SEA - ICE INTERACTIONS

J. E. Weber
Department of Geophysics, University of Oslo, Norway

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Summary

This article presents an overview of the formation and the distribution of sea ice. The physical processes related to the freezing of seawater are discussed. The article identifies several cases where the ice and the ocean interact. This encompasses phenomena such as salt-driven convection beneath growing sea ice. In cases where the density stratification prior to ice formation is weak, salt-driven convection may mix the ocean vertically to considerable depths. Also ocean currents generated by wind-driven ice are discussed. We here include the effect of under-ice topography in the discussion of the angle between the wind direction and the ice drift direction. Furthermore, dynamical processes that occur where the ice meets the open ocean are discussed. Here the action of ocean surface waves is important. Such waves tend to break up the ice when they enter ice-covered areas. As they penetrate into the ice, their amplitudes diminish. This damping is most pronounced for surface waves with small periods, or equivalently, with short wavelengths. Another important process that may occur where the ice meets the ocean is upwelling of water from below. This phenomenon is due to the differences between the wind stress over the open ocean and over the ice. The upwelled water is usually rich in nutrients and may enhance the production of plant

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algae near the surface. We also discuss how the calving of the large ice shelves in Antarctica may influence the production of dense, Antarctic bottom water. This phenomenon is important for the renewal of the deep water in the world's oceans.

1. Introduction

Ice in the sea originates mainly from the freezing of seawater. Such ice is referred to as sea ice. In polar areas the temperatures may be so low that glaciers or continental ice sheets do not lose mass due to melting at the front. In such cases the ablation process is achieved by calving into the sea to form icebergs. Although the presence of icebergs is important as far as the navigation of ships is concerned, they are isolated features that contribute little to the physics of the ocean - atmosphere system. One exception here is icebergs that are formed by the calving of the Antarctic ice shelves. The Antarctic ice shelves interact with the ocean through a slow circulation and cooling of seawater beneath the floating ice, which produces dense Antarctic shelf bottom water. This process is fairly stationary in time. In addition, the calving of the Antarctic ice shelf exposes very cold ice to the much warmer seawater in an abrupt fashion. Due to immediate freezing, downward salt-driven currents are formed adjacent to the vertical ice walls. This brings heavy, saline water to the bottom of the Antarctic shelf. This freezing process is transient in time, and will finally halt. In the end, the heat from the warmer ocean will cause the ice to melt. The melting is very slow at the remaining ice shelf front, amounting to about 0.6 m yr^{-1}. The icebergs, however, drift under the influence of winds and ocean currents into warmer water and melt much more quickly.

Sea ice covers large areas of the world's oceans (approximately 10 percent of the Northern hemisphere and 13 per cent of the Southern hemisphere). Besides being a severe hindrance to navigation, it has a deep impact on the momentum transfer from the atmosphere to the ocean (**see *Air - Sea Interactions. Surface waves***). Furthermore, the ice cover changes the albedo at the earth's surface and it acts as insulation against heat transfer from the ocean to the atmosphere. Melting of sea ice usually increases the static stability of the upper ocean due to the added fresh water supply, while the freezing process increases the upper-layer salinity and may lead to instability in the form of salt-driven convection (**see *Mixing and Fronts***). Also, in the regions where sea ice meets the open ocean, i.e. the so-called marginal ice zone (MIZ), incoming surface waves become rapidly damped when they propagate into the ice-covered ocean (**see *Waves***). Furthermore, the effect of the wind stress on the ocean is not the same over the open sea, as it is in ice-covered areas. In particular, when the ice is stationary, the wind stress becomes discontinuous at the ice edge. For winds along the edge, the horizontal Ekman transport perpendicular to the ice edge cannot be matched by a corresponding transport under the ice, and upwelling or downwelling may occur.

