of 1.0; constant ice thickness of 1.1 meter; ocean temperature and salinity using Levitus (1982) climatological temperature and salinity. These initial conditions were used for a seven-year spin-up with atmospheric forcing from 1979 to 85. This spin-up was carried out using a
diagnostic ice-ocean-only model. The output after the seven-year spin-up was then used as initial conditions for two main experiments: a simulation with a full diagnostic ice-ocean-only model and a simulation with a full diagnostic ice-ocean-mixed-layer model. In addition, a series of sensitivity studies were carried out around these two main simulations to identify the effects of different processes on the ice-ocean system. All the sensitivity studies are done using the same seven-year-spin-up output as initial conditions. Although both the full ice-ocean-only model and the full ice-ocean-mixed-layer model have been described in some detail through this chapter, further descriptions are needed for these sensitivity studies made possible by varying some conditions used in the full models, which will be given immediately before their simulation results are presented or compared in the following related chapters.

Table (3.9.1) Streamfunction ($\psi$) values (Sv) at grid cells along Bering Strait

<table>
<thead>
<tr>
<th>Cell</th>
<th>(1,82)</th>
<th>(1,83)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\psi$</td>
<td>0.0</td>
<td>-1.0</td>
</tr>
</tbody>
</table>

Table (3.9.2) Streamfunction ($\psi$) values (Sv) at grid cells along Denmark Strait

<table>
<thead>
<tr>
<th>Cell</th>
<th>(91,1)</th>
<th>(92,1)</th>
<th>(93,1)</th>
<th>(94,1)</th>
<th>(95,1)</th>
<th>(96,1)</th>
<th>(97,1)</th>
<th>(98,1)</th>
<th>(99,1)</th>
<th>(100,1)</th>
<th>(101,1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\psi$</td>
<td>0.0</td>
<td>-0.3</td>
<td>-0.7</td>
<td>-1.3</td>
<td>-2.3</td>
<td>-3.5</td>
<td>-4.7</td>
<td>-5.7</td>
<td>-6.3</td>
<td>-6.7</td>
<td>-7.0</td>
</tr>
</tbody>
</table>
Table (3.9.3) Streamfunction ($\psi$) values (Sv) at grid cells along Faroe-Shetland Passage

<table>
<thead>
<tr>
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Figure (3.6.1) Grid configuration of the 40 km resolution ice-ocean model used in the North Polar Oceans. Stronger relaxation and open boundary cells are marked in the grid. The grid is divided into three areas for analysis of results. The numbers on the horizontal axis and the vertical axis represent the i transects and j transects respectively.
Figure (3.6.2) Contour map of bottom topography represented in the 40 km resolution ice-ocean model.
Figure (3.6.3) Bottom topography represented by numbers of ocean levels in vertical columns of T and S points. Letters a, b, c, d, e, f, g, h, i, j and k represent 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, and 20 respectively, while "*" stands for land point. The grid size is 130 x 102.
Chapter 4. Experiment with Ice-Ocean-Only Model

The main experiment with the ice-ocean-only model was carried out with a simulation over a seven-year period (1979-85) after a seven-year spin-up. The model was a diagnostic model described in detail in the previous chapter. The simulation results from the main experiment are presented in this chapter and some of the following chapters. Along with the main simulation are some related sensitivity studies. These sensitivity studies are achieved by running the ice-ocean-only model over the same seven-year period with changed or artificially specified conditions in order to examine the effects of different processes. Unless stated otherwise, the initial conditions for these studies are from the same 7-year spin-up as used for the main experiment. However, the majority of the results presented in this chapter are from the main experiment, while the results from those sensitivity studies are mainly used for comparison.

In the following sections of this chapter the predicted ocean fields of velocity, temperature, salinity, pressure, oceanic heat flux as well as the vertically integrated mass transport streamfunction are presented and discussed. Also presented are the fields of ice thickness compactness and velocity. Whenever possible, observational results are used for comparison. Most of the model results in this chapter will reflect the long-term mean characteristics of the ice-ocean system. While the characteristics of the interannual variability of some aspects of this model will be dealt with in Chapter 6. The ocean characteristics of the model are described in Sections 4.1 to 4.8. Sections 4.9 to 4.10 present the simulated ice conditions.

4.1 Ocean Surface Velocity Fields
The simulated fields of ocean surface velocity, taken to be the velocity at the third level of the model at a depth of 37 m, are shown in Fig. (4.1.1). The 1983 velocity fields are chosen to be shown in this figure because they represent the simulated general ocean circulation pattern. Included in the figure are not only the annual mean field but also the monthly mean fields of February and September, typically representing winter and summer. As shown in this figure, the model gives a two-gyre ocean surface circulation pattern in the Arctic Ocean. One gyre is a clockwise circulation in the Beaufort Sea, the other an anti-clockwise one in the Eurasian Basin. These two gyres seem to be separated by the Lomonosov Ridge and roughly to be of the shapes of the two deep basins in the Arctic: the Canadian Basin and the Eurasian Basin. These two gyres are existing with slight seasonal and interannual variations in circulation intensity. Because of the existing Eurasian gyre, the transpolar stream is deflected somewhat compared to what is shown in Fig. (2.3.3), but it still converges to and flows out of the Fram Strait.

An interesting flow is exhibited just off the Alaskan coast with some of the water from the Bering Strait flowing down along the coast against the strong Beaufort gyre. It is particularly robust in September of 1983. Such a flow is very similar to what was observed at that segment of the coast as shown in Fig. (2.3.3). Interest is therefore increased since this flow is of concern to local offshore activities and was not simulated by various previous studies, even those with 80 km horizontal resolution and 15 vertical levels (Ries and Hibler, 1991 and Ranelli 1991). To distinguish this from other currents in the Arctic, it
will be referred as the Beaufort Sea undercurrent. A further analysis of this current is made in Section 4.4. Although the model obtains a realistic local current off the Alaskan coast, it does not predict very well some other local currents such as the current off the Siberian coast (Fig.(2.3.3)).

Absolute pressure distribution is a good representation of the currents since the flow is close to geostrophic in the interior of the ocean. So the pressure field is analyzed here in relation to the ocean surface velocity distribution. The pressure field is determined by using the integrated form of Eq.(3.2.5). It includes a computation of surface pressure $p_S$ and an integration of water weight down to 37 m. $p_S$ is not explicitly calculated in a rigid lid ocean model. However, the surface pressure gradient can be determined and then used to calculate the Laplacian of the pressure. This results in a Poisson equation which can be solved to obtain $p_S$ (Ranelli, 1991) and hence the total pressure. Fig.(4.1.2) includes contours of annual mean pressure fields at second and third levels. These pressure fields represent the geostrophic surface currents. It confirms those main features shown in Fig.(4.1.1) such as the two-gyre circulation pattern and those major currents. The illustrated two gyres coincide well with those shown in Fig.(4.1.1).

4.2 Water Transport Patterns

- The ocean surface velocity fields do not necessarily represent the whole water mass transport patterns in the North Polar Region. Vertically integrated flow fields are a better measure of the whole water mass. The integrated flow can be represented by contours of the integrated mass transport streamfunction, $\psi$. Fig.(4.2.1) shows monthly mean streamfunction contours for February and September of 1983, typically representing winter and summer. Similar to the surface circulation, the water transport patterns are of a rather strong anti-clockwise circulation in the Eurasian basin in addition to a clockwise one
in the Beaufort Sea. In the summer, the Eurasian circulation is even of more intensity. The West Spitsbergen Current can sometimes be observed in the integrated flow contours (Fig. (4.2.1b)). The Beaufort Sea undercurrent may transport water up to 0.5 Sv occasionally.

Also seen from Fig. (4.2.1) as well as Fig. (4.1.1) is the Atlantic water which flows north into the Greenland and Norwegian Seas through the Faroe-Shetland Passage. Most of the water seems to turn around to join the East Greenland Current and flow back to the Atlantic. Certain amounts of it further flows northward into the Barents Sea while some of it tries to force its way into the Arctic Basin through the Fram Strait. The latter is not very successful in breaking into the Arctic Basin from surface level because it encounters the massive cold Arctic surface water flowing out of the Fram Strait.

Due to the contact with cold Arctic water near the Fram Strait, the Atlantic water is cooled somewhat and sinks to an intermediate level of a depth around 400 m. Only at this level does it flow into the Arctic at the Fram Strait. Fig. (4.2.2) gives a picture of velocity and temperature distributions at a transect across the Fram Strait (marked in Fig. (3.6.1)). This figure exhibits that the surface water (Arctic cold water) generally flows out of the Arctic Basin while the relatively warmer Atlantic water consistently flows into the basin near the coast of Spitsbergen. The depth at which Atlantic water flows into the Arctic ranges from 200 m to 1000 m with its core at a depth of about 400 m. The annual mean volume rate of the Atlantic water crossing the transect is shown in Fig. (4.2.3) with some interannual variations. The 7-year mean value is 1.0 Sv, which is close to the estimate of Rudels (1987) as mentioned in Section 2.3. Also included in this plot is a curve from the short-term (1979-85) prognostic sensitivity study with a prognostic ice-ocean-only model. As is stated before, a prognostic simulation can be obtained by removing the diagnostic terms in Eq.s (3.2.2) and (3.2.3). It seems that the prognostic model has a better
performance in bringing more Atlantic water into the Arctic although this is a short period of simulation so a fully prognostic state has not been obtained. This may be due to the fact that the temperature and salinity structure, without the influence of the climatological data, can freely make adjustment which results in more baroclinic inflow at the Fram Strait.

Finally, some comments can be made about the Eurasian gyre shown both in the fields of the surface velocity and the integrated water transport. The observation shown in Figs (2.3.3) and (2.3.4) and the previous Arctic numerical studies do not indicate a consistent existence of the Eurasian gyre. However, the gyre generated by this model is quite justifiable and we have reasons to believe it may actually exist. We note that the model is of rather high resolution both horizontally and vertically. It can very well resolve the complex topography in the Arctic, particularly those narrow ridges which crude resolution models from previous studies could not resolve.

Consequently this model is greatly subject to the effects of so called topographic steering which is deduced from the potential vorticity conservation theory (Pond and Pickard, 1983). The potential vorticity is defined as \((\zeta + f)/H\), where \(\zeta = (∂V/∂x - ∂U/∂y)\) is the relative vorticity due to fluid motion (also see Pedlosky, 1987). Topographic steering plays a particularly important role near the North Pole because the spatial variation of planetary vorticity is relatively small while that of ocean depth is large so that the contour map of the potential vorticity \(f/H\), an approximate version of potential vorticity with \(\zeta\) neglected, is similar to that of bottom topography, as is shown by comparing Fig.(3.6.2) and Fig.(4.2.4). The topographic steering will force the flow along the contours of constant \(f/H\). Therefore those resolved ridges, particularly the Lomonosov Ridge, would certainly deflect the flow to be along the direction they stretch due to the topographic steering effects. This situation will be further examined in Section 4.5 where an analysis of the integrated vorticity balance at a transect across the Arctic Basin is conducted.
4.3 Trace of the Atlantic Water in the Arctic

In fact, the Atlantic water intrusion at the Fram Strait penetrates quite deep into the Arctic Basin before being mixed up with the Arctic cold water and losing its identity. This feature is shown in Fig.s (4.3.1) and (4.3.2). In Fig.(4.3.1) the velocity and temperature distributions at transect J63 (a transect at grid cells j=63, similarly for some other transects cited later; see Fig.(3.6.1) for their location) are plotted. In Fig.(4.3.1) the Atlantic water clearly distinguishes itself from the Arctic water in terms of the differences in velocity and temperature. However, along the way up into the Arctic, the water gradually shrinks and eventually loses its distinction, which is shown in Fig.(4.3.2), a plot of temperature distributions at transects J72 and J80 respectively. On the other hand, the Atlantic water also comes, in a long journey, into the Arctic Basin through the Barents Sea. However, by the time it reaches the Arctic Basin, it is not what it was and is hard to identify.

Fig.(4.3.3) presents an annual mean vector plot for velocity distribution at a depth around 400 m. It shows an overall picture of the simulated movement of the Atlantic water in the Arctic. The Atlantic water, similar to the surface circulation, also forms a clockwise circulation in the Beaufort Sea and an anti-clockwise one in the Eurasian Basin, with some flowing out at the west of the Fram Strait. This circulation pattern is basically in agreement with what is shown in Fig.(2.3.4). Fig.(4.3.4) shows two pressure fields, one is at the same depth (407 m) as the velocity field in Fig.(4.3.3), the other is at the depth 300 m. Following the contours at the Fram Strait, one can see the indication of Atlantic water inflow, but the outflow is not that obvious.

4.4 Beaufort Sea Undercurrent
Velocity distribution at a transect across the Arctic provides a good reflection of the major circulation features. From Fig.(4.3.1a) one can not only recognize the two-gyre circulation in the Beaufort Sea and Eurasian Basin respectively and know that these two gyres mainly form at layers above 400 m, but also can see the coast current reversal against the Beaufort gyre off Alaska (the Beaufort Sea undercurrent). This current reversal can be seen more clearly from a localized figure, Fig.(4.4.1). In this figure, the simulated monthly mean Beaufort Sea undercurrent is illustrated for March and September, 1983. Also included in the figure is observed Beaufort Sea undercurrent based on hydrographic data and current measurements (Mountain, 1974). The simulated current is over 2 to 3 grid cells (80 to 120 km wide) with typical summer mean current speed of 15 cm/s, while the observed current (shaded area) in Fig.(4.4.1c) is about 74 km wide with speed up to 30 cm/s. As is noticed before, previous numerical studies were not able to obtain such a realistic current. The models used in those earlier studies generally have cruder resolution than this model has. In addition, some models did not have ocean open boundary at the Bering Strait.

