MODELS OF AIR-ICE-SEA INTERACTION

Motoyoshi Ikeda
Faculty of Environmental Earth Science, Hokkaido University, North 10 West 5, Sapporo, 060-0810, JAPAN

Keywords: polar ocean, sea ice, heat flux, mixed layer, ocean current, brine rejection, convection, polynya, climate, global warming

Contents

Summary

Sea ice forms on the sea surface once the atmosphere has cooled sufficiently to lower the surface of the water to its freezing point. It often acts as an obstacle for our activity, while we feel a strong connection to nature when we view the ice-covered ocean. From a physics point of view, sea ice provides unique properties such as plastic behavior and salt rejection from freezing sea water during sea ice formation. These characteristics are explained first, followed by a description of ice-ocean interactions considering both thermodynamic and dynamic aspects. Sea ice is a distinct indicator for climate change, as its growth/decay is strongly linked to a colder/warmer climate. Therefore, we pay attention to its variability, both in thickness and extent, while particularly noting changes to sea ice cover due to the current warming phase, and discuss whether the sudden sea ice

reduction in the Arctic might be a consequence of global warming or indicator of year-to-year variability. Overall information on the ice-covered ocean, together with an introduction to air-sea-ice modeling, are key components of the information presented herein.

1. Introduction

Sea ice is a concern for both off-shore oil operations in the Polar Regions and transportation through an ice-covered ocean. The role sea ice will play under future climate change scenarios is uncertain. Since sea ice floats on the sea surface, interactions between sea ice and ocean are naturally included in sea ice modeling. The atmosphere applies a significant dynamic force to a sea ice cover in the form of wind stress, while feedback from the ice-ocean system to the atmosphere has a significant effect on both local weather and global-scale climate. Here we first introduce thermodynamic and dynamic concepts with respect to sea ice, and then describe ice-ocean coupling and air-ice-ocean interactions.

In the last thirty to forty years, since the 1960s and 1970s, the Arctic has experienced a reduction in sea ice which is considered to be an indicator of global warming. On the other hand, climate variability has various time scales; hence we should not take all reduction in sea ice cover to be a consequence of global warming. In particular, climate variability at time scales of up to a few years can be significant, which could lead to a false alarm if we attribute the current reduction in ice cover to anthropogenic climate change alone. This document discusses recent climate variability and considers future projections into the 21st century.

2. Ice Modeling

2.1. Principles of Sea Ice Modeling

Sea-ice models have been discussed in numerous papers (e.g. Hibler 1979, 1980). The mechanisms included in the models are categorized into the areas of thermodynamics and dynamics. The thermodynamics are mainly controlled by heat flux between the atmosphere and the ice-ocean system, in addition to solar radiation. A positive heat flux from water to sea ice immediately initiates ice melting, and water cooled below its freezing point yields ice formation. Sea ice dynamics are typically driven by wind forcing, while ice-water stress, which is usually in balance with air-ice stress, typically counters this forcing. Under a high ice concentration near a coastal boundary, the internal stress of the ice becomes a crucial parameter and also acts to resist wind forcing. A minor addition to these forcing terms also comes from the Coriolis force and gravity associated with sea surface slope.

2.2. Thermodynamics of Sea Ice

Let us consider an idealistic sea ice cover with a uniform thickness. A schematic view is shown in Figure1. Heat flux occurs from the ice surface to the atmosphere and from the underlying seawater to the bottom of the sea ice. The surface heat flux consists of incoming short-wave radiation, net long-wave radiation, and sensible and latent heat fluxes. Sea ice is often covered by snow, which tends to reflect solar radiation. Conductive heat flux occurs in the ice due to the temperature gradient between the ocean and atmosphere. As the bottom of the ice cover is kept nearly at the in situ freezing point of the underlying seawater (about -1.8^{o}C), the conductive heat flux is inversely proportional to ice thickness. Sea ice typically grows (melts) in response to upward (downward) heat flux. Ice formation, which occurs at the interface of the ice cover and the ocean, occurs faster under thinner ice due to a relatively high upward conductive heat flux, and conversely occurs more slowly under a thicker ice cover where the conductive heat flux is reduced. Snow cover has an insulative effect which results in a small thermal conductive coefficient and hence tends to reduce the rate of ice formation. An interested reader is directed to a typical modeling paper on sea ice thermodynamics (Maykut and Untersteiner, 1971) for further information on this process.

Each component in the surface heat flux is now described more in detail. The components are parameterized from bulk formulae given by Smith and Dobson (1984). Incoming short-wave radiation increases with higher solar angle and is zero during night.