2. Sea Ice Formation

The freezing point T_f of seawater is a monotonically decreasing function of the salinity. At atmospheric pressure, fresh water freezes at 0°C. The depression of the freezing point can be approximated as $T_f = -0.054S$°C psu^{-1}, where S is the salinity of seawater in psu (practical salinity unit). The density of seawater near the surface (at atmospheric pressure) is a function of temperature and salinity (**see *Sea Water and its Physical***

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Properties: Heat and Fresh Water Balance of the Oceans). When seawater of salinity less than 24.7 psu is cooled, the density attains a maximum before the freezing point is reached. For fresh water this maximum occurs when the temperature is about 4°C. When the salinity is larger than 24.7 psu, the density increased monotonically with decreasing temperature till the freezing point is reached. This anomalous behaviour will affect the speed at which sea ice is formed at different locations. In coastal areas influenced by rivers, we may find brackish water of salinity less than 24.7 psu. For example, in the Bothnian Bay in the Baltic Sea we find salinities as low as 5 psu. In such areas cooling below the temperature of maximum density will produce a thin, dynamically stable layer at the surface, with water of less density above water of higher density. This thin layer loses heat quite rapidly, and freezing soon commences.

In open ocean areas the salinity is larger than 24.7 psu. Here we usually find salinities close to the standard value of 35 psu. When such seawater is cooled, the water at the surface will always have a larger density than that below the surface. This is a dynamically unstable situation. As a result, convective overturning will occur that mixes the water in the vertical. How far down the penetrating front of this mixed layer will reach, depends on time, on the surface temperature and on the vertical density distribution in the water column prior to cooling. But in general, sea ice will not start to form until a water column of considerable depth is cooled down to the freezing point. In high latitudes this column depth could well be 30 to 50 m. This will delay the onset of ice formation as compared to the case of brackish water.

When freezing occurs, pure water transforms into ice during the phase transition, and the salt is rejected back into the sea. At 0°C the density of pure water is 999.9 kg m^{-3} and the density of pure ice is 916.8 kg m^{-3}. However, due to the crystal structure at the advancing ice front, high salinity water may be trapped among the ice crystals, forming so-called brine pockets within the ice. This will increase the density of sea ice. Values of 924 kg m^{-3} have been recorded in the Arctic Sea. It should also be mentioned that bubbles of air might be present in the ice as a result of the formation process. This will reduce the value of the sea ice density.

The tendency for mechanical trapping of salt in the ice increases with increasing ice growth rate. Since the ice grows most rapidly in the beginning of the freezing process, newly formed sea ice may have salinities as high as 20 psu, while the average salinity has been reduced to about 4 psu after a year or so. However, the salt will not stay permanently in the ice. Normally, there will be vertical temperature gradients in the sea ice, and hence across the brine pockets. For phase equilibrium to exist, this must imply that the ice at the warm end of the brine pocket will melt, diluting the brine. At the cold end freezing occurs, increasing the brine concentration and possibly causing some of the salts to crystallise out. As a result, there is a migration of the brine pocket towards the warm side of the ice. Also pressure build-up in the brine pockets, leading to a separation between liquid and vapour, may cause the ice to fail along the basal planes of crystals, which may allow brine to escape. This is called brine expulsion. As the ice thickens and rises above sea level, gravity acts to drive the brine out of the ice. The salt actually leaves the ice through interconnecting drainage channels of the order 5 mm in diameter from which it trickles back into the sea. This discharge may lead to brine streamers and the formation of hollow ice tubes called ice stalactites. Also gravity drainage of salt may

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occur in the spring and the summer due to the pressure created by surface melt water. This process is called flushing. As a result of all these processes, sea ice that is several years old contains very little salt.

In very calm seas, the ice crystals grow laterally to form a continuous skin. However, in nature, the wind and wave-induced turbulence during the initial phase of ice growth causes the ice crystals to become stirred in the upper layer to form so-called frazil ice. In some cases the frazil ice crystals are herded by the action of the wind to form a soupy agglomeration known as grease ice. The grease ice may form a high viscous-like, soupy surface layer of thickness of up to one metre. Due to the pumping action of ocean waves, frazil ice lumps together to form pancake ice, often with grease ice in between the agglomerated pancakes. Finally, the pancakes and the grease ice freeze together to form a continuous sheet of ice. Young ice, resulting from one freezing season, is called first-year ice. Ice, which survives the first summer and continues to grow the next winter, is called multiyear ice.

