Pergamon

Marine Pollution Bulletin Vol. 43, Nos. 1–6, pp. 61–70, 2001

Printed in Great Britain
0025-326X/01 $ - see front matter

PII: S0025-326X(00)00234-4

Two Regimes of the Arctic's Circulation from Ocean Models with Ice and Contaminants

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A two-dimensional barotropic, coupled, ocean–ice model with a space resolution of 55.5 km and driven by atmospheric forces, river run-off, and sea-level slope between the Pacific and the Arctic Oceans, has been used to simulate the vertically averaged currents and ice drift in the Arctic Ocean. Results from 43 years of numerical simulations of water and ice motions demonstrate that two wind-driven circulation regimes are possible in the Arctic, a cyclonic and an anti-cyclonic circulation. These two regimes appear to alternate at 5–7 year intervals with the 10–15 year period. It is important to pollution studies to understand which circulation regime prevails at any time. It is anticipated that 1995 is a year with a cyclonic regime, and during this cyclonic phase and possibly during past cyclonic regimes as well, pollutants may reach the Alaskan shelf. The regime shifts demonstrated in this paper are fundamentally important to understanding the Arctic's general circulation and particularly important for estimating pollution transport.

Introduction

The variability of the general circulation of the Arctic is not as well known as it is in many other oceans. Knowledge of the circulation is important to understanding the potential for pollutants and radionuclides to be carried to locations distant from a source. The thrust of the present work is to examine the major forcing mechanisms that drive the Arctic's circulation, and with the use of a basin scale model, identify the relative strengths of each forcing mechanism, and the Arctic's response. The potential for pollutants to be dispersed depends on the total circulation comprised of the flow resulting from each forcing mechanism.

Two different historical views describe factors determining the movement of water and ice in the Arctic Ocean. Shokal'skii (1940), Gordienko and Karelin (1945), and Buinitskii (1951, 1958) assumed that the surface circulation in the Arctic Ocean was mainly driven by the inflow of Atlantic water through Fram Strait, producing an excess of water in the basin and causing an outflow of surface water and ice along the shorter routes to the Greenland Sea. Data from vessel drifts in the Eurasian subbasin indicated a movement of surface water from the Eurasin subbasin to Fram Strait. Nansen (1902), who discovered waters of Atlantic origin in the Arctic Ocean, mentioned that water exchange between the Arctic Ocean and the Greenland and Norwegian Seas was caused by differences of water temperature and salinity, e.g., he assumed a thermohaline circulation. These scientists believed the Arctic Ocean water and ice circulation was mainly thermohaline in nature and closely related to the world's ocean circulation.

Subsequently, some authorities (Zubov and Somov, 1940; Shuleikin, 1941; Shirshov, 1944; Treshnikov, 1959) adopted a different view on the nature of the water circulation, namely, that the inflow of Atlantic waters into the Arctic Basin was caused principally by the outflow of surface water to the Greenland Sea under the influence of the existing wind system. This point of view was confirmed by experiments undertaken by Gudkovich and Nikiforov (1965) with a hydraulic model simulating the system of wind-driven currents in the Arctic Ocean. The experiments showed winds to be a principal factor in forming the system of permanent currents in the Arctic Basin. The winds of the Arctic anti-cyclone generate anti-cyclonic currents in the Amerasian subbasin. The wind system in the Eurasian subbasin and at the periphery of the Arctic anti-cyclone induces a wide Trans-Arctic current originating in the Chukchi Sea and emerging through Fram Strait. Simulating the inflow of water as the only factor driving the circulation did not produce a circulation pattern similar to that actually observed. In accordance with this concept, Gudkovich (1961a,b) explained the seasonal and year-to-year variation in the circulation of surface water and ice by a suitable rearrangement of the baric fields.

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To confirm Gudkovich's conclusions, a series of two-dimensional numerical models were run (Fel'zenbaum, 1958; Campbell, 1965; Galt, 1973; Ponomarev and Fel'zenbaum, 1975; Proshutinsky, 1988, 1993), where wind fields were the major factor driving circulation in the Arctic Ocean. It is interesting that Galt (1973) found that inflow/outflow forcing and bottom topography both play a more important role in the circulation than does wind forcing. However, in the research of Treshnikov and Baranov (1972), it was demonstrated that the major features of the water circulation may be obtained by means of diagnostic calculations according to density fields. These authors stated that 'we could not evaluate the role of individual factors in the formation of currents because the observed field of masses already contains the combined (and still indistinguishable) effects of wind, baroclinicity, and bottom relief'. A similar conclusion could be made according to the results of the diagnostic model by Ponomarev (1977). Prognostic models (Semtner, 1976) testify to the significant role of the irregularity of the field of mass in the formation of the three-dimensional water circulation.

A series of different model runs was presented by Semtner (1987), Fleming and Semtner (1991), Häkkinen and Mellor (1992), Hibler and Bryan (1986), Riedlinger and Preller (1991), Piacsek *et al.* (1991). All of these models and numerical experiments were directed to developing new models and to describing better observed features of water and ice dynamics without special attention to the variability of the oceanographic regime of the Arctic Ocean. Recent papers by Häkkinen (1993, 1995) were dedicated to modeling and explain such phenomena as deep convection in the Greenland Sea and the Great Salinity anomaly. Her investigations demonstrated again the significance of atmospheric circulation in the coupled ice–ocean system in the Arctic.

In this paper we assume that the general circulation of the Arctic can be characterized by four broad responses to different forcings. The response from the observed sea-level slope between the North Pacific ocean and the Arctic is northward flow through Bering Strait and out through Fram Strait, with the interior response yet to be clearly described. A second response is due to forcing by the atmospheric winds and pressure fields that drive both a baroclinic and a barotropic response. In this paper we treat the Arctic as a barotropic fluid. A third response is due to the inflow from rivers. Associated with river inflow is a fourth response, the total thermohaline flow, due to the introduction of fluid at a boundary.