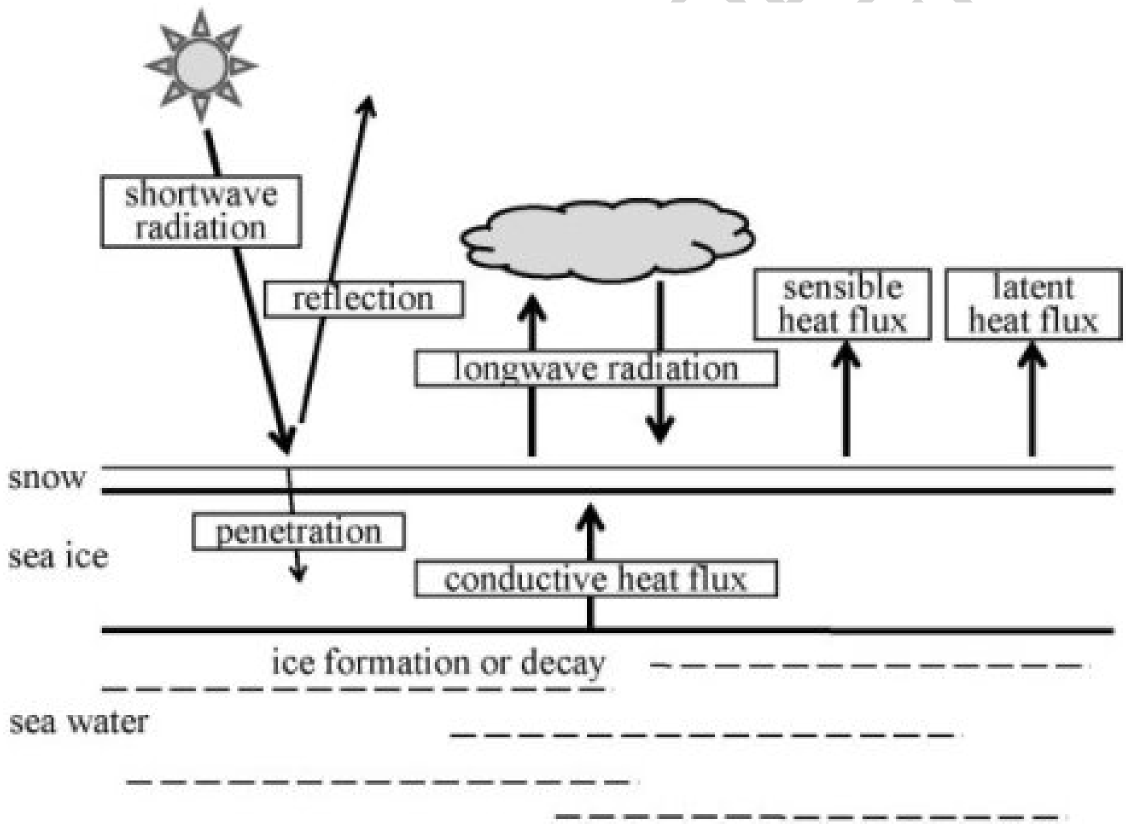

Figure 1. Schematic view of heat fluxes around sea ice.

The values of albedo (a measure of reflectivity) are about 0.05 for water, and in a range of 0.5 for bare ice to 0.8 for snow. The effective solar heat flux reaches a mean value in summer of 100 Wm^{-2}. Long-wave radiation emanating upward from the ice surface is partly cancelled by downward radiation from the atmosphere. Clouds contribute to the downward radiation and reduce net long-wave radiation from the ice. The long-wave radiation is typically stable throughout the year.

Sensible and latent heat fluxes are expressed as follows:

$$SH = \rho_a C_a C_h (T_i - T_a) U_a \tag{1.1}$$

$$LH = L C_e (q_s - q_a) U_a \tag{1.2}$$

The upward fluxes are defined to be positive in both equations. In (1.1), ρ_a (1.3 kg m^{-3}) is air density, C_a (10^3 J kg^{-1} K^{-1}) is the specific heat of air, C_h is the sensible heat flux coefficient, T_a is air temperature, T_i is ice surface temperature, and U_a is wind speed at 10 m above the surface. In (1.2), L (2.5·10^6 J kg^{-1}) is the latent heat of water vapor, C_e is the latent heat flux coefficient, q_s is the saturation water vapor density, and q_a is the water vapor density in air. Both fluxes increase under larger wind speed. The sensible heat flux is proportional to a temperature difference between the ice surface and the atmosphere. The latent heat flux becomes larger with a dryer atmosphere. The sensible and latent heat flux coefficients, C_h and C_e, are on the order of 10^{-3}, based on extensive studies of air-open water interactions by the likes of Smith and Dobson (1984). The coefficients are found to increase with higher wind speed, while the values are comparable for open water and sea ice. For polar to subpolar regions, the sensible heat flux is the largest during winter (a few hundreds of Wm^{-2}). In contrast, solar radiation becomes a major component in summer. These thermodynamic calculations require input fields of wind, air temperature, dew point and cloudiness.

Snow cover tends to insulate ice in winter and reduces ice formation; hence we need precipitation data in order to accurately include its effects on ice growth. Snow reflects more short-wave radiation in spring than ice and slows down ice melting. Snow on sea ice can transition into what is known as “snow ice” due to sea water flooding and subsequent freezing, hence it can be difficult to estimate snow depth in spring. Omission of the snow cover in models yields more ice formation in winter and faster melting in spring.

Sea ice exhibits unique properties due to the fact that it is made up of solid ice crystals interspersed with pockets of liquid brine. As ice forms, salt is rejected from the solid part to the liquid part. Salinity concentrations are about 35 ‰ in sea water and 5 ‰ in sea ice. Salt is partially accumulated in the liquid fraction of the sea ice. The liquid fraction is called a brine pocket and it increases the effective heat capacity of the sea ice, as latent heat is required to maintain brine pocket expansion under warming. This extra heat capacity in brine pockets tends to maintain ice volume in spring and results in less short-wave radiation absorption by open water.