In order to examine the mechanism that leads to such a current, two sensitivity studies were carried out besides the main simulation. One sensitivity study is a simulation using only air stress as ocean surface stress input with the ice interaction force dropped. This study is based on the assumption that the fine horizontal and vertical resolution of the model may result in a favorable topographic steering or a better ice-ocean interaction. Dropping the ice interaction term is seen as a changing of ice-ocean interaction. The other study is a simulation that closes all the open boundaries, therefore no inflow is available at the Bering Strait and the Faroe-Shetland passage and there is no outflow at the Denmark Strait. This study, however, is based on the reasoning that the Bering Strait inflow is a major factor behind the current reversal. For the purpose of easy identification, the former is hereafter called No-Fi while the latter No-OB. Both simulations are over a period of 7
years after the initial 7-year spin-up. The 7-year spin-up for No-Fi is the same as the main experiment, whereas the 7-year spin-up for No-OB started from scratch with all open boundaries shut at the very beginning.

Fig. (4.4.2) shows the annual (1979) mean surface velocity fields for the three cases: Ice-Ocean-Only, No-Fi and No-OB. Ice-Ocean-Only gives a sizable Beaufort Sea undercurrent, while both No-Fi and No-OB have a current westward. Fig. (4.4.3) is a comparison of streamfunction contours among the three cases, which reflects the same features as exhibited in Fig. (4.4.2).

In searching for what is causing the different patterns of the coast current, ocean surface stress time series at grid cell (14,69) are examined. This grid cell is right where the current reversal occurs (Fig. (3.6.1)). Fig. (4.4.4) illustrates the stress component in the y direction (the westward direction of the Alaskan coast line at that location) and the coastal water transport from the three cases. As can be seen from the comparison between Ice-Ocean-Only and No-Fi, the water transport by Ice-Ocean-Only is mostly eastward, while that by No-Fi is toward the opposite direction. The addition of the ice interaction force to the air stress reduces the magnitude of the total surface stress input. This is particularly true in March through June when ice accumulates at that grid cell and significantly reduces the air stress, which is shown in Fig. (4.4.4a). However, the ice interaction force generally does not change its direction, which is conceivable because the ice serves as a medium to dampen the action of wind on the ocean. Nevertheless, two different flow patterns are shown with and without the ice interaction force. An explanation for this is that a portion of inflow from the Bering Strait may have a tendency to flow eastward along the Alaskan coast toward the Beaufort Sea if there exists ice to subdue the air stress, otherwise the often easterly wind will frequently prevent this from happening. If the inflow was cut off, the flow pattern changes as there is no inflow to move along the coast.
This is confirmed by Fig.(4.4.4b) which is a comparison between the results from Ice-Ocean-Only and No-OB. The coast current generated by No-OB almost always flows toward the Chukchi Sea. An interesting feature with it is that during the period of March through June, the surface stress and water transport are almost constant in magnitude and opposite to each other in direction. One can imagine that in that period the wind stress has little effect on the ocean and due to the lack of the Pacific water inflow from the Bering Strait the flow tendency of the Beaufort gyre prevails which is Chukchi-bound. The relative motion between the flowing water and the relatively steadfast ice results in a surface stress in the opposite direction of the flow. For the full model, however, the inflow from the Bering Strait has a tendency to flow against the Beaufort gyre and prevails for some distance. This is why the Beaufort Sea undercurrent is referred to as a continuation of the Bering Sea water flow in Mountain (1974).

The formation of the Beaufort Sea undercurrent underlines the importance of the bottom topography since the eastward flow appears to be steered by the topography contours and follows the shelf break. On the other hand it also shows the necessity of using a high resolution model to resolve the topography, the shelf break and the narrow current. This is why previous crude resolution models (ranelli and Hibler, 1991; Ries and Hibler, 1991) failed to simulate this current even though some of the models did have inflow conditions at the Bering Strait.

A by-product of the model effort is the finding that No-Fi produces generally stronger surface current with a local gyre in the East Siberian Sea. This is because in case No-Fi the ocean no longer has an ice cover to reduce the wind impact since the air stress is treated as directly acting on the ocean surface. However, No-Fi does not necessarily enhance the Atlantic water inflow at the Fram Strait since the Atlantic inflow is below the
surface layer there. This is confirmed by Fig.(4.4.5) which shows monthly 7-year mean Atlantic water inflow volume values. The results in this figure indicate that No-Fi brings into the Arctic a water volume close to Ice-Ocean-Only, while case No-OB significantly reduce this inflow. This reveals that the open boundary conditions used in the full model not only help in generating a realistic Beaufort Sea undercurrent but also result in a robust Atlantic water inflow to the Arctic.

4.5 Analysis of Integrated Vorticity

In Section 4.2, a qualitative discussion has been made regarding the role of potential vorticity conservation in shaping up the ocean circulation patterns in the Arctic. In this section we further analyze the factors that influence the patterns of water transport by looking at the vertically integrated vorticity equation. This equation describing vertically integrated vorticity balance can be used to quantitatively examine the major contributors for the circulation patterns. This vertically integrated vorticity balance is obtained by integrating the ocean momentum equation (Eq.(3.2.1)) from -H to 0, taking the curl and replacing the vertically integrated velocities by the vertically integrated volume transport streamfunction $\psi$. After some manipulation the following equation is obtained:

$$\frac{\partial}{\partial t}(\frac{\partial^2}{\partial x} + \frac{\partial^2}{\partial y})\psi = J(\psi, \psi) + J(p_b, H) + \text{curl}_{x} \tau^S + \text{curl}_{x} R,$$

In Eq.(4.5.1), $p_b$ is the bottom pressure (summation of $p_s$ and weight of water column at related depth) and $\tau^S$ the surface stress. $R$ is the integrated effect of lateral frictional stress:

$$R = \rho_0 A_H \int_{-H}^{0} (\nabla_H^2 U + \nabla_H^2 V) dz.$$

The Jacobian operator $J$ is given
In Eq. (4.5.1) the first term is the vorticity component due to local change with time, the second one is the planetary vorticity tendency, the third one is the bottom pressure torque, the fourth one is the wind curl and the fifth one is the vorticity term due to lateral frictional stress. Two other terms, bottom friction and non-linearity terms, are found negligibly small and discarded. Semtner (1973) found the term of local change is also small, but in this model it is sometimes not negligible and therefore kept. Instead the planetary term \( J(f, \psi) \) has been found small here but nevertheless included in the equation. So basically there are four integrated vorticity terms in the balance.

These four terms' monthly mean values of June, 1979 across a transect J69 in the model configuration are plotted in Fig. (4.5.1), including results from cases Ice-Ocean-Only and No-Fi. For both cases, the monthly mean local change is relatively small while the mean wind curl is even smaller. This is because these two terms often change sign and become small after being averaged over a one month period. Therefore they play a secondary role in determining mean transport. The dominant monthly mean terms are from bottom pressure and lateral friction. These two terms are basically in balance with each other, and the monthly mean vertically integrated flow across J69 are closely related to them, showing a two-gyre circulation pattern in the Arctic Basin. This implies that the bottom topography is important in shaping up this circulation pattern following the rule of vorticity balance.

The term of local change in No-Fi is generally larger than that in Ice-Ocean-Only since the wind stress only is subject to more variation with time. However, the difference between Ice-Ocean-Only and No-Fi in terms of bottom pressure and lateral friction terms are not very significant in these two major basins in the Arctic. This is why the water

\[
J(f, \psi) = \frac{\partial f}{\partial x} \frac{\partial \psi}{\partial y} - \frac{\partial f}{\partial y} \frac{\partial \psi}{\partial x}
\]
transport patterns between these two cases are basically the same except near a part of Alaskan coast where the simulated coast current often has different directions in the two different cases. The vorticities induced by surface pressure (for flat bottom) and the lateral friction terms from the two cases do have some differences from each other near Alaska, but the differences are not significant enough to draw any conclusion, as pointed out in Section 4.4.

It should be also noted that flow along closed contours will not affect the bottom pressure term. Consequently increase in surface stress can increase net flow in the basin (see Fig. (4.4.3)) even though the vorticity differences are not highly noticeable in Fig.(4.5.1)

4.6 Temperature and Salinity Fields

In Section 4.3, the temperature distributions at some transects are illustrated in conjunction with velocity distributions to trace the Atlantic water flow. Nevertheless in this section some more mention is made of the temperature fields. However, more attention will be put on the salinity fields not because they have been under-described but because they are crucial in determining the density fields and baroclinic circulation.

- Fig.(4.6.1) shows calculated 7-year mean vertical temperature and salinity profiles by the ice-ocean-only model in a number of locations chosen to correspond to some of those shown in Fig.(2.3.1) (also see the locations marked in Fig.(2.1.2)). The overall pattern of the simulated temperature and salinity profiles for these different locations is basically close to the observation (Fig.(2.3.1)), with some differences in magnitude for some individual locations. Particularly, the observed surface salinity in Eurasian Basin is
higher than the simulated while the observed surface salinities in the Beaufort Sea are lower than the simulated.

The salinity values at different locations are determined by seasonal cycle of freezing and melting and may not be as important as salinity patterns and gradients. Therefore, Fig.(4.6.2) is created to show the computed 7-year mean salinity fields in February and September at the surface layer from the ice-ocean-only model. As shown in this figure, the summer surface salinity is smaller than the winter surface salinity due to larger summer river runoff and ice melting. For both winter and summer seasons most of the surface fresh water is near the Siberian coast and the Alaskan coast. A comparison between Fig.(4.6.2b) and the observation shown in Fig.(2.3.2) shows that the surface salinity in the summer is in satisfactory agreement with the observation both in magnitude and in pattern.

Fig.s (4.6.3) and (4.6.4) show 7-year mean temperature and salinity distributions at transects J10 and J32 located in the GIN Sea (see Fig.(3.6.1) for the locations of J10 and J32). In these figures and other similar figures the dashed lines represent temperature less than zero or salinity less than 32.4. From these two figures one can easily identify the cold and fresher Arctic water near the east coast of Greenland and warmer and more saline Atlantic water at the east of GIN Sea. Fig.(4.6.5) shows the salinity distributions at transects J63 and J80 (see Fig.s(4.2.2), (4.3.1) and (4.3.2) for temperature transects).

4.7 Ocean Heat Budget

The North Polar Oceans serve as a huge heat sink for the balance of global heat by releasing tremendous amount of heat to space during the long polar winters, it is therefore important for a model to be able to maintain a balanced heat budget. This is also a
good indication of how critical the diagnostic terms are. On account of this an analysis of heat balance is conducted for the whole region under consideration and its three different areas marked in Fig.(3.6.1). Two cases are examined here: the main experiment (Ice-Ocean-Only) and the prog nostic simulation mentioned also in Section 4.2.

Fig.(4.7.1) illustrates the 7-year mean monthly varying heat transfer in the three different areas and the whole region. In this figure, the surface heat losses in the different areas are shown to be of strong seasonal dependence with maximum loss in winter and minimum in summer. Whereas the heat exchange across the Denmark Strait and the Faroe-Shetland Passage is hardly seasonally dependent because the heat stored in the Atlantic water does not dramatically change seasonally. The Atlantic water is the main heat supply to the north polar oceans with some contributions from summer surface heat gain and the Pacific water coming in through the Bering Strait. The seasonal variations of the heat exchange across the Fram Strait are also small from Fig.(4.7.1c). Note that the reference temperature is chosen to be -1.96°C when calculating water carried heat, this is why sometimes the heat coming into the Arctic through the Fram Strait is negative.

Fig.(4.7.2) gives annual mean heat transport in the whole region for the diagnostic main simulation and the prognostic sensitivity study. In Fig.s (4.7.2a) and (4.7.2b), curve A is the heat coming in through the Denmark Strait and the Faroe-Shetland Passage, curve B is the heat coming in through the Bering Strait and curve C is the heat loss to the ice or space through ocean surface. In addition to giving an idea of how much heat transfer is occurring, the figure indicates that the total heat for the main experiment is only slightly off balance in this seven-year period. However, it is considered satisfactory in that the amount of unbalanced heat is small relative to the total heat loss or gain. It is interesting to note that the prognostic sensitivity run results in a less loss of heat through the ocean surface and a better heat balance, this is more clearly shown in Fig.(4.7.2c),
where annual net heat gain is plotted for both simulations. The reason is that the removal of the diagnostic terms forces the model to adapt itself to the environment with a balanced heat budget.

Fig.(4.7.3) shows the annual mean heat transfer in the Arctic Basin through several ocean openings and the surface. It is noticed from this figure that due to the existence of the ice cover as a good heat insulator, the heat loss to space is much smaller compared with the loss in the whole region. The 7-year heat budget from Ice-Ocean-Only is slightly off balance, which is exhibited in Fig.(4.7.3c). However, the prognostic run again shows its ability to maintain a better balance. And it is not hard to see, from Fig.(4.7.3), that the main simulation results in more surface heat release to the space and gains less from the Fram Strait compared to the prognostic simulation. The less surface heat loss in the prognostic simulation may be attributable to the changed temperature and salinity fields which result in less oceanic heat flux. The more incoming heat from the Fram Strait in the prognostic simulation is because of a larger warm Atlantic water inflow (see Section 4.2).

In the Barents Sea, the situation is slightly different: the main simulation loses much more heat through the surface whereas it gains slightly more through the Barents Sea Opening than the prognostic run. It seems, from Fig.(4.7.4), that the heat from the Atlantic water is not sensitive to the diagnostic terms. If the prognostic model brings more Atlantic water and its heat into the Arctic, then it must bring less into the Barents Sea. This is verified in Fig.(4.7.5) where the heat budget of the Greenland Sea is plotted. This figure shows that both models make little difference in the Greenland Sea regarding the heat budget. All this may suggest that the diagnostic model is of some limitation due to the constraint of the diagnostic terms while the prognostic model has potential to bring more Atlantic water and heat into the Arctic and eventually reach heat balance there and at the same time maintain a balance in other areas by adjusting surface heat loss. As discussed
before, a prognostic model is not subject to the constraints imposed by diagnostic terms and its temperature and salinity fields freely evolve according to the surface and lateral flux conditions. Consequently, in a period as short as seven years, the prognostic model, once free of the effect of the diagnostic terms, cannot adjust itself all the way to equilibrium which would need many years because of ocean's long time-scale processes. It should be interesting to conduct a long-time prognostic simulation at some point with a stratified but horizontally uniform initial temperature and salinity fields using such a high resolution ice-ocean model. However, such a simulation is beyond the scope of this study.