We have modeled the response due to sea-level slope, winds and river input. The relative strength of the circulation regimes driven by these forces can be compared by examining the model results. A comparison with the thermohaline flow is possible by comparing the results of modeled thermohaline flow by Proshutinsky (1988, 1993) or Polyakov (1994). Such a comparison shows that the mean annual wind-driven circulation in the Arctic is at least the same order of magnitude or larger, in terms of maximum observed current speeds, as the thermohaline forcing. Thus, the circulation regimes presented in this study may be observable in the ocean, depending on the Arctic's linearity.

For this paper we have examined the way the circulation in the Arctic responds due to the sea-level slope, the winds (derived from pressure), and the input from rivers. We identify two wind-driven regimes, a cyclonic and an anti-cyclonic regime, which alternate over a ten to fifteen year period. When estimating pathways that pollutants may follow, it is important to distinguish between these two regimes, and identify which phase is present.

That the position of the sea surface may serve as an integral index of processes of interaction between the atmosphere and the ocean was an idea first stated by Duvanin (1951). This result was confirmed later by different authors depending on the problem they solved, e.g., 'the position of the sea surface is an integral indicator of most processes and phenomena related to the formation of the hydrological regime' (Galerkin, 1961), or 'the only indicator of synoptic variability of hydrological conditions in arctic seas, for which corresponding field data exist, consists of the sea-level heights' (Krutskikh, 1978) Year-to-year and seasonal fluctuations of the ocean level in this sense may be viewed as an indicator of large-scale interaction between the atmosphere and ocean. In particular, the sea surface can evidently be used as an indicator of large-scale circulation of water and ice of the Arctic Ocean and as an indicator of the regime driving pollution transport. We use sea level in our paper as an integral index to describe variability of the Arctic Ocean circulation.

This paper is organized in the following manner. Basic equations and boundary conditions are described in Model Equations and Boundary Conditions. The data base is discussed in Data Base. The model results are given in Results and the final section contains a summary and discussion.

Model Equations and Boundary Conditions

We shall use a system of equations of motion and continuity written in the vector notation and in the stereographic polar coordinate system

$$\frac{d\mathbf{U}}{dt} + f\mathbf{k} \times \mathbf{U} = -gDm\nabla\zeta + N_h m^2 \nabla^2 \mathbf{U} + \frac{c\mathbf{T}_i + (1-c)\mathbf{T}_s - \mathbf{T}_b}{\rho}, \quad (1)$$

$$\frac{\partial\zeta}{\partial t} = -m^2\nabla(\mathbf{U}/m), \quad (2)$$

and the equations of motion and continuity for ice,

$$\frac{d\mathbf{u}_i}{dt} + f\mathbf{k} \times \mathbf{u}_i = -gm\nabla\zeta + \frac{\mathbf{T}_{is} - \mathbf{T}_i}{\rho_i h_i} + \mathbf{F}_i, \quad (3)$$

$$\frac{\partial c}{\partial t} = -m^2 \nabla(\mathbf{u}_i c/m). \tag{4}$$

Here, x, y are the lateral coordinates, with their origin at the North Pole; t is the time; ζ denotes free surface elevation; $\mathbf{U}$ is a vector of volume transport with components U, V along x- and y-directions; $\mathbf{u}_i$ is a vector of ice velocity; $\mathbf{T}_i$ is a vector of ice stress between water and ice; $\mathbf{T}_{is}$ is a vector of ice stress between atmosphere and ice; $\mathbf{T}_b$ is a vector of bottom stress; $\mathbf{T}_s$ is a vector of water stress between atmosphere and water; $\mathbf{F}_i$ is a vector of internal ice forces; ρ_i is ice density; ρ is water density; c is ice concentration, $0 \leqslant c \leqslant 1$; h_i is ice thickness; N_h is horizontal eddy viscosity $(= 5 \times 10^9\ \mathrm{cm}^2/\mathrm{s})$; D is the total depth $(= H + \zeta)$; f denotes the Coriolis parameter; $\mathbf{k}$ is unit vector along vertical direction and m denotes map coefficient.

$$\frac{\mathrm{d}}{\mathrm{d}t} = \frac{\partial}{\partial t} + m\left[\frac{U}{D}\frac{\partial}{\partial x} + \frac{V}{D}\frac{\partial}{\partial y}\right].$$

In the above operator only the first-order nonlinear terms are retained.

The stereographic polar coordinate system used in Eqs. (1)–(4) is very close to the rectangular system of coordinates. The difference is due to the map coefficient m. It describes map correction from spherical projection to polar stereographic projection. Its value changes from 1 at 90°N to 1.071 at 60°N.

The interaction of water and atmosphere is described by

$$\mathbf{T}_s = \rho_\alpha R_\alpha \mathbf{W}|\mathbf{W}|, \tag{5}$$

where $\mathbf{W}$ is a vector of surface wind and ρ_α and R_α are air density and friction coefficient between air and water, respectively.

The surface wind was determined from geostrophic relationships with consideration of the transitional coefficient C_α and angle of deviation of surface wind from the geostrophic direction A_α. Implemented in the model calculations is the algorithm in which (Ashik *et al.*, 1989)

$$C_\alpha = 0.7 \quad \text{if } \mathbf{W} < 15\ \mathrm{m/s},$$
$$A_\alpha = 30^\circ \quad \text{if } \mathbf{W} < 15\ \mathrm{m/s},$$
$$C_\alpha = 0.8 \quad \text{if } \mathbf{W} > 15\ \mathrm{m/s},$$
$$A_\alpha = 20^\circ \quad \text{if } \mathbf{W} > 15\ \mathrm{m/s}.$$

The air–water drag coefficient was estimated as a function of the wind speed as follows:

$$R_\alpha = (1.1 + 0.04|\mathbf{W}|) \times 10^{-3}.$$

Bottom stress is described by

$$\mathbf{T}_b = \rho R_b \frac{|\mathbf{U}|\mathbf{U}}{D^2}, \tag{6}$$

where R_b is a bottom friction coefficient $(= 2.6 \times 10^{-3})$.