2.3. Thermodynamics in a Partial Ice Cover

Near the ice edge, it is more appropriate from a modeling perspective to consider open water interspersed with uniform sea ice. This two-level representation of ice has been used as in Hibler (1979). As long as sea ice is present, water temperature in a surface water layer is set at the in situ freezing point. This assumption implies an infinitesimal time scale of heat transfer between ice and water, although this condition could be easily relaxed by incorporating a time-derivative term with a finite time scale.

In most cases, a numerical model is used to simulate ice growth and decay. A model typically represents the region of concern using a horizontal grid system. Within each grid cell, thickness and concentration of ice are variables used to describe ice distribution in time and space. Within one cell, ice formation or decay is calculated separately in open water and ice-covered portions, i.e. heat deficit in open water is used to form thin ice, while heat removed through the ice-covered portion is converted to an increased thickness of (thick) ice. The thin ice is immediately merged into the thick ice via conservation of mass. Under a warming condition, heat flux into an open water portion of the surface layer melts the thick ice in an adjacent grid cell. Heat flux into the ice-covered portion decreases the thickness of the thick ice as long as the ice is thicker than a given minimum thickness (e.g. 0.5 m). Once the thickness is reduced to the minimum thickness, all remaining heat is used to melt the ice from the side and form open water. Ice is assumed to have no heat storage.

Heat flux is larger in open water both in winter and summer. In winter, water is warmer than sea ice and hence emits larger sensible and latent heat fluxes. In summer, larger solar radiation is absorbed into the open water portion due to its lower albedo (~0.05) than sea ice (0.5 – 0.8). Once the percentage of open water increases, more heat is absorbed, and sea ice reduces. This mechanism is called ice-albedo feedback and it is active in the summer months.

2.4. Multi-Category Ice Model

A two-level formulation of ice thickness contains only thick ice and open water categories, i.e. the predicted outputs are concentration (a ratio of the area of uniform ice to the total area) and effective thickness (thickness multiplied by concentration). This configuration is categorized into a two-category model, which was later extended to a multi-category model (Walsh et al., 1985). In a multi-category model one category is converted to the adjacent category due to both thermal effects (growth/decay) and kinematic effects (convergence). The conversion is parameterized as a function of these thermal and kinematic processes.

The simplest version is a three-category (3C) model (Ikeda et al., 2004). The new ice category, which is very thin ice with 0.1 m thickness, is added to the two-category model (Figure2). The advantage of this approach is to reproduce ice concentration more appropriately in the seasonal ice region. The thermodynamics of the thin ice are dealt with independently of the thick ice, but under identical formulations. Heat loss through the thin ice and the open water portion is converted to an increase in thin ice concentration, with its thickness kept to the specified value (0.1 m). The thin ice portion contributes to ice concentration, but has little effect on ice volume. The thin ice can prevent air-sea heat flux during the ice formation season and reflect short-wave radiation during the ice melting season. Thus, a seasonal ice cover is simulated better in the three-category model than in the two-category model.

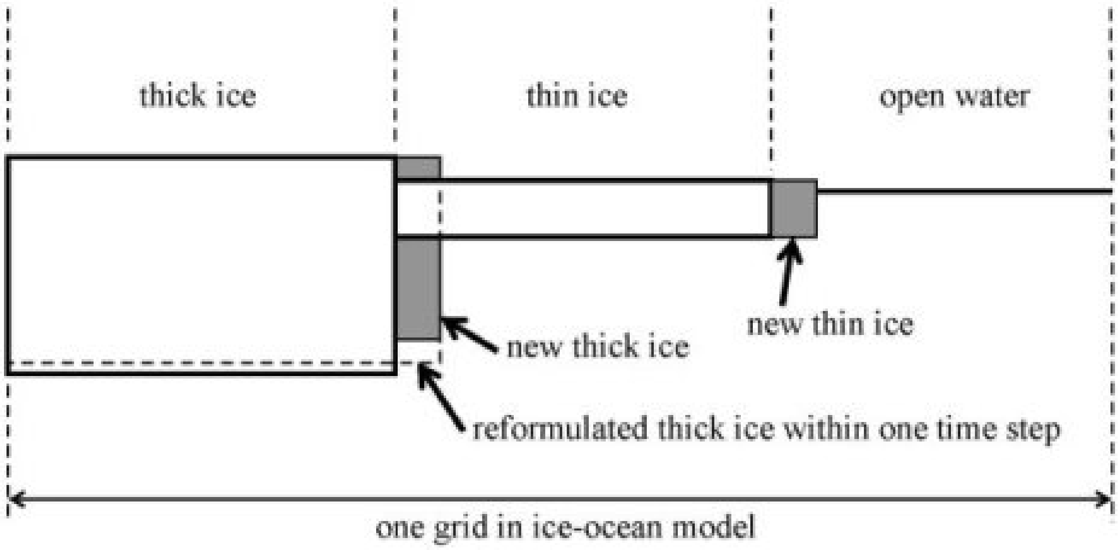

Figure 2. Three category sea ice model.