Another feature worth mentioning is that both the Greenland Sea and the Barents Sea release more heat to the space than the Arctic Ocean. This is because less area is ice covered in these two areas and, hence, the heat there is more subject to direct surface loss.

4.8 Oceanic Heat Flux

As is recognized, oceanic heat flux plays a role in ice growth and decay, particularly in the ice marginal zone. Rather than presenting an overall heat flux distribution in the Arctic, Greenland and Norwegian Seas as did Hibler and Bryan (1987), here the different heat flux components at some typical locations are examined in order to gain an insight of how different ocean processes contribute to the whole heat flux. Here the oceanic heat flux at a grid cell is defined as the heat coming from its adjacent water and the deep ocean to its upper two levels. The oceanic heat flux components include those due to lateral and vertical advection, lateral and vertical diffusion and overturning. This heat flux is a summation of these five. The components at a number of selected locations are plotted out in Fig.s (4.8.1) to (4.8.5). In these figures, 'LA' stands for lateral Advection, 'VA' vertical advection, 'LD' lateral diffusion, 'VD' vertical diffusion, 'OT' overturning and
TOTAL' total heat flux. LA, LD and OT are drawn in solid lines while VA, VD and TOTAL dashed line.

Fig.s (4.8.1) shows the heat flux components at grid cell (i,j)=(25,54) (see Fig.(3.6.1)) in a region of the Beaufort Sea where ice always exists. In such a region under ice cover and far away from the ice margin, the water is relatively uniform horizontally in the upper ocean and the heat flux component due to the lateral diffusion is rather small. Due to the water stratification, the component of heat flux due to the vertical diffusion is relatively greater than that due to the lateral diffusion. Note that the contribution of the vertical diffusion is almost always positive in the Arctic ice covered areas, and hence its long-term effects on the ice conditions in the Arctic may be significant. Separately the components due to the lateral advection and vertical advection (upwelling or downwelling) are not trivial. However, their net contribution is almost negligibly small because of the continuity of water flow and the relative horizontal uniformity of temperature in the upper several levels. Consequently the sole major contributor is convective overturning which results from the instability of the water column due to salt rejection from freezing ice in winter. The overturning destroys the old stratified water structure and supplies heat to the mixed layer by bringing up the warmer deep water. Note that the overturning term usually is a positive contribution to the total flux since in most cases overturning is mainly due to salt rejection from ice (when ice exists) or cooling (when at open water) and this occurs when the ocean temperature below the surface layer is normally warmer than the surface layer.

For grid cell (94,63) in the Barents Sea, the picture of its heat flux components, as shown in Fig.(4.8.2), is quite different from that at (25,54). Grid cell (94,63) is in a region where the ice retreats and advances in full swing, the atmospheric forcing has more direct effect on the ocean heating or cooling, and, hence, the temperature field is subject to
more horizontal as well as vertical differences. Therefore the net contribution by lateral and vertical advections is not trivial. In other words, the advection terms help redistribute the temperature field. The lateral and vertical diffusion terms sometimes are quite large too and they tend to be negative at the same time in summer. This is because the direct atmospheric heating on the ocean surface results in downward heat supply and horizontal heat supply to adjacent grid cells. As for the overturning component, it still has a considerable contribution, although with more interannual variations.

More attention has been paid to the behavior of heat flux in the Greenland Sea due to its generally more active heat exchange between upper and deep waters and possibly more influence on the ice growth there. Four grid cells along transect J35 have been selected for comparison, (83,35), (88,35), (96,35) and (115,35). The first one is located just off the northeastern Greenland coast where ice cover is observed to be always existing and the cold Arctic water flowing through, the heat flux terms at the grid cell act more or less like those at (25,54), and, therefore are not illustrated here. The second one, (88,35), is located where the Atlantic water meets the Arctic water and ice cover is observed most of the time. The encounter of the two waters results in a more varying field and, hence, the advection terms often do not cancel each other out, as it is shown in Fig.(4.8.3), the water there seems so stratified vertically that very little is contributed to the mixed layer by overturning. An interesting situation can be seen occurring in August of 1979 when most of the heat supplied by lateral advection is transferred to other grid cells by lateral and vertical diffusions instead of vertical advection due to probably lack of vertical flow. Furthermore, in the following month, September of 1979, convective overturning introduces heat to the deeper ocean from the mixed layer. All this could be explained due to the active mixing of the Atlantic water and the Arctic water. Different from (88,35), (96,35) is right in the marginal ice zone where ice retreats and advances seasonally as at (94,63) in the Barents Sea. Its heat flux terms are drawn in Fig.(4.8.4). As can be seen in this figure, there are
two features that distinguish this grid cell from others analyzed before. One is that, except for 1984 and 85, the overturning brings up tremendous heat in winter because of the rather warm intermediate Atlantic water. The other is the large net contribution by the advection terms, implying an important role the advection plays in this region in resisting ice advance in winter and pushing ice back in the summer. Grid cell (105,35) is farther away from the ice edge and open water is almost always observed there. The net contribution of the advection terms becomes smaller again and the lateral diffusion term is close to nil. However, in winter overturning brings particularly large amounts of heat up while in the summer vertical diffusion processes supply sizable heat to the cells beneath it, as shown in Fig.(4.8.5).

4.9 Ice Compactness and Thickness Fields

To obtain an idea of how the ice concentration is distributed, the simulated 7-year (1979-85) mean ice concentration from the main simulation and the corresponding observations are illustrated in Fig.(4.9.1). This simulated 7-year mean fields indicate that over a long period the model overestimates the ice concentration in the Greenland, slightly so in the Barents Sea, and is about right in the Arctic. Fig.(4.9.2) shows a 7-year mean seasonal ice coverages in the three regions: the Arctic Ocean, the Barents Sea and the Greenland Sea. The ice coverage is defined here as the fraction of area that is covered by ice, therefore it can be taken as a measure of ice concentration.

From Fig.(4.9.2) it seems that in the Greenland Sea, the model tends to overestimate ice coverage every month. One reason may be that the Hansen air temperatures modified and used (also see Hibler and Zhang, 1993) in this model are slightly over-modified so that the ice feedback effects are a little too strong. Another reason may be related to the behavior of heat flux in that region. As discussed earlier in Section 4.8 for
the heat flux at grid cell (88,35) shown in Fig.(4.8.3), the lateral and vertical diffusion and overturning terms of heat flux are too small in the ice-covered area in the Greenland Sea, despite the possible mixing process of the flow of the Atlantic water with the Arctic outflow. This suggests that the vertical mixing may not be adequate and an appropriate mixed-layer model may be needed to bring up more heat, particularly associated with warm Atlantic water, so as to at least partially improve the prediction of ice extent in that region.

In the Barents and Arctic Seas, the estimated 7-year mean seasonal ice coverages by the three models are shown to compare favorably with the observation in winter, however, the ice extent does not retreat properly fast in the Barents Sea when approaching summer while the ice advance occurs too slow after summer in both the Barents and the Arctic. It is conceived that this situation may come from the two-level ice model that oversimplifies the categories of ice thickness (see Hibler, 1979) and can not act quickly to follow the changing pace. A variable ice thickness model (Hibler, 1980a; Flato, 1991) is thought to be able to make some improvements in this regard, particularly in the Barents Sea where the greatest seasonal ice extent variations exist.

A 7-year mean Ice thickness distribution is shown in Fig.(4.9.3). This thickness field is compared with Fig.(4.9.4) which shows an ice thickness distribution estimated from submarine sonar data (reviewed in Hibler, 1980b). As can be seen, the simulated ice thickness field is in reasonable agreement with the observation shown in Fig.(4.9.4) and described in Section 2.2 as well.

4.10 Ice Drift and Outflow
Fig.(4.10.1) shows the monthly mean ice velocity distributions for February and September of 1983 and the annual mean velocity of this year respectively. As can be seen, the monthly mean ice drift pattern is not necessarily similar to that of the ocean surface currents because the ice drift is mainly driven by wind. Although ocean surface currents are not a dominant factor in shaping up ice drift pattern, they do affect the ice drift in terms of ice drift speed and direction. Their long term effect is particularly noticeable because of the ocean surface currents' relatively consistent flow direction.

The effect of ocean currents on ice motion is illustrated by comparing simulated ice drift with observed buoy drift. Fig.(4.10.2) presents the simulated ice-drift tracklines with three typically observed buoy tracklines in different areas. The simulated ice-drift tracklines, based on the integration of five-day mean ice velocity, are from Ice-Ocean-Only and a sensitivity study. This sensitivity study (denoted as No-Current) is also a 7-year run which was jointly conducted with the main simulation. In this study, the ocean surface currents are not included in calculating the water drag on the ice so that the ice velocity is free of ocean currents' effect. The ice velocity so obtained is certainly different from that of the Ice-Ocean-Only, but every other model quantity is deliberately kept to be the same during simulation. The observed buoy drift data were obtained from World Data Center A for Glaciology and are based on buoy data collected by the Arctic Buoy Program (see Thorndike et al, 1983).

As can be seen in Fig.(4.10.2), the simulated ice drift is improved with the ocean surface currents in these three individual cases. Fig.(4.10.2a) actually shows simulated and observed ice drifting from the Beaufort Sea into the Chukchi Sea. Once a pack of ice gets into the Chukchi Sea, the inflow from the Bering Strait plays a role in its drift revealed by comparing the results from Ice-Ocean-Only and No-Current. This inflow influence also can be observed in Fig.(4.10.2b) which shows an ice drift pattern right in front of the
mouth of the Bering Strait. The effect of the transpolar current and the East Greenland
Current on ice drift is particularly noticeable on account of the ice drift tracklines in
Fig.(4.10.2c). Without such currents the ice drift would be left far behind what it should be
in reality.

Fig.(4.10.2) only compares some individual buoy drifts. The overall behavior of the
simulated ice drift is evaluated by statistical analysis. It is based on the simulation results
and buoy drift data available in the whole region over the seven-year period of 1979-85.
The statistics verifies that the simulated ice drift generally benefits from the ocean surface
currents, as shown in Table (4.10.1) giving statistic estimations based on 30-day simulated
ice drift and buoy drift in the whole region under consideration. In this table, the drift ratio
gives a measure of how fast the ice drifts relative to the relevant buoy is, while the error
radius tells how far the simulated ice drift is deflected away from the observation.

These statistical values in Table (4.10.1) indicate that the existence of ocean
currents speeds up ice drift to such an extent that is in rather close agreement statistically
with observation. However, the simulated ocean currents are not quite successful in
reducing the deflection of ice drift from buoy drift to a satisfactory degree. On the other
hand, the model without the effect of ocean currents would slow down ice drift by about
0.5 km per day against the buoy drift, and, hence, leave the ice about 360 km behind
statistically after two-year drifting. It would also result in 0.2 km more ice-drift deflection
per day relative to that from the main simulation. That is 168 km deflection after two-year
drifting:
Table (4.10.1). Statistics of model ice drift and buoy drift over the whole region in 79-85

<table>
<thead>
<tr>
<th></th>
<th>Mean buoy drift (1)</th>
<th>Mean ice drift (2)</th>
<th>Drift ratio (1)/(2)</th>
<th>Mean error radius</th>
</tr>
</thead>
<tbody>
<tr>
<td>Standard</td>
<td>6.6 km/day</td>
<td>6.5 km/day</td>
<td>1.01</td>
<td>3.1 km/day</td>
</tr>
<tr>
<td>No-Current</td>
<td>6.6 km/day</td>
<td>6.0 km/day</td>
<td>1.10</td>
<td>3.3 km/day</td>
</tr>
</tbody>
</table>

As shown in Fig.(4.10.1), considerable ice flows out of the Arctic Basin through the Fram Strait. Even sizable ice moves further down south out of the Greenland Sea through the Denmark Strait. The amount of Arctic ice flowing out of the Fram Strait and heading further south is of interest because obviously it is closely related to the salt or fresh water budget in the Arctic. In addition, from the heat budget perspective, the ice absorbs considerable latent heat when melting, its outflow is therefore an addition of heat and closely related to the whole heat budget in the region. From the standpoint of the outside oceans, on the other hand, the Arctic ice outflow may also affect the thermohaline circulation in the Atlantic and probably the global ocean thermohaline circulation.