3. Distribution of Sea Ice

Near the shore, sea ice forms as fast ice. This ice is anchored to islands or coastal boundaries. It forms rather quickly under freezing atmospheric conditions, since the near-shore water usually is shallow and therefore cools down more easily. The fast ice stretches out towards the open sea where it becomes more directly exposed to wind, incoming ocean waves and ocean currents. Here it breaks up to form pack ice, which consists of ice floes. These ice floes may vary considerably in size and in their degree of packing. Closest to the fast ice we find the largest ice floes. The floes tend to become smaller as one approaches the open sea. This area is often termed the marginal ice zone (MIZ). In the outer part of the MIZ, the ice is more or less broken down by the action of ocean waves. Here the individual ice floes are very small and often closely packed so that they form a soupy ice-agglomeration. The pack ice is very much influenced by wind and ocean currents, being the main agencies by which the ice moves. In regions with converging ice motion, compression and shear forces within the ice lead to the formation of individual pressure ridges or rubble fields. When an ice sheet deforms under compression, ice blocks, created by bending and buckling, are piled up above and below the surface. The part of the pressure ridge formed above the surface is called the sail, while the under-water part is called the keel. One usually finds that the keel depth is about 3 to 4 times larger than the sail height. In regions with diverging ice motion, we may find areas with open waters even in winter times. Such open waters are called leads. In winter times the leads quickly re-freeze. We may also find larger areas with open water in the pack ice, called polynyas. The polynyas may have a more permanent character and are often generated by upwelling of warmer ocean water that melts the ice from beneath.

3.1. Ice in the Arctic and Adjacent Seas

In the Northern hemisphere the Arctic Sea, the Canadian Archipelago, Baffin Bay/Davis Strait, Hudson Bay, Labrador Sea and the Barents Sea are covered, or partly covered, by ice during the larger part of the year. It is worth mentioning that the western coast of north Norway, which is situated nearly at the same latitude as Baffin Bay, is completely

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free of ice all winter, except for the occasional freezing over of the inner parts of fjords. This is due to the influence of the warm, saline Norwegian Current that carries water of Gulf Stream origin along the western coast of Norway. In the open North Pacific ice does not occur. However, in winter, ice is formed in the Bering Sea, the Sea of Okhotsk and the north of the Sea of Japan. In Baffin Bay/Davis Strait there is mostly first-year ice of 1.5 to 2 m thickness. We also find some older ice up to 3 m thick entering at the north from the Arctic through Smith Sound. In this region it is more common to find ice cover than open water. In the Barents Sea we also have mainly first-year ice, but occasionally we find multiyear ice that has arrived from the Arctic Sea and survives the summer. The ice extent in the Barents Sea is largest at the end of March, where ice reaches as far south as Bear Island, and as far east as Novaya Zemlaya. In September the ice extent is at its minimum. Then the ice edge is located north of Spitsbergen.