Once the thin ice is included, the thick ice increases its concentration by rafting of thin ice. This assumption is justified by the observation that thin ice often rafts over itself in the ocean. The process is represented in the model by deformation, which is a process that converts thin ice into thick ice. Ice velocities are calculated from the momentum equations including internal stresses, as described in the next section. The ice rheology is identical to the viscous-plastic formulation proposed by Hibler (1979). It is assumed that the bulk ice strength, which is a key parameter in ice deformation, can be computed from taking the areal average of both categories of ice. The bulk strength approaches the strength of either category under the condition that the other category diminishes. When advection forces the ice concentration to exceed 100%, thin ice is converted into thick ice within one time step. Once all the thin ice has been converted to thick ice, the thick ice is assumed to deform. In addition to dynamic forcing, thick ice also increases in thickness due to thermodynamics forcing as described earlier.

2.5. Dynamics of Sea Ice

Ice dynamics in models are governed by a momentum balance between air and water stresses, Coriolis force, gravity due to sea surface slope and internal ice stress. The internal ice stress is typically determined using a plastic constitutive law, which is replaced with a viscous law for very small rates of strain. Although the ice rheology used in models has not been confirmed in field (Rothrock, 1979), it has been found to adequately represent alongshore ice movement, which is not sensitive to the internal shear stress except for within the vicinity of a coastal boundary. Furthermore, it also adequately represents the case where the ice resists onshore wind forcing and becomes solid like. Sea ice is often fixed to a coastal boundary, and it is known as “fast ice” when this occurs. The case of free-drifting ice, where the internal stress of the ice is considered to be negligible, is taken to represent fundamental sea ice dynamics and is thus examined here first.

Free-drifting sea ice receives air-ice drag and ice-water drag as its main forcing:

$$\tau_a = \rho_a C_d U_a{}^2 \qquad (1.3)$$

$$\tau_w = \rho C_w (U_i - U_w)^2 \qquad (1.4)$$

Here, ρ_a is air density, ρ is water density, U_i is ice velocity and U_w is water velocity. The velocity of free-drifting ice is almost completely determined by the wind velocity and a ratio of the air-ice drag coefficient ($C_d = \sim 2\times10^{-3}$) to the ice-water drag coefficient ($C_w = \sim 7\times10^{-3}$), with a minor contribution from Coriolis force along with gravity due to sea surface slope. Water density is roughly 1000 times larger than air density, hence ice velocity is about 2% of the wind velocity (e.g., Pease et al., 1983).

A minor input from the Coriolis force nearly balances with gravity due to sea surface slope under a no-wind condition. In a horizontally uniform ocean with wind forcing, only the Coriolis force is at work (i.e. sea surface slope is uniformly zero) and it is responsible for vertical profiles of water speed and direction under sea ice, in a similar manner as those in an open water area. For the northern hemisphere the near-surface ocean layer, called the Ekman layer, has a vertically averaged flow which is directed towards the right when facing in the downwind direction. Its thickness is around 30 m, and its velocity diminishes at the bottom of the boundary layer. Thus, wind forcing balances with Coriolis force in the entire boundary layer. It is natural to expect sea ice motion to also turn to the right when facing the downwind direction. As shown in Figure3, the sea ice part receives forcing from the air-ice drag, ice-water drag and the Coriolis force. The ice velocity is directed between the wind and the water velocity under sea ice so that the air-ice drag is in balance with the ice-water drag and the Coriolis force. An interested reader is directed to physical oceanography textbooks on the Ekman spiral (e.g., LeBlond and Mysak, 1978) for further information.