Based on these considerations, the ice transports through the Fram Strait and Denmark Straits are examined and plotted here. Fig.(4.10.3) draws the annual ice areal transports out of the Fram Strait from both models. Also in the plot is the estimated areal transport by Colony (1990) based on observation. The main simulation predicts an ice areal transport that closely matches the observational estimate, while the exclusion of ocean currents in computing ice velocity considerably under-estimates the transport. Since the description of areal transport does not give a full picture of how much ice is involved, Fig.(4.10.4) is given where the annual mean ice volume transported out of the Fram Strait and Denmark Strait are plotted. Both ice volume outflows demonstrate considerable interannual variations, although little observational data is available for comparison.
Figure (4.1.1) Mean ocean surface velocity (represented by the third level velocity in the model). Velocity vector at every other grid cell is plotted.
(a) Feb., 1983; (b) Sept., 83; (c) Annual mean of 83.
Figure (4.1.1) (cont.)
Figure (4.1.1) (cont.)
Figure (4.1.2) Annual (1983) mean pressure contours at ocean levels 2 (18 m) and 3 (37 m). The numbers at the bottom of each plot reading from the left to the right are the minimum and maximum contour values and the contour interval, respectively.
Figure (4.2.1) Monthly mean streamfunction contours. (a) February, 1983; (b) September, 1983. Contour interval is 0.5 Sv.
Figure (4.2.2) Annual (1983) mean velocity and temperature distributions at the transect across the Fram Strait. (a) Velocity; The dashed lines stand for flow into the Arctic and the solid lines for flow out of the Arctic. Contour interval is 0.25 cm/s. (b) Temperature; The dashed lines stand for temperature below 0°C. Contour interval is 0.35°C.
Figure (4.2.3) Annual mean Atlantic water flowing into the Arctic through the Fram Strait. Solid line: ice-ocean-only model; Dash-dot line: prognostic sensitivity study.
Figure (4.2.4) Contours of potential vorticity, $f/H$. Unit: $10^{10} \text{s}^{-1} \text{cm}^{-1}$. Contour interval is 4.0.
Figure (4.3.1) Annual (1983) mean velocity and temperature distributions at transect J63. (a) Velocity; The solid lines stand for flow in a direction into the paper and the dashed lines out of the paper. The contour interval is 0.5 cm/s. (b) Temperature; The dashed lines are for temperatures below zero. The contour interval is 0.35°C.
Figure (4.3.2) Annual (1983) mean temperature distributions across (a) transect J72 and (b) J80 with contour interval 0.35°C. The dashed lines are for temperatures below zero.
Figure (4.3.3) Annual (1983) mean ocean velocity at a depth of 407 meter.
Figure (4.3.4) Annual (1983) mean pressure fields. (a) At depth 300 m; (b) At depth 407 m. The numbers at the bottom of each plot reading from the left to the right are the minimum and maximum contour values, and the contour interval, respectively.
Figure 4.4.1 Simulated monthly mean velocity distributions across transect J63 off the Alaskan coast and observed Beaufort Sea undercurrent. The solid lines stand for flow westward, while the dash lines for flow eastward. (a) March, 1983; (b) September, 1983; (c) Observed current (Mountain, 1974). The horizontal spacings for (a) and (b) are actual grid cell spacings. The contour interval for (a) and (b) is 1 cm/s.
Figure (4.4.2) Annual (1983) mean ocean surface velocity.

(a) From Ice-Ocean-Only;

(b) From case No-Fi;

(c) From case No-OB.
Figure (4.4.2) (cont.)
Figure (4.4.2) (cont.)
Figure (4.4.3) Annual (1983) mean streamfunction contours. (a) From Ice-Ocean-Only; (b) From case No-Fi; (c) From case No-OB. Contour interval is 0.5 Sv.
Figure (4.4.4) Daily y-component ocean surface stress (unit: dynes/cm²) at grid cell (14,69) and y-component water transport within the cell (unit: 5x10³ m³/s). (a) Solid line: Ice-Ocean-Only; Dashed line: No-Fi. (b) Solid line: Ice-Ocean-Only; Dashed line: No-OB.
Figure (4.4.4) (cont.)
Figure (4.4.5) Comparison of monthly mean Atlantic water flowing into the Arctic through the Fram Strait in 1979 among cases Ice-Ocean-Only, No-Fi and No-OB.
Figure (4.5.1) Monthly (June, 1979) mean vertically integrated vorticity components at transect J69 over the Arctic Basin. The solid lines stand for water transport across the transect (in y direction) within one grid cell, $V_{\text{mean}}*H*h$, with unit $5\times10^3$ m$^3$/s. (a) From the ice-ocean-only simulation; (b) From the sensitivity study (No-Fi) in which the ocean surface velocity is not used to compute ice velocity.
Figure (4.5.1) (cont.)
Figure (4.6.1) Simulated 7-year (1979-85) mean vertical profiles of temperature and salinity at selected locations in the North Polar Region.
Figure (4.6.2) Simulated 7-year (1979-85) mean surface (5 m) salinity (ppt) fields in (a) February and (b) September. The numbers at the bottom of each plot reading from the left to the right are the minimum and maximum contour values, and the contour interval, respectively.
Figure 4.6.3. Temperature and Salinity distributions at Transect J10. (a) Temperature, dashed-lines for temperature below zero, contour interval: 0.35°C; (b) Salinity, dashed lines for salinity below 34.2 ppt, contour interval: 0.05 ppt.
Figure (4.6.4) Temperature and Salinity distributions at Transect J32. (a) Temperature, dashed lines for temperature below zero, contour interval: 0.35°C; (b) Salinity, dashed lines for salinity below 34.2 ppt, contour interval: 0.05 ppt.
Figure (4.6.5) Salinity distributions at (a) Transect J63 and (b) J80. Dashed lines are for salinity below 34.2 ppt, contour interval: 0.05 ppt.
Figure (4.7.1) 7-year mean monthly heat transport (unit: $10^{12}$ W) in different areas and the whole region. (A) The Greenland Sea; curve A: incoming heat form the Fram Strait; B: incoming heat from the Denmark Strait and Faroe-Shetland Passage; C: incoming heat from the Barents Sea Opening; D: surface heat loss. (B) The Barents Sea; A: incoming heat from the Barents Sea Opening; B: incoming heat from the other openings to the Arctic; C: surface loss. (C) The Arctic Sea; A: incoming heat from the Fram Strait; B: incoming heat from the Bering Strait; C: incoming heat from the other openings to the Barents Sea; D: surface loss. (D) The whole region; A: incoming heat from the Denmark Strait and Faroe-Shetland Passage; B: incoming heat from the Bering Strait; C: surface loss.
Figure (4.7.2) Annual heat transport in the whole region under consideration (unit: $10^{12}$ w). (A) From Ice-Ocean-Only (standard); (B) From the prognostic study; (C) Net heat gain. In (A) and (B) the curves A, B and C are corresponding to those in Figure (4.7.1D).
Figure (4.7.3) Annual heat transport in the Arctic Sea (unit: $10^{12}$ W). (A) From Ice-Ocean-Only (standard); (B) From the prognostic study; (C) Net heat gain. In (A) and (B) the curves A, B, C and D are corresponding to those in Figure (4.7.1C).
Figure (4.7.4) Annual heat transport in the Barents Sea (unit: $10^{12}$ W). (A) From Ice-Ocean-Only (standard); (B) From the prognostic study; (C) Net heat gain. In (A) and (B) the curves A, B and C are corresponding to those in Figure (4.7.1B).
Figure (4.7.5) Annual heat transport in the Greenland Sea (unit: $10^{12}$ W). (A) From Ice-Ocean-Only (standard); (B) From the prognostic study; (C) Net heat gain. In (A) and (B) the curves A, B, C and D are corresponding to those in Figure (4.7.1A).
Figure (4.8.1) Monthly mean heat flux components (w/m²) at grid cell (25,54) in the Beaufort Sea.
Figure (4.8.2) Monthly mean heat flux components (w/m²) at grid cell (94,63) in the Barents Sea.
Heat flux components at cell (88,35)

Figure (4.8.3) Monthly mean heat flux components (w/m²) at grid cell (88,35) in the GIN Sea.
Figure (4.8.4) Monthly mean heat flux components (w/m²) at grid cell (96,35) in the GIN Sea.
Figure (4.8.5) Monthly mean heat flux components (w/m²) at grid cell (105,35) in the GIN Sea.
Figure (4.9.1) 7-year mean ice compactness fields.
(a) from the ice-ocean-only model; (b) Observed.
Contour interval 0.1.
Figure (4.9.2) 7-year mean monthly ice concentrations from the standard ice-ocean-only model and the observation, in the Arctic, Greenland and Barents Seas respectively. The unit is the fraction of ice concentration of the whole area.
Figure (4.9.3) 7-year mean ice thickness distributions from (a) the ice-ocean-only simulation and (b) observation (from Hibler, 1980b). Contour interval for (a) is 0.3 m.
Figure 4.10.1 Mean ice velocity distributions.
(a) February, 1983;
(b) September, 1983;
(c) Annual mean of 1983.
Figure (4.10.1) (cont.)
Figure (4.10.1) (cont.)
Figure (4.10.2) Comparison between simulated ice drift (based on 5-day average ice velocity) and observed buoy drift in 3 different areas. (a) In the Beaufort sea; (b) In the Chukchi Sea; (c) From the Eurasian Basin to the Greenland Sea. 'Standard' stands for Ice-Ocean-Only; 'No current' stands for the sensitivity study in which the ocean velocity is not used in calculating ice velocity.
Buoy 13805, 5/20/81 - 2/24/82, Chukchi Sea

Buoy 11905, 3/1/79-10/2/79, Eurasian Basin->Greenland Sea

Figure (4.10.2) (cont.)
Figure (4.10.3) Annual mean areal ice transport out of the Fram Strait from the Arctic.
Figure (4.10.4) Annual mean ice volume transport out of (a) the Fram Strait and (b) the Denmark Strait.
Mean ice transport rate out Denmark Strait (1000 m^2 m/s)

Figure (4.10.4) (cont.)
Chapter 5. Experiment with the Ice-Ocean-Mixed-Layer Model

The formulation and numerical implementation of the ice-ocean-mixed-layer model have been described in detail in Chapter 3. This model was intended to be integrated for a period of seven years (1979-1985) after the same 7-year spin-up as for the ice-ocean-only simulation. However, to ensure a smooth transition to a better working condition for the mixed layer model, the 1979 calculation was carried out twice. In this chapter some major experimental results from this fully coupled model of a variable depth mixed layer are presented. Meanwhile comparisons are often made between this model and the previous ice-ocean-only model with fixed depth mixed layer. Additionally, the results from one sensitivity study are also used for comparison.

The sensitivity study is based on the standard ice-ocean-mixed-layer model with a somewhat changed parameterization. This sensitivity study is intended to examine how different degrees of vertical mixing would affect the ice-ocean system. In order to do that, the values of decay coefficients $h_W$ and $h_C$ were changed for the sensitivity study. These coefficients are changed in such a way that whenever the mixed layer depth $h_m$ becomes larger than 60 m, they are no longer determined by Eq.(3.3.15), but instead by $h_W = \min\{((60-10)/10 + 4), ((h_m-10)/10 + 4)\}$ and $h_C = \min\{((60-10)/7.5 + 4), ((h_m-100)/7.5 + 4)\}$. This would further suppress deep mixing after $h_m$ reaches 60 m and hence reduce the effects of the embedded variable depth mixed layer. This sensitivity study was also integrated over the 7-year period with a previous year 1979 run for transition. The results from this study are compared with the other models' results whenever appropriate.

In the following Section 5.1, the characteristics of the simulated variable depth mixed layer are presented. Whereas the effects of the mixed layer on the ice conditions and the ocean circulation are discussed in Sections 5.2 and 5.3 respectively.
5.1 Characteristics of the Variable Depth Mixed Layer

Fig.(5.1.1) shows the calculated 7-year (1979-85) mean monthly mixed layer thickness fields in a sequence in which the mixed layer deepens and retreats. As can be seen from the figure, the retreating mixed layer after going through a whole summer and into October has already grown to be about 40 m deep in some areas as the ocean becomes cooler and new ice starts to form. January is a time when the mixed layer actively continues its deepening because of continued ice freezing and cooling. During April, however, it reaches its deepest limit since the summer is approaching. In the summer, the atmospheric heat input melts the ice and warms the ocean surface and accordingly the mixed layer becomes shallower and shallower, which is shown in Fig.(5.1.1d). Apparently, the mixed layer evolution is in a seasonal cycle of growing and decaying which is closely associated with the cycle of atmospheric cooling and warming. This is because atmospheric cooling and warming play a major role in determining the ocean surface buoyancy flux, which is important for the mixed layer formation.

Fig.(5.1.1) also exhibits some features regarding spatial distribution of the mixed layer thickness. An obvious feature is that the mixed layer in the Arctic is generally shallower than in the GIN Sea and the Barents Sea. Whereas in the GIN Sea and the Barents-Sea the deepest mixed layer more often occurs near the ice edges and in the areas not covered by ice. Furthermore it is noticed that the mixed layer tends to become deeper near the west of Spitsbergen where the West Spitsbergen Current tries to break into the Arctic through the Fram Strait. This feature plays a role in bringing the predicted ice margin at that location closer to the observed as will be discussed later. All these features are attributable to the spatially different conditions of ocean surface mechanical energy
input, due to wind or ice motion, and buoyancy flux, due to surface warming, cooling or ice salt rejection, which all play a role in mixed layer variation as pointed out before.