The interaction of ice and atmosphere is described by

$$\mathbf{T}_s = \rho_\alpha R_{i\alpha} \mathbf{u}_i |\mathbf{u}_i|. \tag{7}$$

In a first approximation the ice–atmosphere drag coefficient $R_{i\alpha}$ was equal to air–water friction coefficient, R_α.

The interaction of ice and water is described by

$$\mathbf{T}_i = \rho R_i \left|\mathbf{u}_i - \frac{\mathbf{U}}{D}\right| \left(\mathbf{u}_i - \frac{\mathbf{U}}{D}\right). \tag{8}$$

The ice–water drag coefficient R_i $(= 5.5 \times 10^{-3})$ was estimated by McPhee (1986).

To describe internal ice forces the nonlinear viscous constitutive law proposed by Rothrock (1975) is taken

$$\mathbf{F}_i = \eta m^2 \nabla^2 \mathbf{u}_i + \Lambda m^2 \nabla(\nabla \mathbf{u}_i) - m \nabla p. \tag{9}$$

Here

$$\nabla(\nabla \mathbf{u}_i) = \mathrm{grad}(\mathrm{div}\,\mathbf{u}_i).$$

Rothrock (1975) suggested that the tensile stress in ice is negligible compared to compressive stress. Pressure (p) in (9) is given by:

$$p = -A_p m \nabla \mathbf{u}_i \quad \text{if } \nabla \mathbf{u}_i < 0,$$

$$p = 0 \quad \text{if } \nabla \mathbf{u}_i \geq 0. \tag{10}$$

In the above formulas both bulk (Λ) and shear (η) viscosity coefficients are taken to be equal; A_p is the ice pressure coefficient.

The magnitude of the frictional coefficients used in the above equations should result in numerical stability and reasonable reproduction of the turbulent processes in the water and ice. Horizontal eddy viscosity $N_h = 5 \times 10^9\ \mathrm{cm}^2\ \mathrm{s}^{-1}$ is taken close to the threshold of numerical stability $N_h = 5 \times 10^8\ \mathrm{cm}^2\ \mathrm{s}^{-1}$ (Kowalik, 1981). A variable level of the vertically generated turbulence is controlled through the ice–water stress. In the ice-free areas this term is set to zero and in the fast ice regions the friction is expressed in a manner similar to the bottom stress.

To obtain a unique solution to the above system in the domain of integration it is sufficient to specify initial and boundary conditions.

Initially, the dependent variables in the integration domain are taken as 0:

$$\zeta(x, y)_{t=0} = 0, \tag{11}$$

$$\mathbf{U}(x, y)_{t=0} = 0, \tag{12}$$

$$\mathbf{u}_i(x, y)_{t=0} = 0. \tag{13}$$

Along the solid boundary (S) we assume a no-slip condition for water transport and ice velocity

$$\mathbf{U}(x, y, t)_S = 0, \tag{14}$$

$$\mathbf{u}_i(x, y, t)_S = 0. \tag{15}$$

On the open boundary (O) of the domain, often sea level or water transport is known. If sea level is known we proceed as follows: in the vicinity of the open boundary (along the first line parallel to the open boundary), the linear hyperbolic problem is solved (horizontal friction and advective terms in (1) are omitted). This procedure yields a unique solution for the volume transport with sea level defined at the open boundary (e.g., Kowalik and Murty, 1993). When transport is specified along the line parallel to the open boundary, the solution process for the full set of equations (parabolic problem) can be extended into the integration domain. With the above restriction, the open boundary conditions pertinent to Eqs. (1) and (2) can be stated as follows:

$$\zeta_O = \zeta(x, y, t), \tag{16}$$

$$\mathbf{U}(x, y, t)_O = \mathbf{U}(x, y, t). \tag{17}$$

Very often water transport along the open boundary is not known. In this case it is useful and enough to prescribe a radiation condition along the open boundary as

$$\mathbf{U} = \pm\zeta_{in}(gH)^{-2}, \tag{18}$$

where ζ_{in} is a sea level inside computational domain.

For ice cover at the open boundary the following conditions are prescribed:

$$\frac{\partial \mathbf{u}_i}{\partial n} = 0, \tag{19}$$

$$\frac{\partial c}{\partial n} = 0, \tag{20}$$

where n is a normal to the open boundary.

Along the open boundaries in the North Pacific we prescribe sea level (2 m), and along the open boundaries in the North Atlantic we prescribe the radiation condition (18). These conditions are used to reproduce sea slope between the Pacific and Atlantic Oceans. This sea slope is responsible for permanent water inflow from the Bering Sea into the Arctic Ocean and stable circulation in the northern Bering Sea and the Chukchi Sea (Proshutinsky, 1986) and apparently plays a significant role in formation of the Trans-Arctic current. To include river run-off we prescribe water transport (**U** or **V**) for every river included in the model (Mackenzie, Kolyma, Indigirka, Lena, Khatanga, Yenisei, Ob and Pechora).

Data Base

To cover the Arctic Ocean with a 55.5 km spatial grid step we use bathymetry of the Arctic Ocean with a 14-km spatial grid step prepared by Kowalik and Proshutinsky (1994).

The atmospheric forcing fields are obtained from the CD-ROM of the National Center for Atmospheric Research (NCAR). We use sea-level pressure analysis for the daily wind forcing case (NCAR, 1990).