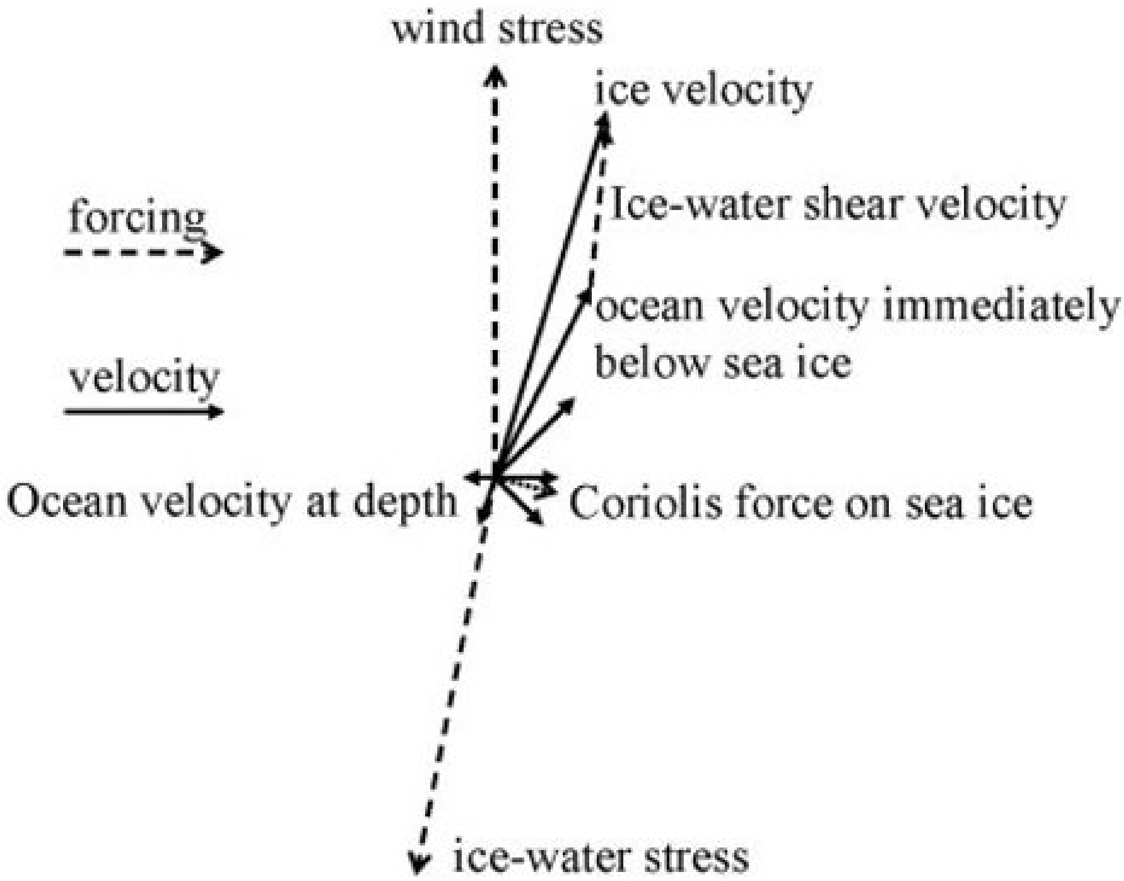

Figure 3. Ekman spiral under sea ice with wind forcing in the northern hemisphere.

The governing momentum equations contain time-derivative terms, however a practical modification is made to the equations in the models, i.e. that the time derivative terms are negligible and the forces acting on the ice are balanced at each time step. This modification is reasonable, because a balance between air-ice stress and ice-water stress is established within about 1000 s (Parkinson and Washington, 1979), and it also remains valid when the internal ice stress is included.

2.6. Rheology of Sea Ice

Internal ice stress only plays an important role in sea ice dynamics near a coastal boundary at time scales longer than a few days. However, it becomes more influential in any location at shorter time scales when ice forcing generated by tides and wind waves is considered. The principle of internal stress in the ice is typically expressed using a plastic constitutive law, in which stresses are related to strain rates within a horizontal two-dimensional domain. An interested reader is directed to read Coon et al. (1974) and Hibler (1979) for further information.

A more recent approach is to replace the plastic constitutive law with an elastic one, as proposed by Hunke and Dukowics (1997). The elastic constitutive law requires an extremely small time step for stable computation. However, a computer composed of parallel processors can handle this type of computation more easily than the heavy iterative computation required in the case of the plastic constitutive law. There is essentially no difference in results between these two types of models.

Fluid rheology in numerical models is usually expressed by the Newtonian viscous law, in which shear stress is proportional to strain rate. One of the most significant differences between the plastic law and the viscous law is realized as an asymmetric response to forces inducing ice compression from those forces inducing ice expansion. Sea ice which is represented using the plastic law resists compression up to a critical pressure, but freely moves under expansion. Ice strength is on the order of 10^4 Nm^{-2}. Sea ice with a 1 m thickness can resist the pressure accumulated over an ice cover with a transverse length of 100 km under shoreward wind forcing of 0.1 Nm^{-2}. Once the applied stress reaches the ice strength, ice is crushed into deformed ice. Thinner ice is more easily deformed, even in the middle of the Arctic Basin. Most observed ice ridges are thought to be created from thin ice growing in a polynya (an ice free region within a mostly ice-covered area) under spatially variable wind forcing.

Another difference between the plastic law and the viscous law is also clear when shear stress is applied to sea ice. As derived by Ikeda (1988), sea ice becomes motionless at a coastal boundary below a critical condition, but slips at the boundary where the shear stress is acting above the critical condition, independent of alongshore velocity. The critical value is determined by a cross-shore pressure gradient, which increases under a shoreward wind stress. This critical value increases even when there is only an alongshore wind forcing, as ice velocity tends to turn shoreward due to the Coriolis force, e.g. towards a coastal boundary on the right when facing downwind in the northern hemisphere, and towards the left in the southern hemisphere.

-
-
-

Bibliography

Aoki, S., N.L. Bindoff, and J.A. Church, (2005) Interdecadal water mass changes in the Southern Ocean between 30°E and 160°E. *Geophys. Res. Lett.*, **32**, L07607, doi:10.1029/2004GL022220.[Information of water mass change]

Bryden, H.L., H.R. Longworth, and S.A. Cunningham, (2005) Slowing of the Atlantic meridional overturning circulation at 25°N. *Nature*, **438**, doi:10.1038/nature04385.[Information of merridional overturning]

Comiso, J.C., and C.L. Parkinson, (2004) Satellite-observed changes in the Arctic. *Physics Today*, **August 2004**, 38-44.[Information of sea ice change]

Coon, M.D., S.A. Maykut, R.S. Pritchard, D.A. Rothrock and A.S. Thorndike, (1974) Modeling the pack ice as an elastic-plastic material. *AIDJEX Bull.*, **24**, 1-105.[Pioneer work on ice modeling]

Deser, C., J.E. Walsh and M.S. Timlin, (2000) Arctic sea ice variability in the context of recent atmospheric circulation trends. *J. Climate*, **13**, 617-633.[Information of climate change]

Dickson, R.R., J. Meincke, S.A. Malmberg, and A.J. Lee, (1988) The Great Salinity Anomaly in the Northern North Atlantic 1968-1982. *Prog. Oceanogr.*, **20**, 103-151.[Most comprehensive work on salinity anomaly]