The ocean surface conditions are considerably different from region to region in the North Polar oceans. In the Arctic, for example, the ocean is ice covered during most of the year. The surface stress on the ice covered ocean can be much smaller than that on the open water as demonstrated in Fig. (4.4.4a). Therefore the mechanical energy input to the mixed layer there can be relatively insignificant compared to some part of the GIN Sea and the Barents Sea where open water is often observed. On the other hand, the existence of an ice cover effectively insulates the ocean from the direct influence of the atmospheric warming and cooling. The atmospheric effects on the mixed layer are reflected in the ice's melting or freezing which determines the surface salt flux into the mixed layer. While in an area of open water the mixed layer is affected by direct atmospheric warming and cooling through surface heat flux or salt flux when ice grows in addition to a generally larger wind energy input. However, in a marginal ice zone, the ocean surface interactions may be more complicated with probably all the features in either cases. In order to obtain an insight as to how these features affect the mixed layer and how the variable depth mixed layer in turn affects the ice-ocean system, the results at two typical transects are presented for the purposes of somewhat detailed analysis. One transect is J37 which crosses the Greenland Sea at J = 37 (Fig. (3.6.1)). The other is J63 which connects Alaska at one end and Franz Josef Land at the other along J = 63. Transect J37 is right across the often observed ice marginal zone at the Greenland Sea, so it is desirable to select this transect to examine the vertical-mixing processes near the ice edge. While Transect J63 is across the central Arctic where ice exists in all seasons as well as a limited area of marginal ice zone in the Beaufort Sea just off the Alaskan coast sometime in the summer. The results at these two transects are given and discussed separately in the following.
Fig.(5.1.2) shows a transect versus time plot of a 7-year mean mixed layer depth in which the numbers on the horizontal axis are the grid cell numbers in the i direction along J37 (same for following similar figures). Also drawn in dashed line in this figure is the calculated 7-year mean ice edge variation with time by the ice-ocean-mixed-layer model. The ice edge is defined by the 0.2 concentration contour with ice covered areas located to the left and open water to the right. From this one figure, one can see not only the seasonal cycle of the mixed layer's growing and shrinking but also the spatial variations. Generally, the simulated mixed layer is deeper in the open water than in the ice covered area except in the summer when the mixed layer retreates to its minimum depth set to be 10 m by the model. However the most prominent feature the figure shows is that the mixed layer depth is more greatly increased in crossing the ice edge from the ice covered area to the open water in all seasons except the summer. This indicates that extraordinarily active vertical mixing is going on near the ice edge. This is further confirmed by Fig.(5.1.3) which shows similar plots for three consecutive years (1981 and 83) in representing other years in this regard.

Although the intensity of the extraordinary mixing near the ice edge differs from year to year, Fig.(5.1.3) shows that it always exists and the ice edge stops right at the left side of the mixing area. The intensive mixing in the area prevents the ice from extending further to the open water since it brings up relatively larger oceanic heat flux to melt the ice. This is particularly true in an area with the warm Atlantic water underlying the surface water (Fig.(2.3.1)). This is verified in Fig.(5.1.4) which exhibits the transect versus time plot of the 7-year mean oceanic heat fluxes from both the fully coupled ice-ocean-mixed-layer model and the ice-ocean-only model of fixed depth mixed layer. As can be seen, the
variable depth mixed layer obtains more oceanic heat flux from the underlying water than the fixed depth mixed layer, particularly in the ice marginal zone (also see Fig.(5.1.1)).

The manifestation of a larger oceanic heat flux supply to the mixed layer is a reduced ice concentration and thickness since some of the ice is melted by the oceanic heat. Fig.(5.1.5) shows the 7-year mean ice edge variations with time along the transect from the two main models and the observation. The mixed layer model results of ice extent agree better with the observation than the fixed depth mixed layer model which generally over-predicts ice extent, with about 5 grid cells' ice along the transect being over-estimated. Fig.(5.1.6) shows the 7-year mean ice edge spatial distributions over the Barents and GIN Seas for February and September, representing winter and summer. As can be seen in Fig.(5.1.6a), the ice edge from the ice-ocean-mixed-layer model is closer to the observed in the Barents Sea. In the Northern Greenland Sea, the observed ice edge has a dent and, interesting enough, the ice edge from the ice-ocean-mixed-layer model also has one, although the location is not exactly the same. The location of the simulated ice edge dent, however, is where a mixed layer depth contour dent is observed (Fig.(5.1.1.c)), indicating a close correlation between the ice edge shape and the vertical mixing. The ice edge is also improved in the southern Greenland Sea near Iceland. In summer (Fig.(5.1.6b)), noticeable improvement in ice edge is observed in the central Greenland Sea, while elsewhere there is no much difference between the two models.

Although the most active vertical mixing observed is near the ice edge, the reason behind it needs to be explained. Therefore the surface conditions at that transect are examined since they are the governing factors for the vertical mixing processes. The surface conditions consist of surface stress and heat and salt fluxes. Fig.(5.1.7a) shows the 7-year mean friction velocity $u^*$ along the transect. $u^*$ is a measure of the surface mechanical energy input which is always a production term for the mixed layer. As can be
seen, $u^*$ is normally larger in the open water area than in the ice covered area, but its magnitude rarely exceeds 3 cm/s. Although the larger values of $u^*$ in the open water area would generally promote more mixing, it does not seem to be able to result in dramatic mixing near the ice edge, compared to the entrainment velocity pattern shown in Fig.(5.1.7b) and the mixed layer depth distribution in Fig.(5.1.3a). The entrainment velocity distribution also exhibits the feature that across the ice edge there is a steep change in magnitude of the velocity.

The ocean surface salt fluxes at the transect from the two main models are shown in Fig.(5.1.8). Both models have negative surface salt flux along the ice edge in all the seasons, indicating that the ice there is almost in a constant melting state. For the ice-ocean-mixed-layer model, there is always tremendous negative salt flux into the ocean at the ice-covered side of the ice edge except for the summer, much more than that of the ice-ocean-only model, because of the vertical mixing process that brings up more oceanic heat to melt the ice.

In the figure, one can also see a small positive surface salt flux at the open water side of the ice edge in summer. This positive salt flux is not believed to be due to actual ice melting. Instead, it may result from the central finite differencing scheme that is used to calculate ice advection in the ice model. The central differencing scheme is of 2nd order numerical accuracy, but it occasionally produces unreal negative ice (positive surface salt flux) during summer when ice is retreating. In winter, when the ice is expanding, this problem is ameliorated. Therefore, this problem hardly affects the calculation of the vertical mixing since in summer, the mixed layer is retreating anyway. However, further improvement is desirable by tracing the unreal negative ice and removing it after the ice advection calculation in the ice model.
The ice melting induced surface negative salt flux tends to stabilize the mixed layer and reduce the vertical mixing because the potential energy in the mixed layer is decreased. However, as observed in Fig.s(5.1.2) and (5.1.3), the mixing along the ice edge is not small at all in winter and fall. This indicates that the surface cooling (negative heat flux into the ocean) along the ice edge is important, in addition to the wind or ice stirring effects, in maintaining such a complicated mixing process that at the ice covered side, the ice is melting to reduce the vertical mixing, while at the open water side, the mixing is deepening.

The importance of the surface heat flux in the vertical mixing pattern near the ice edge is confirmed by Fig.(5.1.9). In this figure, the 7-year mean surface heat flux distributions at the transect from both models are drawn. The surface heat flux here is defined as the atmospheric heat flux on the ice surface if ice exists or on the ocean surface if no ice exists. Fig.(5.1.9) demonstrates a significant discontinuity of heat flux across each model's ice edge in the cooling seasons with much larger surface heat release to the atmosphere at the open water side of the edge, which is particularly true for the ice-ocean-mixed-layer model. This is because the atmospheric cooling is directly acting on the ocean surface near the ice edge where no ice exists and greatly enhances the mixing there.

Fig.(5.1.10) shows the 7-year surface salinity distributions at the transect from both models. The salinity from the ice-ocean-only model is higher than that from the ice-ocean-mixed-layer model since the latter melts more ice. The surface temperature distributions from the two models are shown in Fig.(5.1.11). The temperature patterns from the models are rather close except for some differences near the ice edge since the different models have different shapes of ice edges at whose ice-covered side, the temperature is at freezing point. The temperature patterns are also close to that obtained in a two dimensional model (Houssais and Hibler, 1993).
(ii) Transect J63

Fig.(5.1.12) shows the transect versus time plot of the 7-year mean mixed layer depth along Transect J63. The depth varies ranging from the specified minimum depth of 10 m to about 70 m. This range is not far from the observations as reviewed in Section 2.3. There is a steep change of depth near grid cell (12,63) close to the Alaskan coast simply because of the topography change. Note that the mixed layer depth in the Arctic is smaller than that in the GIN Sea and the Barents Sea. One reason for this is that the surface stress is generally smaller in the ice covered region comparing Fig.(5.1.7) to Fig.(5.1.13) which illustrates the 1981 friction velocity distribution along the transect. Another reason may be attributable to the fact that in the ice covered area the mixed layer temperature is normally maintained at a constant freezing point and the atmospheric cooling or warming only affects the ocean through the ice melting or freezing which determine the ocean surface salt flux. The salt flux alone is not able to produce a vertical mixing as active as in the ice marginal zone in the GIN Sea.

Fig.(5.1.14) shows the 7-year mean oceanic heat flux distributions along Transect J63 from the two models. As can be seen, the mixed layer model brings much more heat up at the Beaufort Sea marginal ice zone, particularly in the winter and spring, while near Franz Josef Land, the largest heat fluxes occur at slightly different locations for the different models. In the central Arctic, however, the ice-ocean-mixed-layer model has noticeably less oceanic heat flux than its counterpart in the cooling seasons. One possible reason is that the ocean temperature there is rather uniform down to certain depth from the surface as shown in Fig.s (2.3.1) and (4.6.1) and the vertical mixing does not go beyond that depth (Fig. (5.1.12) so that there is little advantage for the mixed layer model to bring up more heat via mixing. Another possible reason may be due to the fact that the model does not allow vertical diffusion processes at the base of the mixed layer. As is discussed
in Section 4.8, vertical diffusion contributes a certain amount of heat to the upper ocean in the Arctic ice covered area. Its long-term effect is not trivial. Once the heat brought up by the mixing is less than the diffusion's contribution, it is not surprising to see a lesser heat flux by the mixed layer model.

Fig.(5.1.15) shows the 7-year mean surface salt flux distributions from the two models at the transect. The most prominent difference between these two models still occurs at the Beaufort Sea ice edge where the ice-ocean-mixed-layer model has more negative salt flux than its counterpart in most months except in January through March when more positive salt flux is produced. Given that the surface salt flux reflects the conditions of ice melting or freezing, it seems that at the ice edge, the seasonal cycle of melting and freezing simulated by the mixed layer model is more variable: in the summer more ice melts while in the winter more ice grows. The other noticeable difference in surface salt flux distribution is in the central Arctic where the mixed layer model results in a little more ice melting, but it is not significant.

Fig.(5.1.16) is a comparison of the 7-year mean surface heat flux distributions between the two models at Transect J63. As is shown in this figure, the ice-ocean-mixed-layer model is subject to more surface cooling at the Beaufort Sea ice edge than the model of fixed depth mixed layer. Elsewhere, there is not much difference between the two models. Note that the patterns of the atmospheric surface heat flux distributions are closely matched with those of surface salt flux distributions in the central Arctic from both models, indicating that the atmospheric heating or cooling is the dominant factor in ice growth and decay, not the oceanic heat flux.

The mechanisms that drive the vertical processes along the ice edges of the Beaufort and Greenland Seas are quite different as can be seen by their respective
distributions of ocean heat flux, surface salt flux and surface heat flux. At the Beaufort Sea ice edge, the surface cooling, the salt rejection due to the ice growth and the winds promote the deepening of mixing, but the mixing does not bring up enough oceanic heat to melt the ice. Whereas at the Greenland Sea ice edge, the mixing, as described before, brings up enough heat to allow a continued ice melting nearby that results in negative salt flux to resist the deep down mixing. Nevertheless, the mixing goes on despite the negative salt flux input. This difference likely results from the difference of the water masses in the two seas which have different temperature profiles.

Fig.(5.1.17) shows the 7-year mean ice thickness distributions at the transect from both models. The ice-ocean-mixed-layer model generates less thick ice in the Beaufort Sea marginal ice zone but produces thicker ice in the central Arctic. This pattern is closely related to that of oceanic heat flux shown in Fig.(5.1.14). The 7-year mean surface salinity distributions (at depth 5 m) are shown in Fig.(5.1.18) where the ice-ocean-mixed-layer generates very fresh water near the Alaskan coast from summer through fall. This may be due to the melted ice that forms a rather stratified ocean so that the continuing supply of fresh water due to the continuing ice melting concentrates in the very shallow mixed layer and therefore results in a layer of very fresh water. This situation may be ameliorated if the vertical eddy diffusion at the base of the mixed layer is put back. Nevertheless, the lowering of surface salinity does not seem to have a negative effect since it mainly happens in a limited area during melting season.

Finally, Fig.(5.1.19) is used to show the vertical profiles of the temperatures and salinities from the ice-ocean-mixed-layer model at the same selected locations as those in Fig.(4.6.1). The temperature profiles at the grid cells in the Beaufort Sea and Eurasian Basin are about the same as those from the ice-ocean-only model, but the surface temperatures at the south and north Greenland Sea are closer to the observation in
Fig.(2.3.1). The first level salinities to the south and north of the Greenland Sea and the
Eurasian Basin are lower than those from the ice-ocean-only model. While at the Beaufort
Sea the first level salinity is about the same. The lowering of surface salinity may be
attributable to the same reasons mentioned in the previous paragraph. However, the
subsurface temperature and salinity profiles from both models do not seem to have large
differences.

5.2 Ice compactness and Thickness

In this section, comparison is made mainly between the two main models to
examine the effects of the variable depth mixed layer on the ice fields. Fig.(5.2.1) shows
the 7-year mean fields of ice compactness and ice thickness from the fully coupled model.
As can be seen from this figure, the mixing mechanism does not reduce the ice
compactness in the Barents Sea compared to that from the ice-ocean-only model. This may
be due to the fact that the ocean is shallow at the location where the ice edge is often
observed and there is no underlying water warm enough for the mixing process to bring up
enough heat. For example, the temperature at grid cell (105,55), located close to the 7-year
mean ice edge in the Barents Sea, has a temperature profile that decreases with the depth
within certain range as shown in Fig.s (5.1.19) and (4.6.1). The Arctic also sees little
change in ice compactness. The most noticeable change occurs in the Greenland Sea where
the mixed layer model generally reduces the ice compactness. And the most interesting
change in the Greenland Sea ice compactness, brought about by the variable depth mixed
layer, is a dent on the ice compactness contour near West Spitsbergen. This dent is quite
similar to the observation shown in Fig.(4.9.1b), although not as much as the observation.
The ice compactness dent is closely related to the vertical mixing there since the mixed layer
depth there is particularly deep. Its winter contour there also has a dent which is remarkably
similar to the observed ice compactness dent.
The 7-year mean spatial ice thickness distribution in Fig.(5.2.1b) is similar to the corresponding field from the ice-ocean-only model shown in Fig.(4.9.3a) with thicker ice in the central Arctic and thinner ice in the Greenland Sea as has already been found out from the transect analysis. What is worth mentioning is that the ice thickness increase is greater in the area off the Canadian Archipelago and the nearby Greenland coast. Two possible reasons may be behind this situation. One has been discussed before, that is, the exclusion of the vertical diffusion at the base of the mixed layer may result in more ice growth in the central Arctic. The other may be that the relatively thicker central Arctic ice converges to the Fram Strait as part of the transpolar ice stream and is more rapidly piled in that area. Therefore the ice thickness at that area seems closer to the observation shown in Fig.(4.9.3b).