The model results are supported by comparing observations to information obtained by different authors in numerical experiments with the Arctic Ocean models. Three kinds of data bases are available to perform this comparison:

(a) Sea-level observations are available from about 15 sites in the Kara, Laptev, East-Siberian and Chukchi Seas. These are mainly statistical characteristics of sea-level regime which were published mainly in Russian scientific journals.

(b) Coordinates of 335 drifted buoys for the period 1979–1988 (Throndike *et al.*, 1982).

(c) We use widely different indirect information related to the hydrological regime of the Arctic Ocean, such as dynamic heights relative to the 200 dbar surface obtained by Treshnikov *et al.* (1976) for the period 1946–1965; water temperature and salinity record in the Faroe-Shetland Canyon for a period of 1946–1973 published by Yanes (1977), air temperature record at the meteorological stations in the Greenland, Norwegian and Barents Seas, published by Dementiev (1991), and water transport, salinity and temperature behavior in the Fram Strait simulated by Häkkinen (1995).

Results

Numerical simulations have been completed to assess the effects of different forcing. In the first experiment we tried to understand the role of Bering Strait inflow and a river run-off in the Arctic Ocean circulation. Antonov (1958, 1968) had explained observed surface circulation and the Arctic Ocean hydrological regime mainly by river run-off variability. Fig. 1(a) shows the surface circulation in the Arctic Ocean for the first experiment. The sea slope between the Pacific and the Arctic Oceans, and the river run-off, drive the well-known, quasi-permanent circulation in the northern Bering Sea, Bering Strait, Chukchi Sea, Laptev and Kara Seas. They also indicate the location of the core of the Trans-Arctic and East-Greenland currents. Magnitudes of the currents are proportionally related to the values of sea-level slope between the Pacific and Arctic Oceans and the river run-off. The general pattern of these currents is related to the ratio between water transport through the Bering strait and the water inflow from the Siberian rivers. Two extreme situations are shown in Fig. 1(b) and (c), where surface circulation is presented for situations when sea slope is zero and only river run-off forms the circulation, and when river run-off is zero and circulation in the ocean is driven only by sea slope between the two oceans.

In the second experiment, both the ocean and ice models were initialized from rest on 1 January 1946, and run for 43 years until 31 December 1988, using the NCAR daily surface atmospheric pressure, the simulated surface winds computed from the pressure data, mean annual river run-off and a sea-level slope between the Pacific and Arctic Ocean which provided reasonable inflow of Pacific waters of about 1 Sv. Barotropic model spin up takes about 4 months if sea-level slope and river run-off are included into the model.

Year-to-year variability

The model shows that wind is a principal factor responsible for the year-to-year variability in the circula-