Fichefet, Th., and Ph. Gasper, (1988) A model study of upper ocean-sea ice interactions. *J. Phys. Oceanogr.,* 18, 181-195.[Pioneer work on water mass formation in the Greenland Sea]

Gong, G., D. Entekhabi, and J. Cohen, (2003) Modeled Northern Hemisphere winter climate response to realistic Siberian snow anomalies. *J. Climate*, **16**, 3917–3931.[Information of feedback from snow to atmosphere]

Guest, P.S., and K.L. Davidson, (1987) The effect of observed ice conditions on the drag coefficient in the summer East Greenland Sea Marginal Ice Zone. *J. Geophys. Res.,* **92**, 6943-6954.[Information of air-ice drag]

Haine, T.W.N., A.J. Watson, M.L. Liddicoat and R.R. Dickson, (1998) The flow of Antarctic bottom water to the southwest Indian Ocean estimated using CFCs. *J. Geophys. Res.,* **103**, 27637-27653.[Information of Antarctic Bottom Water]

Hakkinen, S., (1987) A coupled dynamic-thermodynamic model of an ice-ocean system in the marginal ice zone. *J. Geophys. Res.,* **92**, 9469-9478.[Pioneer work on marginal ice zone modeling]

Hibler, W.D., (1979) A dynamic thermodynamic sea ice model. *J. Phys. Oceanogr.,* **9**, 815-846.[Pioneer work on sea ice modeling]

Hibler, W.D., and K. Bryan, (1987) A diagnostic ice-ocean model. *J. Phys. Oceanogr.,* **17**, 987-1015.[Pioneer work on ice-ocean interactions]

Hilmer, M., and P. Lemke, (2000) On the decrease of Arctic sea ice volume. *Geophys. Res. Lett.*, **27**, 3751-3754.[Information of Arctic sea ice reduction]

Hiraike, Y., and M. Ikeda, (2009) Descending surface water at the Antarctic marginal ice zone and its contribution to the Intermediate Water using an ice-ocean model. *J. Oceanogr*. (submitted).[Information of wind effects on sea ice]

Hogg, N., and W. Zenk, (1997) Long-period changes in the bottom water flowing through Vema Channel. *J. Geophys. Res.,* **102**, 15,639-15,646.[Information of bottom water formation]

Hunke, E.C., and J.K. Dukowics, (1997) An elastic-viscous-plastic model for sea ice dynamics. *J. Phys. Oceanogr*., **27**, 1849-1867.[Information of sea ice modeling]

Hurrell, J.W., (1995) Decadal trends in the North-Atlantic Oscillation - Regional temperatures and precipitation. *Science*, **269**, 676-679.[Information of North Atlantic Oscillation]

Ikeda, M., (1986) A mixed layer beneath melting sea ice in the marginal ice zone using a one-dimensional turbulent closure model. *J. Geophys. Res.,* **91**, 5054-5060.[Pioneer work on sea ice effects on mixed layer]

Ikeda, M., (1988) A model study of wind- and buoyancy-driven coastal circulation. *J. Geophys. Res.*, **93**, 5078-5092.[Pioneer work on ice-edge upwelling]

Ikeda, M., (1991a) Numerical modeling of ocean circulation and ice cover over the continental shelf. *J. Phys. Oceanogr.,* **21**, 97-117.[Pioneer work on ice-ocean circulation on the shelf]

Ikeda, M., (1991b) Wind-induced mesoscale features in a coupled ice-ocean system. *J. Geophys. Res.*, **96**, 4623-4629.[Pioneer work on ice-ocean interactions with mesoscale features]

Ikeda, M., R. Colony, H. Yamaguchi and T. Ikeda, (2005) Decadal variability in the Arctic Ocean shown in hydrochemical data. *Geophys. Res. Lett.*, **32**, L21605, doi:10.1029/2005GL023908.[Pioneer work on geochemical components showing Arctic Ocean variability]

Ikeda, M., J. Wang and A. Makshtas, (2003) Importance of clouds to the decaying trend in the Arctic ice cover. *J. Meteorol. Soc. Japan*, **81**, 179-189.[Information of cloud variability in the Arctic]

Ikeda, M., J. Wang and J-P. Zhao, (2001) Hypersensitive Arctic-subarctic climate system. *Geophys. Res. Lett.*, **28**, 1275-1278.[Pioneer work on decadal ice-ocean variability in the Arctic]

Ikeda, M., T. Yao and Q. Yao, (1996) Seasonal evolution of sea ice cover and shelf water off Labrador simulated in a coupled ice-ocean model. *J. Geophys. Res.,* **101**, 16465-16489.[Information of Labrador Sea ice modeling]

Ikeda, J., H. Shinkai and T. Watanabe, (2004) Parametrization of the thin ice in a coupled ice-ocean model: application to the seasonal ice cover in the Sea of Okhotsk. *Atmos. Ocean,* **42**, 1-12.[Information of Okhotsk Sea ice modeling]

IPCC, (2007) Intergovernmental Panel on Climate Change, Working Group I Report: The physical science basis of climate change. eds. S. Solomon and D. Qin.[Comprehensive review on climate change in the polar oceans]

LeBlond, P.H., and L.A. Mysak, (1978) Waves in the ocean. Elsevier Scientific Publishing Company, Amsterdam.[Textbook of ocean dynamics]