Fig.(5.2.2) shows the 7-year mean ice thickness over the three different regions: the Greenland Sea, the Barents Sea and the Arctic Basin marked in Fig.(3.6.1). The results from the two main models and the mixed layer sensitivity study are shown in this figure. The two different mixed layer models generate more ice in the Arctic and the Barents Seas while less ice in the Greenland Sea. The different degrees of vertical mixing controlled by the decay coefficients do not make a significant difference in terms of ice conditions in the Arctic and Barents Seas while they do in the Greenland Sea, revealing that the vertical mixing in the Greenland Sea may be more important than in the other regions.

Fig.(5.2.3) shows the 7-year mean ice concentrations predicted by the three models over the three regions together with that observed. Compared to the observation, all the models over-estimate the ice concentration in the Greenland Sea although the ice-ocean-mixed-layer model and the sensitivity study bring the results closer to the observation. However, the areas covered by the ice with concentration greater than or equal to 0.2 show
that the simulated results are close to the observed in the Greenland Sea (Fig.(5.2.4)), indicating that the models create more ice with higher concentration than the observation. In addition, the models do not generate an ice cover in the Arctic and the Barents Sea that would advance as quickly as the observed ice cover after the summer shrinking.

5.3 Ocean Circulation

Since the vertical mixing mechanism introduced into the ice-ocean simulation changes to a certain extent the ocean's temperature and salinity structures, it would inevitably result in a change in ocean baroclinic circulation. Some changes are exhibited in Fig.(5.3.1) which shows the 1983 annual mean ocean surface velocity field. In this figure, one can still clearly find the major currents in the North Polar Region including the Alaskan coast current. Moreover, there is little change in velocity distribution in the GIN Sea and the Barents Sea. However, the surface circulation in the Arctic does have some noticeable changes. The most prominent change is the presence of one or two small gyres within the big Beaufort gyre. The gyre centered in the Wrangel Abyssal Plain in the Canadian Basin is most obvious. By examining the 7-year mean mixed layer depth contours of April shown in Fig.(5.1.1c) one finds that this gyre is also centered at an area which a mixed layer depth contour half circled, indicating that intensity of the vertical mixing may play a role. The center of the Beaufort Sea gyre calculated by the ice-ocean-only model is also shifted a little. Despite these changes, the overall ocean surface circulation pattern is more or less the same. This is also true for the pattern of the vertically integrated water transport, which is shown in Fig.(5.3.2). Only some limited local changes can be observed by carefully comparison this figure to Fig.(4.2.1). Other features of the ocean circulation such as the Atlantic water inflow are also not changed significantly, so they are not compared here.
Figure (5.1.1) 7-year mean monthly mixed layer depth fields. (a) October; (b) January; (c) April; (d) June. Contour interval is 20 m.
Figure (5.1.2) Transect (J37) – time plot of 7-year mean monthly mixed layer depth. Contour interval is 7 m. The dashed line is the ice edge defined as of ice compactness 0.2.
Figure (5.1.3) Transect (J37) ~ time plot of monthly mean mixed layer depth. (a) 1981; (b) 1982; (c) 1983. Contour interval is 7 m. The dashed line is the ice edge defined as of ice compactness 0.2.
Figure (5.1.3) (cont.)
Figure (5.1.3) (cont.)
Figure (5.1.4) Transect (J37) - time plot of 7-year mean oceanic heat flux. (a) Ice-ocean-mixed-layer model; (b) Ice-ocean-only model. Contour interval: 50 w/m².
Figure (5.1.4) (cont.)
Figure 5.1.5 Transect (J37) - time plot of 7-year mean ice edge in the Greenland Sea. Solid line: observation; Dashed line: ice-ocean-only model; Dot line: ice-ocean-mixed-layer model.
Figure (5.1.6) 7-year mean ice edge distributions in the Barents and GIN Seas. (a) February; (b) September. Solid line: observation; Dashed line: ice-ocean-mixed-layer model; Dot line: ice-ocean-only.
Figure (5.1.6) (cont.)
Figure (5.1.7) Transect (J37) - time plot of 7-year mean (a) friction velocity and (b) entrainment velocity. Contour interval: (a) 0.2 cm/s; (b) 4 m/day.
Figure (5.1.7) (cont.)
Figure (5.1.8) Transect (J37) - time plot of 7-year mean surface salt flux. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Unit: (centimeter of ice)/day-m² equivalent to 0.35 kg/day-m². Contour interval is 2. Solid lines stand for positive surface salt flux into the ocean, dashed lines for negative flux.
Figure (5.1.8) (cont.)
Figure (5.1.9) Transect (J37) ~ time plot of 7-year mean surface heat flux. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Unit: (centimeter of ice)/day-m² equivalent to 35 w/m². Contour interval is 2. Solid lines are for the ocean surface losing heat, dashed lines gaining heat.
SURFACE HEAT FLUX, TRANSECT (J37) - TIME, 7-YR MEAN, FIXED ML

GRID CELL NUMBER I

MONTH

CONTOUR FROM -0.0008 TO 76.000 CONTOUR INTERVAL OF 3.0000 PT(13,3) = 1.9078

Figure (5.1.9) (cont.)
Figure (5.1.10) Transect (J37) ~ time plot of 7-year mean surface (5 m) salinity (ppt). (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only.
Figure (5.1.10) (cont.)
Figure (5.1.11) Transect (J37) ~ time plot of 7-year mean surface (5 m) temperature. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Contour interval is 0.8 °C.
Figure (5.1.11) (cont.)
Figure (5.1.12) Transect (J63) — time plot of 7-year mean monthly mixed layer depth. Contour interval is 7 m.
Figure (5.1.13) Transect (J63) ~ time plot of 7-year mean friction velocity. Contour interval is 0.2 cm/s.
Figure (5.1.14) Transect (J63) — time plot of 7-year mean oceanic heat flux. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Contour interval is 50 w/m².
Figure (5.1.14) (cont.)
Figure (5.1.15) Transect (J63) – time plot of 7-year mean surface salt flux. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Unit: (centimeter of ice)/day-m² equivalent to 0.35 kg /day-m². Contour interval is 1. Solid lines stand for positive surface salt flux into the ocean, dashed lines for negative flux.
Figure (5.1.15) (cont.)
Figure 5.1.16 Transect (J63) – time plot of 7-year mean surface heat flux. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Unit: (centimeter of ice)/day-m² equivalent to 35 w/m². Contour interval is 1. Solid lines are for the ocean surface losing heat, dashed lines gaining heat.
SURFACE HEAT FLUX, TRANSECT (J63) - TIME, 7-YR MEAN, FIXED ML

GRID CELL NUMBER

MONTH

GRID CELL NUMBER

CONTOUR FROM -5.0000 TO 32.0000 CONTOUR INTERVAL OF 1.0000 PT12.34 1.6884

(b)

Figure (5.1.16) (cont.)
Figure (5.1.17) Transect (J63) - time plot of 7-year mean ice thickness. (a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Contour interval: 0.25 m.
Figure (5.1.17) (cont.)
Figure (5.1.18) Transect (J63) ~ time plot of 7-year mean surface (5 m) salinity (ppt).
(a) Ice-ocean-mixed-layer; (b) Ice-ocean-only. Contour interval is 1 ppt.
Figure (5.1.18) (cont.)
Figure (5.1.19) 7-year mean vertical profiles of temperature and salinity at selected locations, from the ice-ocean-mixed-layer model.
Figure (5.2.1) 7-year mean ice compactness and thickness fields.
(a) Compactness, contour interval: 0.1; (b) Thickness, contour interval: 0.3 m.
Figure (5.2.2) 7-year mean distributed ice thickness in the Greenland Sea, Barents Sea and Arctic Basin.
Figure (5.2.3) 7-year mean ice concentration in the Greenland Sea, Barents Sea and Arctic Basin. The unit is the fraction of ice concentration of the whole area.
Figure (5.2.4) 7-year mean fraction of area covered by ice with compactness greater than or equal to 0.2 in the Greenland Sea, Barents Sea and Arctic Basin.
Figure (5.3.1) Annual (1983) mean ocean surface velocity from the ice-ocean-mixed-layer model.
Figure (5.3.2) Monthly mean streamfunction from the ice-ocean-mixed-layer model in (a) February, 1983 and (b) September, 1983. Contour interval is 0.5 Sv.
Chapter 6. Interannual Characteristics of the Models

In the previous two chapters, the analyses of the numerical results are mostly focused on the characteristics of the simulated mean fields averaged over the whole integration period of 7 years. The mean fields effectively reveals the general patterns of the related processes in the ice-ocean system in a concise manner. However, the mean characteristics do not reflect the interannual variability for some distributions of quantities concerned. Therefore in this chapter, some aspects of the interannual characteristics of the models are analyzed. In the following Section 6.1, the interannual ice concentration variations are given and discussed, while in Section 6.2, the ice budget in the GIN Sea and its interannual variations are given with some discussion too.

6.1 Ice Concentration Variations

One of the interests of this thesis study is the effects of the ocean circulation and surface air temperature on the interannual variations of the ice cover in the North Polar Region. Consequently the main focus of this section is concentrated on the interannual variabilities of the ice concentration in the region under different situations. In order to identify these effects, comparisons are made mainly between the ice-ocean-only model and two additional sensitivity studies while the results from the ice-ocean-mixed-layer model and the prognostic sensitivity study are used when appropriate.

Following the same procedure for the aforementioned sensitivity studies, these two sensitivity studies are all integrated over the same 7-year period after the same 7-year spin-up. The first study is an ice-ocean-only simulation using a thermal forcing that is generated with 7-year (1979-85) mean daily varying modified Hansen temperatures. Needless to say, this study (hereafter called 'Mean-Thermal') is meant to show the effects
of the ocean on the interannual variations of the ice conditions. The second study is an ice-only simulation using the seven-year mean daily varying ocean heat flux and surface currents generated by the ice-ocean-only model. This study (hereafter called 'Ice-Only-Mean-Thermal') also uses the same thermal forcing as Mean-Thermal does. It is, however, meant to single out the effects of ocean circulation on the ice interannual variability.

The cycles of the seasonally averaged ice concentrations in various regions over the 7 years for Ice-Ocean-Only, Mean-Thermal, Ice-Only-Mean-Thermal and the observation are illustrated in Fig.(6.1.1). This figure demonstrates some of the features the same as discussed in Sections 4.9 and 5.2, such as the over-estimate of the Greenland Sea ice concentration and the slow ice advance after summer for all simulations, etc. In addition, Fig.(6.1.1) shows that Ice-Only-Mean-Thermal generally create more ice compared to Ice-Ocean-Only while the results from Mean-Thermal are closer to those of the main experiment.

The interannual characteristics of the ice concentration simulations from Ice-Ocean-Only, Mean-Thermal, Ice-Only-Mean-Thermal and the observation are illustrated in Fig.(6.1.2) for the three different regions. Plotted in this figure are the seasonal ice concentration deviations from 7-year averaged seasonal ice concentrations in each region. With this figure, one has a better idea of how each model behaves interannually by comparing the simulated variations with the observed variations. A more quantitative description of the results from the ice-ocean-only simulation and the two different sensitivity studies is given by correlation coefficients in Table (6.1.1) between the simulated interannual ice concentrations and the observed variations shown in Fig.(6.1.2).

As mentioned earlier, both sensitivity studies use thermal forcing that is calculated based on 7-year mean temperature, the only difference between them is that one is an ice-
ocean model under full influence of the ocean and the other is an ice-only model with 7-year mean ocean input fields of heat flux and ocean surface velocity. Therefore a comparison between them gives to some degree the ocean's effect. It seems that the existence of ocean circulation improves interannual variability of ice concentration in the Barents Sea, which can be seen by comparing the results from the two sensitivity studies in the figure and the table. This is thought to benefit from the interannual variations of the northern Atlantic water that flows and brings heat into that area and possibly affects the ice advection there too. On the other hand, the ocean does not seem to have a positive effect on the ice concentration variations in the Greenland Sea and hardly has it any effect in the Arctic Basin. However, the combination of the normal thermal forcing with the ocean circulation makes a considerable improvements in all areas, particularly in the Barents Sea although the ice there is out of phase in retreating and advancing before and after summer respectively. The relatively low performance of the main simulation in the Greenland Sea is probably due to a general over-estimate of ice concentration by the model. Nevertheless, using interannually varying surface air temperature is crucial in enhancing models' interannual variability as demonstrated by comparing the results shown in Fig.(6.1.2) and Table (6.1.1).