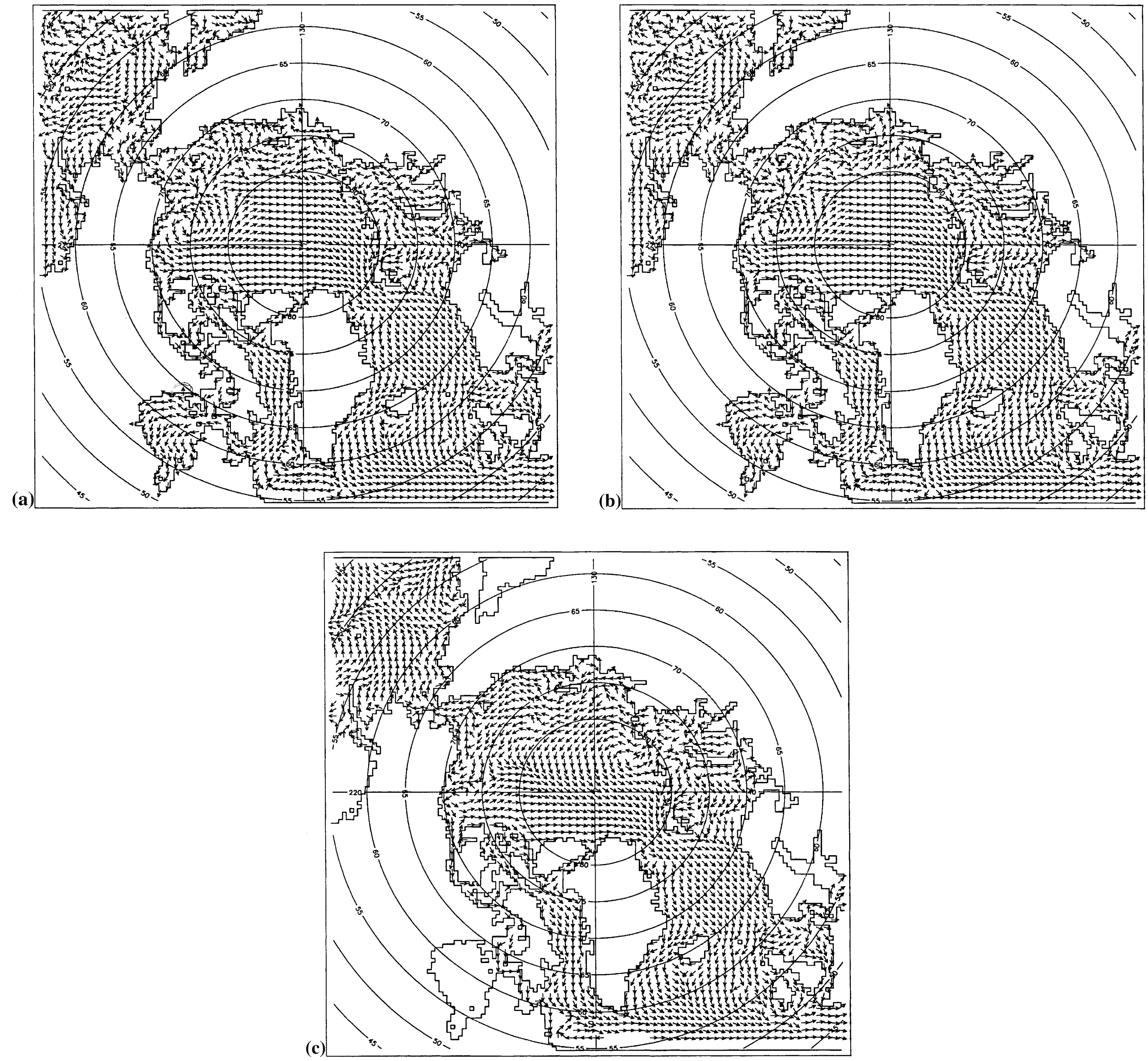

Fig. 1 Surface currents from the barotropic model with prescribed: (a) river run-off and sea slope between the Pacific and Arctic Oceans; (b) only sea slope between the Pacific and Arctic Oceans; (c) only river run-off. Note that with river run-off forcing flow through Bering Strait is southward.

tion of ice and vertically integrated flow. The time for the ocean to respond to changes in thermohaline forcing is generally considered to be too slow to explain shorter term variability of a few years or less.

Two regimes describe the modeled barotropic circulation and ice circulation in the Arctic Ocean. One regime is characterized by anti-cyclonic circulation as observed in the modeled central Arctic Ocean during 1946–1952, 1958–1964, 1972–1979 (Figs. 2 and 3). A second regime is characterized by cyclonic circulation as observed in the model during 1952–1957, 1964–1971 and 1980–1986 (Figs. 4 and 5). Regime shifts between cyclonic and anti-cyclonic flow occur every 5–8 years with the period of oscillation of about 10–15 years.

To determine the time scale of the regime shifts we examined the sea level slope at the center of the Arctic Basin as a measure of cyclonicity. The sea slope was calculated as the difference between the sea level at the center of a closed circulation cell and the sea level at the periphery of this circulation divided by the distance between the center and periphery. Positive slopes corresponds to a raised sea surface and anti-cyclonic ice and water circulation (Fig. 3). Negative sea-level slopes in the center of the Arctic Basin correspond to cyclonic ice and water circulation (Fig. 5). The absolute value of the sea-level gradient in the center of the Arctic Basin may serve as an index of circulation intensity (higher values correspond to a higher velocities in cyclonic or

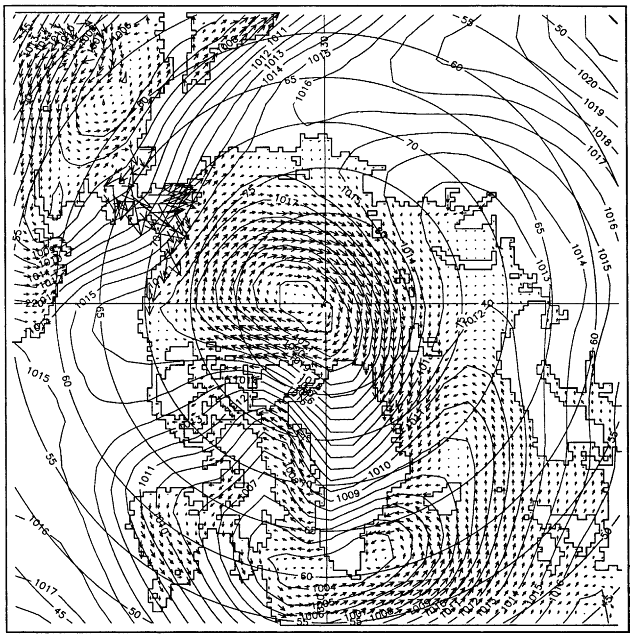

Fig. 2 Surface currents and surface atmospheric pressure distribution for the typical year with anti-cyclonic circulation.

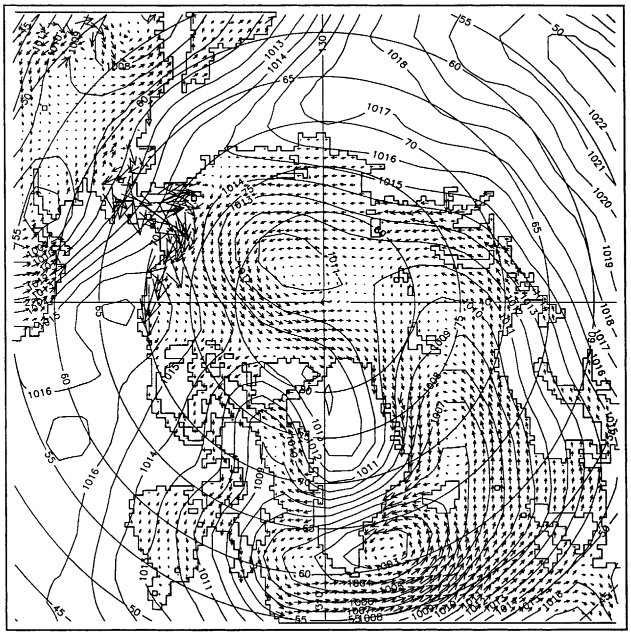

Fig. 4 Surface currents and surface atmospheric pressure distribution for the typical year with cyclonic circulation.