Lemke, P., (1987) A coupled one-dimensional sea ice-ocean model. *J. Geophys. Res.*, **92**, 13164-13172.[Pioneer work on ice-ocean modeling]

Maykut, G.A., and N. Untersteiner, (1971) Some results from a time-dependent thermodynamic model of sea ice. *J. Geophys. Res.*, **76**, 1550-1575.[Pioneer work on sea ice modeling]

McCartney, M.S., (1977) Subantarctic Model Water. *Deep-Sea Res*., **24, suppl.**, 103-119.[Pioneer work on water formation]

McPhee, M.G., (1983) Turbulent heat and momentum transfer in the oceanic boundary layer under melting pack ice. *J. Geophys. Res.*, **88**, 2827-2835.[Pioneer work on ice-ocean observations]

Mellor, G.L., M.G. McPhee and M. Steele, (1986) Ice-seawater turbulent boundary layer interaction with melting or freezing. *J. Phys. Oceanogr.,* **16**, 1829-1846.[Pioneer work on ice-ocean interaction]

Nakanowatari, T., K.I. Ohshima and M. Wakatsuchi, (2007) Warming and oxygen decrease of intermediate water in the northwestern North Pacific, originating from the Sea of Okhotsk, 1955-2004. *Geophys. Res. Lett*., **34**, L04602, doi:10.1029/2006GL028243.[Information of North Pacific Intermediate Water]

Noone, D., and I. Simmonds, (2002) Annular variations in moisture transport mechanisms and the abundanc of delta O-18 in Antarctic snow. *J. Geophys. Res*., **107**, 4742.[Information of moisture flux around Antarctic]

Ohshima, K.I., and S. Nihashi, (2005) A simplified ice-ocean model for the Antarctic ice melt season. *J. Phys. Oceanogr.,* **35**, 188-201.[Information of ice melting in the Antarctic Ocean]

Ohshima, K.I., S.C. Riser and M. Wakatsuchi, (2005) Mixed layer evolution in the Sea of Okhotsk observed with profiling floats and its relation to sea ice formation. *Geophys. Res. Lett.*, **32**, L06607, doi:10.1029/2004GL021823, 2005[Information of ice-ocean interactions in the Okhotsk Sea]

Parkinson, C.L., and W.M. Washington, (1979) A large-scale numerical model of sea ice. *J. Geophys. Res.*, **84**, 311-337.[Pioneer work on climate-scale sea ice variability]

Pease, C.H., S.A. Salo and J.E. Overland, (1983) Drag measurements for first-year sea ice over a shallow sea. *J. Geophys. Res.*, **88**, 2853-2862.[Pioneer work on ice-edge process]

Pease, C.H., (1987) The size of wind-driven coastal polynyas. *J. Geophys. Res.,* **92**, 7049-7059.[Pioneer work on polynya]

Polyakov, I.V., A.Y. Proshutinsky and M.A. Johnson, (1999) Seasonal cycles in two regimes of Arctic climate. *J. Geophys. Res.*, **104**, 25761-25788.[Pioneer work on sea ice with Arctic Oscillation]

Proshutinsky, A., and AOMIP Group, (2005) Arctic Ocean study: synthesis of model results and observations. *EOS, Transaction*, American Geophysical Union, **86**, 368-371.[Review of ice-ocean modeling]

Roed, L.P., and J.J. O'Brien, (1983) A coupled ice-ocean model of the marginal ice zone. *J. Geophys. Res.*, **88**, 2863-2872.[Pioneer work on ice-edge upwelling]

Rothrock, D.A., (1979) Modeling sea-ice features and processes. *J. Glaciol.*, **24**, 359-375.[Pioneer work on sea ice modeling]

Rothrock, D.A., Y. Yu and G.A. Maykut, (1999) Thinning of the Arctic sea-ice cover. *Geophys. Res. Lett.*,

26, 3469-3472.[Pioneer work on sea ice decrease in the Arctic]

Saenko, O.A., and A.J. Weaver, (2001) Importance of wind-driven sea ice motion for the formation of Antarcrtic Intermediate Water. *Geophys. Res. Lett.,* **28**, 4147-4150.[Information of water formation in the southern ocean]

Santoso, A., and M.H. England, (2004) Antarctic Intermediate Water circulation and variability in a coupled climate model. *J. Phys. Oceanogr.*, **34**, 2160-2179.[Information of ice-ocean modeling in the southern ocean]

Sasai, Y., A. Ishida, Y. Yamanaka and H. Sasaki, (2004) Chlorofluorocarbons in a global ocean eddy-resolving OGCM: Pathway and formation of Antarctic Bottom Water. *Geophys. Res. Lett.*, **31**, L12305, doi:10.1029/2004GL019895. [Information of bottom water formation in the Antarctic Ocean]

Schauer, U., and E. Fahrbach, (2004) Arctic warming through the Fram Strait: Oceanic heat transport from 3 years of measurements. *J. Geophys. Res*., **109**, C0606, doi:10.1029/2003JC001823.[Information of warming in the Arctic]

Semtner, A.J., (1987) A numerical study of sea ice and ocean circulation in the Arctic. *J. Phys. Oceanogr.*, **17**, 1077-1099.[Pioneer work on Arctic Ocean]