In Table (6.1.2) are additional correlation coefficients from the short-term prognostic run, the main ice-ocean-mixed-layer simulation and the mixed layer sensitivity study. The prognostic simulation has a slightly better performance in the Arctic compared to those from the ice-ocean-only run, while slightly worse in the Barents Sea. Considering that more Atlantic water flows into the Arctic through the Fram Strait and, hence, less into the Barents Sea, we conclude that more Atlantic water is necessary for a better simulation for both Arctic and Barents Seas. It should be interesting to conduct a long-term prognostic simulation with an ocean freely evolving to equilibrium and to see how the ocean circulation affects the ice extent variations.
The correlation coefficients from the ice-ocean-mixed-layer model indicate that the vertical mixing noticeably improves the interannual variability of the Greenland Sea ice concentration while has less impact in the other regions: the Arctic slightly improved and the Barents Sea slightly worsened. Also exhibited in Table(6.1.2) is that the more restricted vertical mixing in the mixed layer sensitivity study seems to hamper the model’s interannual variability in every region under consideration compared to the standard mixed layer model.

Table(6.1.1). Correlation coefficients between simulated and observed interannual ice concentrations shown in Fig.(6.1.2)

<table>
<thead>
<tr>
<th>Region</th>
<th>Ice-Ocean-Only</th>
<th>Mean-Thermal</th>
<th>Ice-Only-Mean-Thermal</th>
</tr>
</thead>
<tbody>
<tr>
<td>Greenland Sea</td>
<td>0.25</td>
<td>-0.01</td>
<td>0.08</td>
</tr>
<tr>
<td>Barents Sea</td>
<td>0.76</td>
<td>0.21</td>
<td>0.07</td>
</tr>
<tr>
<td>Arctic Sea</td>
<td>0.48</td>
<td>0.39</td>
<td>0.40</td>
</tr>
<tr>
<td>Whole Region</td>
<td>0.35</td>
<td>-0.09</td>
<td>-0.06</td>
</tr>
</tbody>
</table>

Table(6.1.2). Correlation coefficients from the prognostic study, the ice-ocean-mixed-layer simulation and the mixed layer sensitivity study.

<table>
<thead>
<tr>
<th>Region</th>
<th>Prognostic</th>
<th>Ice-Ocean-Mixed-Layer</th>
<th>Mixed Layer Sensitivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Greenland Sea</td>
<td>0.25</td>
<td>0.36</td>
<td>0.32</td>
</tr>
<tr>
<td>Barents Sea</td>
<td>0.73</td>
<td>0.74</td>
<td>0.72</td>
</tr>
<tr>
<td>Arctic Sea</td>
<td>0.51</td>
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<td>0.47</td>
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<tr>
<td>Whole Region</td>
<td>0.33</td>
<td>0.38</td>
<td>0.36</td>
</tr>
</tbody>
</table>
6.2 Ice Budget in the GIN Sea

Section 4.10 has seen considerable Arctic ice flowing into the Greenland Sea through the Fram Strait. In this section the effects of the ice advection on the GIN Sea ice distribution is examined. The ice advection in the GIN Sea is defined as the ice outflow at the Fram Strait and the Barents Sea Opening minus the outflow at transect J7 (J = 7 in Fig.(3.6.1)). Fig.(6.2.1) shows the simulated 7-year mean ice thickness, ice melting rate and the ice advection in the GIN Sea. As can be seen from this figure, the GIN Sea has only two winter months (January and February) that have net ice growth. The ice growth is so small that its contribution to the whole ice distribution is not significant at all. While in other months of the year, the ice there keeps melting, with the melting rate reaching the maximum in summer and getting smaller in the fall. This is why we observe much more negative surface salt flux distribution than positive distribution in the transect versus time plot of Fig.(5.1.8). The ice advection in the GIN Sea is the net contribution of the ice outflow at the Fram Strait, the Barents Sea Opening and the southern boundary (taken as J7) of the GIN Sea. The individual outflow at these three locations are shown in Fig.(6.2.2). As can be seen from this figure, the outflow has large seasonal and interannual variations for Fram Strait and the Barents Sea Opening, except for J7 which has very small outflow compared to others. What is interesting is that the outflow at the Barents Sea Opening is often significant, although smaller than that at the Fram Strait.

The melted ice can not be compensated by locally grown ice but by the ice advection contribution instead. As show in Fig.(6.2.1), the ice supply due to the ice advection is quite significant and is in all seasons, particularly in the fall. It is large enough to meet the demand of ice melting so as to keep the ice budget in the GIN Sea in a cyclic balance. This cyclic balance is indicated by the seasonal variations of the mean ice
thickness. The ice maintains a thickness at about 0.75 m in the middle of winter and decreases in summer due to the ice advection over taken by the melting rate, but increases again in the fall to its winters' level due to a larger ice advection. The balance is hence maintained. All this demonstrates that the ice advection is the dominant factor for the ice budget in the GIN Sea.

Fig.(6.2.3a) shows the interannual variations of both the seasonal mean ice thickness and the ice advection term. It is not hard to find out that the variations of the ice advection lag about one season behind the variations of the ice thickness. If the ice advection contribution were one season ahead, we would have rather good match between the ice advection and ice thickness, which is shown in Fig.(6.2.3b). This is understandable since the mean ice velocity in the Greenland Sea is about 0.1 m/s, it takes about 90 days for the ice travel over about 20 grid cells (800 km) to reach the central part of the Greenland Sea.

Fig.(6.2.4) shows the deviations of the ice thickness and advection from their 7-year mean, based on Fig.(6.2.3b) with one season shifted for the ice advection. The curves in the figure further exhibit a good agreement between the two quantities in terms of interannual variations. The deviation correlation coefficient of the two curves is 0.63, indicating a good correlation between the ice advection and thickness. Whereas if the ice advection is not shifted by one season, the corresponding correlation coefficient is 0.46. However, it should be pointed out that when the ice advection is compared to the observed ice edge, the correlation is not encouraging, only around 0.2. This is not surprising since the correlation between the simulated ice concentration and the observed one in the Greenland Sea is not high either. However, correlation along particular transects may be much higher.
Figure (6.1.1) Seasonal variations of ice concentration from Ice-Ocean-Only, Mean-Thermal, Ice-Only-Mean-Thermal and the observation in the Arctic, Greenland and Barents Seas from 1979 to 1985. The unit is the fraction of ice concentration of the whole area.
Figure (6.1.2) Deviation of the seasonal ice concentration from the 7-year seasonal mean in the Arctic, Greenland and Barents Seas from 1979 to 1985. The solid line represents observed data and the dashed line simulated results. The unit is fraction of ice coverage of the whole area.
Figure (6.2.1) 7-year mean distributed ice thickness, ice advection and ice melting rate in the GIN Sea.
Ice outflow (Sv)

At Fram Strait
At Barents Sea Opening
At transect J7

Figure (6.2.2) Ice outflow. Solid line: at the Fram Strait; Dash-dot line: at the Barents Sea Opening; Dot line: at transect J7.
Figure (6.2.3) Seasonal variations of distributed ice thickness and ice advection in the Greenland Sea. (a) Actual variations; (b) The curve of ice advection is one season shifted toward the right of the horizontal axis.
Figure (6.2.4) Distributed ice thickness and ice advection deviations from their 7-year mean values in the Greenland Sea.
Chapter 7. Summary and Conclusions

This research consists of two main objectives. One is to study the characteristics of the ice-ocean circulation in the North Polar oceans with a coupled high-resolution ice-ocean numerical model. The other is the embedding of a mixed layer model into a general ice-ocean circulation model to examine effects of introducing the physics of vertical mixing on the performance of the ice-ocean numerical simulation of the ice-ocean system. To achieve the first objective, a high-resolution (40 km horizontal resolution and 21 vertical levels of increased thicknesses with depth) ice-ocean-only model was constructed based on the Hibler and Bryan (1987) 160 km resolution ice-ocean model with a number of modifications among which are the introduction of the ocean open boundary conditions and time and area dependent precipitation. The work for realizing the second objective involved development of the Kraus and Turner mixed layer model and efforts to embed the mixed layer model into the high resolution ice-ocean-only model before it was tested in a form of a simplified salinity mixed layer model. The ice-ocean-only model and the fully coupled ice-ocean-mixed-layer model were used for the 7-year (1979-85) period simulations of the North Polar ice-ocean system, in conjunction with a series of sensitivity studies. The major results from all the main simulations and sensitivity studies are summarized in Section 7.1 with some conclusions given. In Section 7.2, some suggestions are made for future work.

7.1 Summary and Conclusions

- Both the simulated ocean surface velocity and the vertically integrated water volume transport show a consistent existence of a two-gyre circulation in the Arctic Basin. One is in the Beaufort Sea with a clockwise ocean circulation, while the other is in the Eurasian Basin with an anti-clockwise circulation. By integrated vorticity analysis, it is revealed that the existence of the two-gyre pattern in the Arctic Basin is closely related to
the bottom topography which steers ocean flows by following the rule of vorticity conservation in which the bottom pressure and lateral friction induced vorticities are dominant terms. In addition, the coupling of the mixed layer model has some effect on the ocean circulation in the Arctic such as the one or two small local circulation gyres within the big Beaufort gyre. This is thought to be attributable to the spatial differences in the vertical mixing which modifies the temperature and salinity structure and hence results in some change in baroclinic circulation which may be the dominant part of the ocean circulation in the Arctic. However, in the GIN Sea and the Barents Sea, the mixing effects on the circulation do not seem noticeable, indicating that the limited change of surface temperature and salinity distribution in these areas do not significantly influence the baroclinic flow. Despite some changes in the Arctic, the same overall characteristics of the ocean circulation as those by the ice-ocean-only model are preserved.

• The high-resolution models used here demonstrate some advantage in resolving those major currents as well as some local currents in the North Polar Region. In particular, both the ice-ocean-only and the ice-ocean-mixed-layer models all realistically create the Beaufort Sea undercurrent. This local current has not been simulated by previous Arctic numerical studies because those studies used relatively crude resolutions that can not resolve such a narrow current. However, high-resolution models do not necessarily guarantee to generate this current. The influence of the ice cover on the ocean through modifying the ocean surface stress input with the ice interaction force and the properly introduced Bering Strait open boundary conditions are important in maintaining the Alaskan coast flow pattern.

• The Atlantic water inflow proves to be rather robust, having a mean value about 1 Sv with some interannual variations. The mean value is close to the estimate based on observation (Rudels, 1987) while differs from othe reports (Section (2.3)). This Atlantic
water penetrates deep into the Arctic before finally losing its identity in terms of temperature and salinity and flows out of the Arctic in the sublayer. The introduction of the open boundary conditions at the Bering Strait, Denmark Strait and Faroe-Shetland passage is also a factor that enhances the Atlantic water inflow. If these open boundaries are closed, the inflow is cut by about half. In addition, the short-term prognostic simulation tends to allow some more Atlantic water to flow into the Arctic.

- The simulated ocean surface salinity fields by the ice-ocean-only model seem reasonable by a comparison between the simulated and the observed fields in summers. The ice-ocean-mixed-layer model results in a relatively fresher water at the ice edges in the Beaufort Sea and the Greenland Sea because this model agitates more vertical mixing at ice edges than in other areas. Therefore near the ice edges it results in more fresh water at the surface layer.

- The treatment of the open boundaries works well in that it allows proper heat exchange between the whole region under consideration and the outside oceans so that the overall heat budget in the region is basically in balance. However, the short-term prognostic sensitivity study seems to have a better adaptivity and be able to maintain a better heat balance.

- The oceanic heat flux analysis shows that the convective overturning process is always the principal contributor of the oceanic heat flux in all the three different regions in the North Polar Region. Whereas other oceanic heat flux components play a different role from region to region. In the ice covered area such as the most part of the Arctic Ocean, the heat flux components due to the lateral and vertical advection basically cancel each other out, the lateral diffusion term is negligibly small and the vertical diffusion process is the second largest term although not big. In the marginal ice zone, the open water areas and the
areas where the ice cover is subject to great seasonal changes such as the Barents Sea, the lateral advection is quite significant and the net contribution due to the lateral and vertical advection is not trivial because of less uniform horizontal temperature fields. The lateral advection term can be important since it compensates the heat loss at the ice edge where additional heat is needed to melt ice. In the marginal ice zone, the vertical diffusion processes often transfer the heat in the mixed layer to the water below in the summer due to the effects of the atmospheric warming.

- The simulated ice drift in the whole region is in fair agreement with the observed buoy drift statistically. The robust ocean circulation resulting from the use of the lower horizontal viscosity and diffusion allowed by the high resolution improves the ice drift. The inflow at the Bering Strait may play a role in improving the ice drift in the Chukchi region, while the transpolar stream and the East Greenland Current enhance the ice drift from the Arctic to the Greenland Sea. The ice outflow at the Fram Strait is quite significant and in good agreement with the corresponding observational estimates.

- The ice outflow at the Fram Strait is crucial to maintain a balanced ice budget in the GIN Sea because it is the major ice source of this region. The ice budget in the region is basically in a balance between the ice melting and the ice advection in the GIN Sea, while the local ice growth is very small. The Fram Strait is the major ice contributor to the GIN Sea, and the outflow at the Barents Sea Opening is often significant too. The interannual variations of the ice conditions in the GIN Sea are closely related to the interannual variations of the ice advection with the latter lagging about one season behind the former since the ice at the Fram Strait needs about 90 days to reach the central part of the Greenland Sea.
• Compared to observations, the models generally over-predict the ice concentration in the GIN Sea. However, the simulated areas covered by the ice with concentration larger than or equal to 0.2 are close to that observed. This indicates that the models over-predict the ice that is of higher concentration and perform reasonably well in predicting ice of lower concentration. The use of the mixed layer model substantially improves the prediction of the ice edge in the GIN Sea by allowing particularly intensive vertical mixing at the ice edge and effectively stops the ice from further extending to the open water. However, since the dominant factor for ice formation is still the atmospheric conditions, the degree of the vertical mixing allowed in this study can not bring the predicted ice concentration to a good agreement with the observation. In the Arctic Basin and the Barents Sea, the vertical mixing does not significantly change the ice concentration fields. The reasons behind this are different for different regions. For the Arctic Ocean, the mixed layer depth is limited to about 70 m using the region-independent decay coefficients, hence the oceanic heat flux resulting from the vertical mixing is very limited. In the Barents Sea, more mixing occurs than in the Arctic, but the temperature profile in the ocean surface layer near the normally observed ice edge does not allow significantly more oceanic heat flux either. In addition, the Barents Sea ice extent simulated by all the models does not advance fast enough after the summer compared to observations.