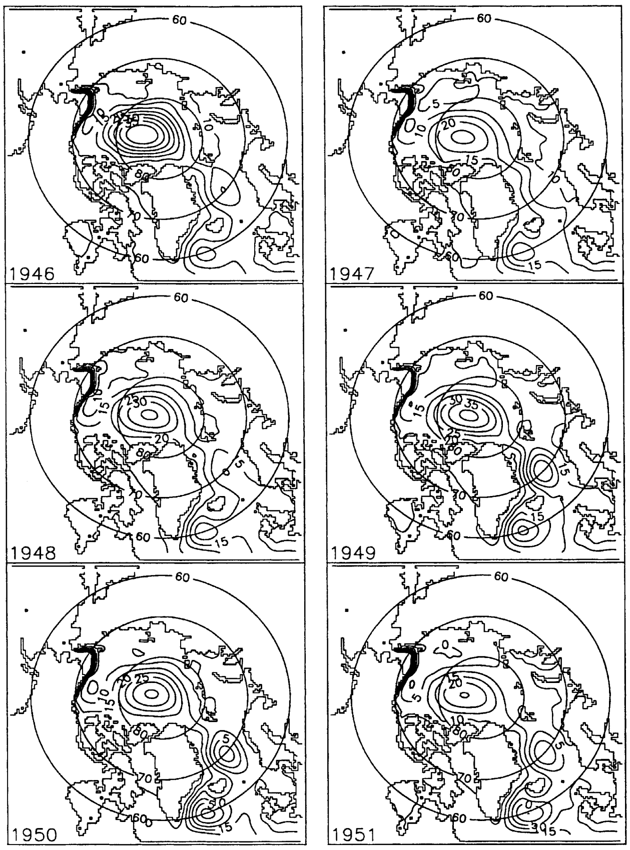

Fig. 3 Sea-level heights for the years of anti-cyclonic circulation. High heights are in the center of the Arctic Basin and low heights are along the coast. There is a cyclonic circulations in the Greenland–Islandic–Norwegian Seas (GIN) and in the North Atlantic.

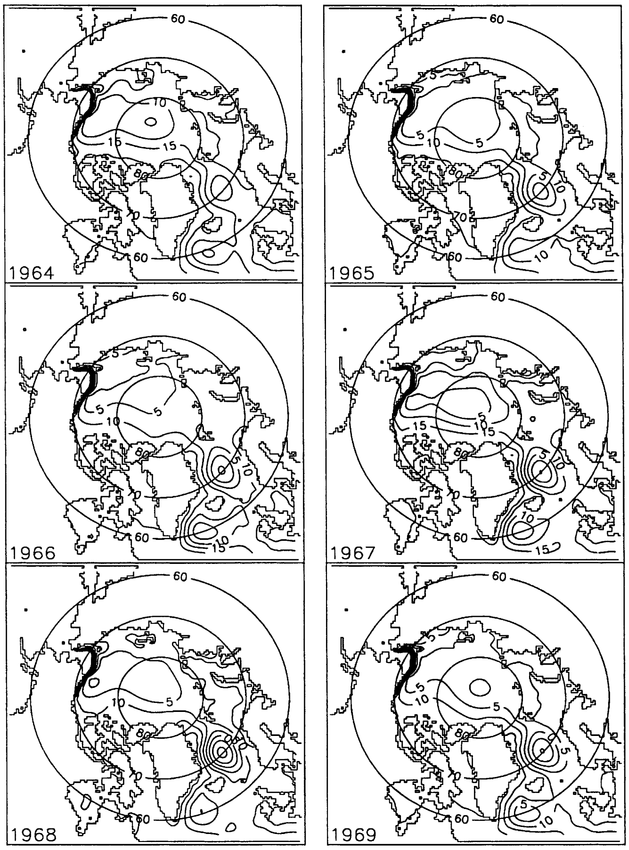

Fig. 5 Sea-level heights for the years of cyclonic circulation. Low heights are in the center of the Arctic Basin and high heights are along the coast. There is a cyclonic circulations in the Greenland–Islandic–Norwegian Seas (GIN) and in the North Atlantic.

anti-cyclonic circulation). Fig. 6 shows a time-series of the gradient of this slope and identifies the circulation and strength of the two regimes.

We have found references to the existence of only a 5–7 year oscillatory regimes in the Russian meteorological and oceanographic literature. This is probably

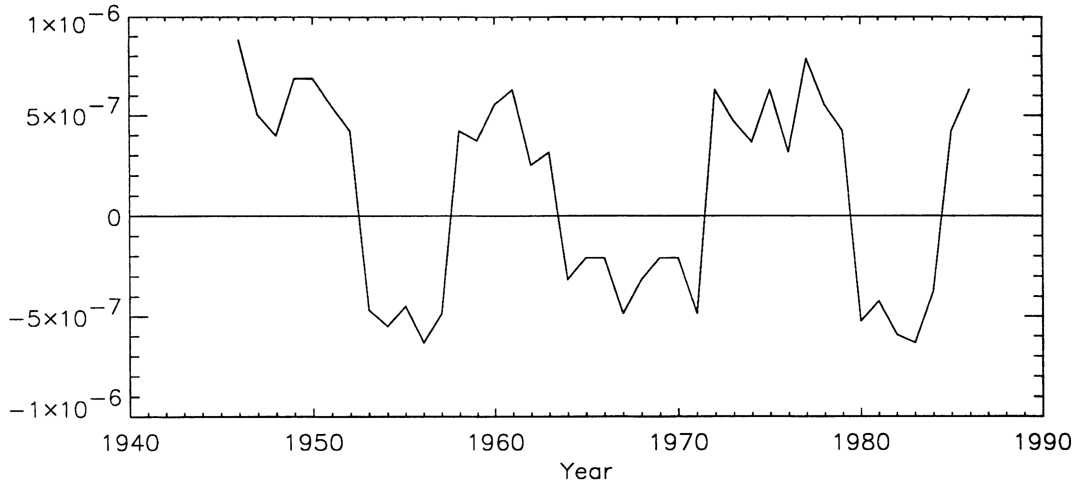

Fig. 6 Gradients of sea level in the central Arctic Basin. Positive gradients indicate anti-cyclonic circulation and negative gradients indicate cyclonic circulation. In 1995–1999 we expect a cyclonic regime to prevail.