The surface circulation in the Arctic Sea is dominated by an anti-cyclonic gyre in the Beaufort Sea, and a transpolar drift from Siberia towards Fram Strait, located between Spitsbergen and Greenland (**see *Physical Oceanography Overview***). In the central part of the Arctic Sea we find the polar ice cap. It covers about 70 percent of the Arctic Sea. This ice is always several years old, i.e. multiyear ice. The average thickness of the polar ice cap is 3 to 4 m in the winter. Some melting occurs in the summer, giving an average thickness of 2 to 3 m. In the Arctic Sea the polar ice cap is surrounded by the pack ice, covering about 25 per cent of the ocean area. The largest pack ice extent is found in May and the least in September. The shore fast ice in the Arctic Sea develops to a thickness of 1 to 2 m in the winter and melts completely in the summer together with some of the pack ice. The cap ice and the pack ice do not stay permanently in the Arctic Sea. The ice has an estimated residence time of 5 to 7 years, and moves with a mean velocity of about 2 km day^{-1}, following the general clock-wise, or anti-cyclonic, circulation pattern of the upper Arctic Sea. Most of the ice leaves the area through Fram Strait. Here the ice is carried away in the East Greenland Current. The drift speed is estimated about 15 km day^{-1}. On an annual basis about 10 percent of the ice cover in the basin leaves through the Fram Strait. This amounts to about 3×10^3 km^3 yr^{-1}, which is nearly the same flux as the total freshwater input to the Arctic Sea from rivers. Obviously this transport of ice, from the Arctic Sea to lower latitudes where it finally melts, cools the ocean locally by the loss of heat needed to raise the temperature of the ice to its melting point, and by the further supply of latent heat needed to melt it. Production of pack ice in polar areas, and the subsequent melting of this ice at lower latitudes, is an important process for the earth's climate. It works to reduce the temperature difference between the equator and the poles.

3.2. Sea Ice in Antarctica

The Southern Ocean is not surrounded by land. The dominant feature here is the Antarctic Circumpolar Current, which is a deep and strong current that flows eastwards. It is also referred to as the West Wind Drift, since its direction follows the prevailing westerly winds of the area. Closer to the Antarctic continent, we find easterly winds, which are caused by the prevailing high-pressure system over the cold Antarctic continent. This wind system gives rise to the westward flowing Antarctic Coastal Current, which tends to follow the continental margin. It is also called the East Wind Drift. The dominant feature of the Antarctic Continent is the Antarctic Ice Sheet that

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covers most of it and extends out into the sea as floating ice shelves. The ice shelves are particularly prominent in the Weddell Sea and the Ross Sea.
The fact that the Arctic Sea is a closed basin surrounded by land effectively restricts the southward expansion of the ice during the winter. The only major outflow occurs through the relatively narrow Fram Strait. The ice that survives the summer may be trapped in the polar basin for a number of years. In contrast, the Antarctic sea ice is formed in an ocean with no northward limitations. Also the winds are generally stronger here, which means that the surface wave climate becomes more severe than in the Arctic. Hence, more vigorous surface wave action tends to break up the ice at the MIZ in Antarctica. Finally, the ice melts in contact with warmer surface water in the Antarctic Circumpolar Current.

The surface water in the Antarctic zone is always close to the freezing point, i.e. –1.9°C. The average extent of sea ice in summer is typically 4×10^6 km^2 and in winter 20×10^6 km^2. In Antarctica the marginal ice zones (MIZ) are considerably larger than in Arctic. More open water generally exists in the Antarctic pack ice than in the Arctic because of the much lager MIZ and the stronger wind fields that produce divergence in the ice. Furthermore, the Antarctic sea ice is thinner than the ice in the polar basin. This is partly due to a larger heat flux from the underlying water in the Southern Ocean that tends to melt the ice, and also the fact that the residence time for the ice here is smaller. As a rule, Antarctic sea ice hardly survives more than two years before it is advected to the ice edge, where it melts. Another interesting contrast is that while the Arctic sea ice extent nearly doubles in winter, the Antarctic ice cover experiences a fivefold increase in size. The larger vertical heat flux from the ocean in Antarctica is probably also responsible for the Weddell polynya, which is a large area of open water occasionally observed in the eastern Weddell Sea.

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worth reading].

Biographical Sketch

Dr. Jan Erik Weber is a full Professor of Oceanography at the University of Oslo, Norway, with a background in applied mathematics and geophysical fluid dynamics. His research interests cover areas such as air- sea - ice interaction, buoyancy-driven convection, sea ice formation, topographic wave generation and propagation, turbulent mixing, wind- and wave-induced ocean currents, and the dynamics of very viscous fluids. His interests in polar oceanography have been stimulated by lecturing at the University Courses on Svalbard (Spitsbergen), Norway. Dr. Weber is a member of the Norwegian Academy of Science and Letters, and a Life Member of Clare Hall, University of Cambridge, UK.

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