Shoutilin, S.V., A.P. Makshtas, M. Ikeda, A.V. Marchenko and R.V. Bekryaev, (2005) Dynamic-thermodynamic sea ice model: ridging and its application to climate study and navigation. *J. Climate,* **18**, 3840-3855.[Information of ridged sea ice modeling]

Smith, S.D. and F.W. Dobson, (1984) The heat budget at ocean weather station Bravo. *Atmos. Ocean*, **22**, 1-22.[Pioneer work on air-sea heat flux measurement]

Smith, S.D., R.D. Muench and C.H. Pease, (1990) Polynyas and leads: an overview of physical processes and environment. *J. Geophys. Res.,* **95**, 9461-9479.[Pioneer work on polynya modeling]

Steele, M., and T. Boyd, (1998) Retreat of the cold halocline layer in the Arctic Ocean. *J. Geophys. Res.*, **103**, 10419-10435. [Pioneer work on climate change in the Arctic]

Stroeve J., M.M. Holland, W. Meier, T. Scambos and M. Serreze, (2007) Arctic sea ice decline: Faster than forecast, *Geophys. Res. Lett.,* **34**, L09501, doi:10.1029/2007GL029703.[Topical work on sea ice reduction in the Arctic]

Talley, L.D., (1991) An Okhotsk Sea water anomaly: Implications for ventilation in the North Pacific. *Deep-Sea Res., Part A*, **38, suppl**., S171-S190.[Pioneer work on change in the North Pacific Interrnediate Water]

Tamura T., K. I. Ohshima and S. Nihashi, (2008) Mapping of sea ice production for Antarctic coastal polynyas. *Geophys. Res. Lett.*, **35**, L07606, doi:10.1029/2007GL032903.[Information of ice formation in polynya]

Tang, C.L., and M. Ikeda, (1989) Ice edge upwelling off the Newfoundland coast during LIMEX. *Atmos. Ocean*, **27**, 658-681.[Topical work on ice-edge process]

Thompson, D.W.J., and J.M. Wallace, (1998) The Arctic oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, **25**, 1297-1300.[Pioneer work on Arctic Oscillation]

Thompson, D.W.J., and J.M. Wallace, (2000) Annular modes in the extratropical circulation. Part I: Month-to-month variability. *J. Climate*, **13**, 1000-1016.[Comprehensive work on Arctic Oscillation]

Tokinaga H., and Y. Tanimoto, (2004) Seasonal transition of SST anomalies in the tropical Indian Ocean during El Nino and Indian Ocean Dipole years: *J. Meteor. Soc. of Japan*, **82**, 1007-1018.[Pioneer work on air-sea interactions at climate time scale]

Vinnikov, K.Y., A. Robock, R.J. Stouffer, J.E. Walsh, C.L. Parkinson, D.J. Cavalieri, J.F.B. Mitchell, D. Garett and V.F. Zakharov, (1999) Detection and attribution of global warming using northern hemisphere sea ice. *Science*, **286**, 1934-1937.[Pioneer work on Arctic sea ice modeling under global warming]

Walsh, J.E., W.D. Hibler and B. Ross, (1985) Numerical simulation of Northern Hemisphere sea ice variability, 1951-1980. *J. Geophys. Res.*, **90**, 4847-4865.[Pioneer work on climate change in the Arctic Ocean]

Wang, X., and J.R. Key, (2003) Recent trends in Arctic surface, cloud and radiation properties from space. *Science*, **299**, 1725-1728.[Information of cloud variability in the Arctic]

Willmott, A.J., and L.A. Mysak, (1989) A simple steady-state coupled ice-ocean model, with application to the Greenland-Norwegian Sea. *J. Phys. Oceanogr.*, **18**, 501-518.[Pioneer work on simple ice-ocean modeling]

Yao, T., and M. Ikeda, (1990) A model of sea ice and the upper ocean mixed layer off Labrador. J. Geophys. Res., 95, 11603-11612.[Information of ice-ocean modeling on the Labrador Shelf]

Zhang, J., D. Rothrock and M. Steele, (2000) Recent changes in Arctic sea ice: the interplay between ice dynamics and thermodynamics. *J. Climate*, **13**, 3099-3114.[Information of Arctic sea ice change]

Biographical Sketch

Motoyoshi Ikeda born in Tokyo, Japan, 1946. Ph.D in Aeronautics, University of Tokyo, 1974. Research activities have been made in climate variability, ocean data assimilation and socio-climate modeling.

He has taken the professional positions as follows:

Research Student, Institute of Space and Aeronautical Sciences, April 1974 - Aug. 1979

Research Associate of the U.S. National Research Council at NOAA/PMEL, Sep. 1979 - Aug. 1981

Research Associate of the University of British Columbia, Sep. 1981 - Nov. 1983

Research Scientist, Bedford Institute of Oceanography, Dec. 1983 - Mar. 1994

Professor, Hokkaido University, Apr. 1994 - present

Program Director for the IARC, Frontier Research System for Global Change, Oct. 1997 to Mar. 2002

Dean, Graduate School of Environmental Earth Science, Apr. 2002 - Mar. 2007

His publication includes:

Ikeda, M., and F. Dobson (Eds.), Oceanographic Applications of Remote Sensing, CRC Press, 1995

Professor, Dr. Ikeda, American Geophysical Union, Oceanographic Society of Japan, Canadian Meteorological and Oceanographic Society