because the time series of analyzed data were too short. Karklin (1977) found that the ice conditions and ice drift in the Arctic Seas has a periodicity of 6–7-years. A 7–8-year period was found in the temporal variability of dynamical heights relative to 200 dbar for the period 1949–1973 (Treshnikov *et al.*, 1976). An investigation of the Arctic Ocean water budget (Treshnikov and Baranov, 1976) revealed a 2–3 and a 5–7 year period. Our results show that 2–3 year and a 5–7 year variability of the circulation exists, too, on a background of cyclonic or anti-cyclonic circulation (this is usually 2 year variability of intensity of the circulation without changing the sign of rotation). Year-to-year variability of temperature and salinity in the Faroe-Shetland Canyon shows periods of 5–7 years as well (Yanes, 1977). Air surface temperature at the stations in the Greenland and Norwegian Seas has the same period 5–7 year oscillations (Dement'ev, 1991). It is interesting that in the most recent paper of Häkkinen (1995), one can find the same 10–15 years variability in the simulated potential temperature and salinity for the top 100 m, and in the change in heat and salt content of the upper 2000 m in the Greenland Sea on her Figs. 6(a) and (c) and 7(a), respectively.

The presence of two wind-driven circulation regimes has several important implications. During years of anti-cyclonic ice and water circulation, the core of the Trans-Arctic current is intensified and shifted toward Siberia, enhancing outflow of ice from the East-Siberian, Laptev and Kara Seas. The ice transported through Fram Strait is expected to be thinner because of its origin in these coastal seas, implying decreased transport of fresh water into the Greenland Sea. During years with cyclonic circulation, the core of the Trans-Arctic current is shifted toward Canada and Greenland, enhancing the flow of old ice into the Greenland Sea. According to this ice circulation, there should be more transport of ice and fresh water through Fram Strait at this time because the ice transported from the Canadian sector of the Arctic is

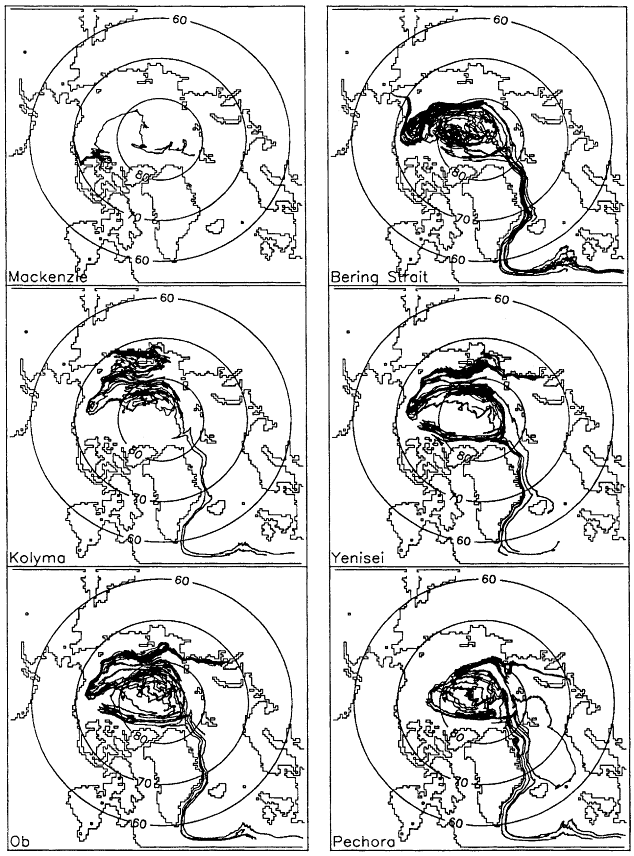

Fig. 7 Trajectories of water parcels released at different regions in 1946. Period of simulations is 1946–1988.

thicker than in the Siberian Seas by a factor of 2 or more. Navigation in the Kara Sea is likely to be better in the years with anti-cyclonic type of circulation and worse in the cyclonic years when ice tends to block the Vilkitskogo Strait.

The annual cycle

After analyzing the 'average' annual cycle we have found that the spatial–temporal structure of the surface heights of the Arctic Ocean is determined mainly by the surface atmospheric circulation: by the intensity and location of the Arctic anti-cyclone and Icelandic depression. In the cold season from November to April there is significant development of high pressure in the Arctic with a rise in the sea level in the center of the anti-cyclonic system of circulation and a decrease in sea level along the coast. In the second half of the year with a weakening of the arctic atmospheric anti-cyclone, the resulting barotropic anti-cyclonic ocean circulation weakens, with sea level in the center of circulation dropping while there is a sea level increase along the coastline of the arctic seas.

In the Norwegian and Greenland Seas, under influence of the system of winds and the baric field of the Icelandic depression from November to April, a vast cyclonic water circulation develops. In the warm season the atmospheric depression fills and wind-driven circulation weakens.

The annual cycle has been examined with respect to the two general wind-driven circulation regimes. We have averaged into monthly bins data from the anti-cyclonic circulation regime and averaged into monthly bins data from the cyclonic circulation regime for ice and surface motions to look at the seasonal cycle of each of the two circulation regimes. In the anti-cyclonic regime, strong anti-cyclonic circulation dominates for about 10 months from October to July. Thus, in this case, the winter regime of annual circulation prevails. In the years with a cyclonic regime, the summer circulation, characterized by weak cyclonic circulation and low ice and water transport through the Fram Strait, dominates the annual cycle.

Pollutant transport

Understanding the two circulation regimes is useful for investigating the temporal and spatial variability of ice, water and pollutant transport in the Arctic Ocean. For example, in Fig. 7 we present results based on the history of trajectories of water markers released monthly beginning in 1946 at Bering Strait and at different river mouth locations for the period 1946–1986. Water markers moving with vertically averaged velocities have comparatively stable trajectories following bathymetric features and are consistent with the two circulation regimes. For example, parcels released at Bering Strait have trajectories with both cyclonic and anti-cyclonic rotation. Eighty percent reach Fram Strait and 20% were entrained into the circulation of the central Arctic Basin. Parcels originating in the Kolyma River mouth move to the east or west as a function of the regime type. Pollutants from the Kolymya River can reach the coastal waters off Alaska.