• The ice-ocean-mixed-layer model creates thicker ice than the ice-ocean-only model. The thicker ice distribution seems closer to observations. It is believed that this is because the vertical eddy diffusion at the mixed layer base is not included in the model so that less exchange is allowed between the mixed layer and the layers below when the mixed layer is retreating and no entrainment occurs. From the analysis of the heat flux components, although small, the diffusion processes in the Arctic are still the consistent contributor for the upward oceanic heat flux second only to the overturning component. The overturning term in the Arctic ice covered area is much smaller than its counterpart in
the GIN sea marginal ice zone. This allows the vertical diffusion term in the Arctic plays more prominent role than that in the GIN Sea. Exclusion of the diffusion induced heat flux therefore results in an additional ice growth.

- The interannual variations of surface air temperature dictate the ice concentration variations, while the ocean's effects are only noticeable in the Barents Sea where the incoming Atlantic water may play a role because it carries considerable heat with it and may also affect the ice advection. The embedded standard mixed layer model considerably improves the interannual variability of the ice concentration in the Greenland Sea compared to the ice-ocean-only model, while the mixed layer sensitivity study with less mixing makes relatively smaller improvement. In the Arctic, improvement by the standard mixed layer model is not very significant statistically. The difference the mixed layer model makes in the Barents Sea is also not statistically significant.

- The simulated mixed layer depth, as a measure of the degree of the vertical mixing, is so distributed that it is generally shallower in the Arctic Basin and deeper in the Barents Sea and, particularly, in the GIN Sea using the same energy decay parameterization. This is because in the often ice covered Arctic, the mixed layer is close to a salinity mixed layer without the surface cooling and heating being directly involved, while, in the GIN Sea for example, both the salt flux and heat flux are at play to drive the mixed layer, particularly at the ice edge where more entrainment occurs. In addition, the relatively larger wind energy input also plays a role to promote more mixing in the GIN Sea. Moreover, it seems the mixing process at the Greenland Sea ice edge is different from that at the Beaufort Sea ice edge. At the Greenland Sea ice edge, the agitated entrainment brings up enough deep oceanic heat to keep the nearby ice melting so that the mixing and the melting are going on at the same time in a band along the ice edge. At the Beaufort Sea
ice edge, however, the mixing does not bring up enough heat to result in a melting condition. Therefore there does not exist a parallel mixing and melting.

7.2 Suggestions for Future Work

The models used in this study are diagnostic ones subject to the constraints by the climatological temperature and salinity fields. The short-term prognostic study has already demonstrated that it can better adapt itself to the surface and lateral boundary conditions so that more Atlantic water is brought into the Arctic from the Fram Strait and better heat budget is maintained there. Considering this feature, a long-term prognostic simulation with a horizontally uniform but stratified initial temperature and salinity fields is suggested with a high-resolution grid like the one used here. In this way the influence of the observed temperature and salinity are excluded and the ocean can freely evolve to its equilibrium after a many-year integration period. It should be interesting to carry out such prognostic ice-ocean simulations with a conventional fixed depth mixed layer and a variable depth mixed layer and compare the differences they make.

The delay of ice advance in the Arctic may indicate that the two-level ice model of Hibler (1979) may be slow in response to the atmospheric cooling in terms of ice growth. An incorporation of a multi-level ice thickness model (see Hibler, 1980a; Flato, 1991) may be able to improve the seasonal variability of the ice-ocean models.

From the experience of using the Kraus and Turner mixed layer model in this study, the author would suggest some more considerations in the mixed layer parameterization. In considering the different conditions in the different regions in the North Polar Oceans, a region-dependent turbulence decay parameterization may be useful. For example, if relatively large decay coefficients are chosen for the GIN Sea, more mixing
is expected there. The diffusion calculation at the mixed layer base may be retained too so that there will be tracer exchange between the mixed layer and the layers below. This will prevent the water in the summer mixed layer getting too fresh.
Appendix 1. Imposition of ocean open boundary (OB) conditions

The following measures are taken to introduce open boundary conditions:

1. 30 day relaxation is imposed at every level at OBs and in their vicinity.
2. \( \frac{\partial T}{\partial n} = \frac{\partial S}{\partial n} = 0 \), where \( n \) is the direction perpendicular to OBs.
3. The distribution of vertically integrated transport streamfunction along OB is specified with possible observationally estimated water transport data as mentioned before. Therefore the vertically averaged velocity or external mode of velocity (see Semtner, 1986) at OB is calculated from the specified streamfunction:

\[
\begin{align*}
\bar{u} &= \frac{1}{H} \int_{-H}^{0} Udz = -\frac{1}{Ha} \frac{\partial \psi}{\partial \varphi}, \\
\bar{v} &= \frac{1}{H} \int_{-H}^{0} Vdz = \frac{1}{Hacos \varphi} \frac{\partial \psi}{\partial \lambda},
\end{align*}
\]

where \( H \) is ocean depth, \( U \) and \( V \) are horizontal ocean velocity components, \( a \) is the earth radius, \( \varphi \) is latitude, \( \lambda \) is longitude and \( \psi \) is the streamfunction.

4. A simplified version of the internal mode of velocity or velocity deviation from the external mode can be obtained from geostrophic equations:

\[
\begin{align*}
\bar{u}' &= -\frac{1}{f_p a} \frac{\partial p}{\partial \varphi}, \\
\bar{v}' &= \frac{1}{f_p acos \varphi} \frac{\partial p}{\partial \lambda},
\end{align*}
\]

\[ p(z) = p_s + \int_z^0 g \rho dz' = p_s + p_0, \]

where \( p_s \) is ocean surface pressure and \( p_0 \) is hydrostatic pressure component.

Let

\[
\begin{align*}
\bar{u}^* &= -\frac{1}{f_p a} \frac{\partial p_0}{\partial \varphi}, \\
\bar{v}^* &= \frac{1}{f_p acos \varphi} \frac{\partial p_0}{\partial \lambda},
\end{align*}
\]
then
\[ u'(z) = u^*(z) - \frac{1}{H} \int_{-H}^{0} u^*(z)dz', \]
\[ v'(z) = v^*(z) - \frac{1}{H} \int_{-H}^{0} v^*(z)dz', \]

or numerically,
\[ u_k' = u_k^* - \frac{1}{H} \sum_{k=1}^{kz} u_k^* \Delta z_k, \]
\[ v_k' = v_k^* - \frac{1}{H} \sum_{k=1}^{kz} v_k^* \Delta z_k. \]

Finally \( U = \bar{u} + u', \quad V = \bar{v} + v', \) at OBs.
Appendix 2. Tests of One-Dimensional Salinity Mixed layer Model

A one-dimensional salinity mixed layer is created and tested in order to obtain a general idea how the full mixed layer model described in Sections 3.3 and 3.4 will behave once embedded into the ocean model. The formulation of the salinity mixed layer model is based on that described in Section 3.3, but its one-dimensional temperature is always set to be a constant, freezing point -1.96°C. Therefore the temperature profile has little influence on the formation of the mixed layer, while the salinity plays a dominant role on it. This situation can be found in areas where ice cover exists in all seasons. This means that the salinity mixed layer is a special case of the full model's applications. If the salinity mixed layer model works, we can reasonably believe that the full mixed layer model works too.

The tests include two aspects. One is to see if the mixed layer model can have a good response to the initial and boundary conditions, and forcing fields and maintain a reasonable prediction of the related mixed layer quantities. The other is to see if the embedding procedure can maintain a conservation for those conservative properties. In the salinity model, the salt should be conserved. For this purpose, a one-dimensional salinity ocean model of 21 fixed levels is created into which the two-level mixed layer is to be embedded. These 21 levels are identical to those used in the ice-ocean model. The initial salinity profile for the salinity ocean model is the same as that of mixed layer. The embedding procedure is identical to what is described in Section 3.4, except that the advective contribution to the change of mixed layer depth is excluded because of its one-dimensionality.

Most of the parameters used in the model are from Lemke and Manley (1984), including the surface stress induced by ice keels' stirring, surface buoyancy flux estimated
by specifying ice melting rate, some of the decay coefficients, etc. According to Lemke and Manley the surface stress is calculated by

$$|\tau| = \rho \, C_w \, |u|^2$$

where $C_w = 0.0034$ is drag coefficient. The ice melting rate is expressed as

$$F(t) = a_m \sin \left( \pi t / t_m \right) \quad 0 \leq t \leq t_m$$

$$F(t) = a_f \sin \left[ \pi (t + t_f - t_m) / t_f \right] \quad t_m \leq t \leq t_m + t_f$$

where $a_m = H \pi / 2 t_m$, $a_f = A \pi / 2 t_f$, and $H$ is the thickness of the seasonally frozen of melted ice. Here $H$ uses only one of Lemke and Manley's values, 0.85 m/year. The curve of ice melting rate versus time is shown in Fig.(A2.1). This same melting rate is used for 5 different tests whose related information is listed in Table (A.2.1). As can be seen from the table, these tests are made by changing one condition or parameter at a time. All the test results are shown in Fig.(A2.2), including the mixed layer salinity, depth and total salt. Each plot in the figure corresponds to one test case.

Table (A2.1) Parameters used in different test cases represented by plots in Fig.(A2.2)

<table>
<thead>
<tr>
<th>Test</th>
<th>$\Delta T$</th>
<th>$u$</th>
<th>$h_w$</th>
<th>$h_c$</th>
<th>$S_{mi}$</th>
<th>$S(-h_m)i$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Test1, Fig.(A2.2a)</td>
<td>day/24</td>
<td>6</td>
<td>9.67</td>
<td>19.34</td>
<td>30.4</td>
<td>32.8</td>
</tr>
<tr>
<td>Test2, Fig.(A2.2b)</td>
<td>day/4</td>
<td>6</td>
<td>9.67</td>
<td>19.34</td>
<td>30.4</td>
<td>32.8</td>
</tr>
<tr>
<td>Test3, Fig.(A2.2c)</td>
<td>day/4</td>
<td>6</td>
<td>9.67</td>
<td>31.14</td>
<td>30.4</td>
<td>32.8</td>
</tr>
<tr>
<td>Test4, Fig.(A2.2d)</td>
<td>day/4</td>
<td>12</td>
<td>9.67</td>
<td>19.34</td>
<td>30.4</td>
<td>32.8</td>
</tr>
<tr>
<td>Test5, Fig.(A2.2e)</td>
<td>day/4</td>
<td>6</td>
<td>9.67</td>
<td>19.34</td>
<td>30.4</td>
<td>31.8</td>
</tr>
</tbody>
</table>
In Table (A2.1), $\Delta T$ is timestep, $u$ is ice velocity in cm/s, $S_{mi}$ and $S_{(-hm)i}$ are initial mixed layer and its base salinities in ppt and decay coefficients $h_w$ and $h_c$ are set here to be constants in meters.

The forcing that drives the salinity mixed layer model consists of the surface buoyancy flux and the surface stress. Since the model uses a constant ice velocity, the surface stress or the mechanical energy input to the model is always a constant. So the evolution of the mixed layer is closely related to the variations of the buoyancy flux or ice melting or freezing.

Fig.(A2.2a) representing Test 1 shows the results from a test of small timestep with standard parameters which will be changed one at a time for other tests. The calculated $S_m$ varies between the initial conditions $S_{mi}$ and $S_{(-hm)i}$ according to the ice melting rate. $h_{mix}$ is changing in a reasonable range and the total salt, a salt summation over the 21 levels of the 1-D ocean model, is very well conserved. Another feature is that all the quantities are cyclically varying in response to a periodic ice melting. This feature is desirable for a long-term simulation because the results will not drift away.

Test 2 (Fig.(A2.2b)) uses a much larger timestep, the same as that in the ice-ocean-mixed-layer model, while other parameters are not changed. The results are very similar to those in Fig.(A2.2a), indicating that the model is not sensitive to time step and relatively large time step may be used. Test 3 uses a larger $h_c$ to see how the decay coefficient changes the results. As can be seen from Fig.(A2.1c), $h_{mix}$ is greatly increased and $S_m$ is accordingly increased too, although still within the range of $S_{mi}$ and $S_{(-hm)i}$. Test 4 uses a larger ice velocity to calculate mechanical energy input. The results shown in Fig.(A2.2d) indicate that $h_{mix}$ becomes less shallower when the mixed layer retreats because of a larger kinematic energy input. In response to the less shallower mixed layer,
the variation range of $S_m$ is narrowed. Note also that $h_{mix}$ does not get significantly deeper with the larger energy input than other comparable cases because the influence of the decay coefficient $h_w$. Test 5 uses a different initial salinity profile from Test 1 with $S(-h_M)i$ reduced by 1 ppt. A comparison between Test 5 in Fig.(A2.2e) and Test 2 shows no difference except that $S_m$ from Test 5 is decreased a little less than 1 ppt. In all tests the total salt is the same because of the same surface salt flux determined by the same ice melting rate.

The analysis of those test results leads to a conclusion that the salinity mixed layer model is well behaved and the embedding procedure can guarantee conservation of related quantities both in the mixed layer and ocean models. This conclusion therefore lays an assuring basis for the full mixed layer model to be coupled into the 3-D ocean model.
Figure (A2.1) Ice melting rate.

Figure (A2.2) Mixed layer salinity, mixed layer depth, and total salt. 
(a) Test 1; (b) Test 2; (c) Test 3; (d) Test 4; (e) Test 5.
Figure (A2.2) (cont.)
Figure (A2.2) (cont.)
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