Ice trajectories are more variable because of the direct influence of wind. Tracks of the ice motion can be found everywhere in the Arctic basin with a maximum concentration in the Beaufort Sea and along the core of the Trans-Arctic Current.

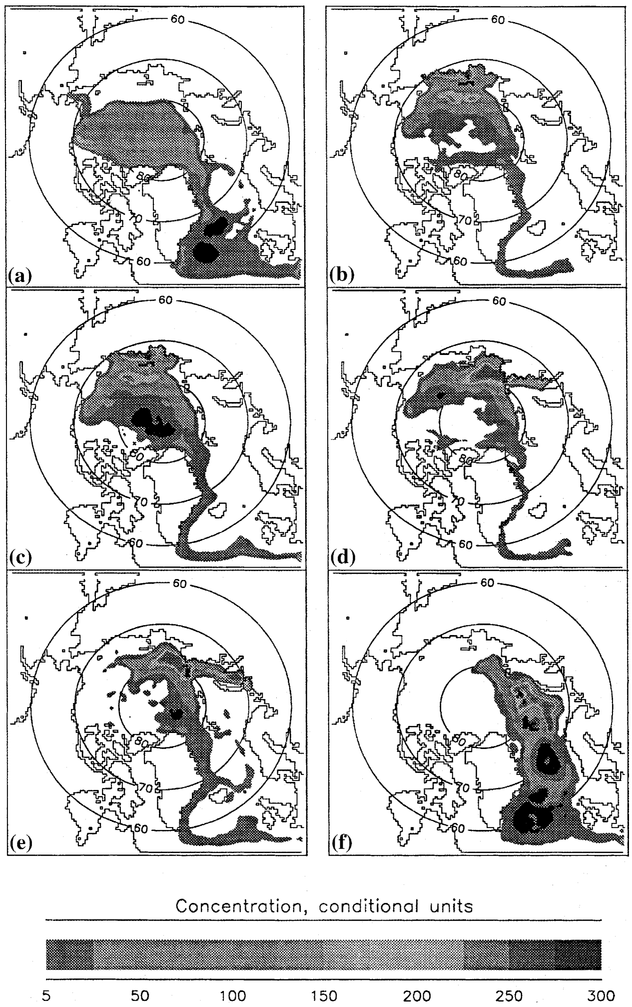

Fig. 8 Conditional concentration of pollution from the different regions during 1946–1988. Note higher concentrations stretching from the Alaskan shelf across the Arctic Basin to Fram Strait; the higher concentrations in the Beaufort Gyre and along the east Siberian shelf.

Fig. 8 shows concentration of pollutants from different sources. Pollutant concentration was simulated using water and ice trajectories, and we assumed that at every time step each 'pollutant particle' leaves an amount of its pollution that is simply summed every model time step. Summing gives a 'concentration' and a contour of this is presented. To have a high concentration of pollution in a region either, (1) many particles pass over the region, or, (2) a particle remains in the region for an extended period. Analysis of the figures shows that there are a few regions with high potential pollutant concentration: the continental slope of the Siberian Seas, the core of Trans-Arctic current, regions with prevailing local cyclonic circulation, and, of course, regions near a source of pollution.

Summary and Conclusions

We have constructed a barotropic, two-dimensional, coupled, ice–ocean model with resolution of about 55 km, obtained solutions to different kinds of forcing, and analyzed data of vertically averaged currents and ice motion in the Arctic Ocean for period 1946–1988. The model results show that:

1. Two wind-driven circulation regimes are possible in the Arctic, a cyclonic and an anti-cyclonic circulation. These appear to alternate at intervals of 5–7 years with the period of 10–15 years. It is important to pollution studies to understand which circulation regime prevails. It is anticipated that we are now in a cyclonic regime, and during this phase, pollutants can indeed reach the Alaskan shelf.
2. Many measurements of temperature and salinity distribution were done in the Arctic Ocean between 1972 and 1978 when an anti-cyclonic circulation regime prevailed. These data were used in forming the Levitus data base as well as for modeling the thermohaline circulation of the Arctic Ocean. Our knowledge about the Arctic Ocean circulation is thus based primarily on data from years with prevailing anti-cyclonic atmospheric and ocean circulation. That a cyclonic circulation regime also exists is supported by examining the temporal variability of dynamic heights in the Arctic Ocean (Treshnikov *et al.*, 1976), temperature and salinity oscillations in the Faroe-Shetland Strait (Yanes, 1977), and air temperature variations in the Norwegian and Greenland Seas (Dement'ev, 1991).
3. The seasonal circulation has two regimes as well. During winter an anti-cyclonic regime prevails and during summer this anti-cyclonic circulation weakens or perhaps even reverses so that weak cyclonic circulation prevails.
4. We anticipate that this research will continue, and we will include the atmospheric conditions for 1989–1994. Following this, we plan to use a three-dimensional model for simulating the Arctic Ocean circulation including the thermohaline effects and ice thermodynamics and further explore the two circulation regimes discussed here.

The regime shifts demonstrated in this paper are fundamentally important to understanding the Arctic's general circulation and particularly for estimating pollution transport. But because the majority of Arctic data reflect the anti-cyclonic (1972–1978) regime, and because the recent Trans-Arctic effort may have sampled during an anti-cyclonic regime as well, it is central to Arctic science that this modeling work be fully tested and a major future field effort sample during the next cyclonic regime (1995–1999).

We would like to acknowledge support for this work from the Office of Naval Research, under grant NOO14-94-1-0570. Numerical computations were carried out using CRAY Y-MP located at the University of Alaska, Fairbanks.

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