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Originally published as:

Maqueda, M. A. M., Willmott, A. J., Biggs, N. R. T. (2004): Polynya dynamics: A review of observations and modeling. - *Reviews of Geophysics*, 42, RG1004 / 2004

DOI: [10.1029/2002RG000116](https://doi.org/10.1029/2002RG000116)

POLYNYA DYNAMICS: A REVIEW OF OBSERVATIONS AND MODELING

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Received 24 July 2002; revised 26 June 2003; accepted 10 September 2003; published 18 March 2004.

[1] Polynyas are nonlinear-shaped openings within the ice cover, ranging in size from 10 to 10^5 km². Polynyas play an important climatic role. First, winter polynyas tend to warm the atmosphere, thus affecting atmospheric mesoscale motions. Second, ocean surface cooling and brine rejection during sea ice growth in polynyas lead to vertical mixing and convection, contributing to the transformation of intermediate and deep waters in the global ocean and the maintenance of the oceanic overturning circulation. Since 1990, there has been an upsurge in polynya observations and theoretical models for polynya formation and their impact on

the biogeochemistry of the polar seas. This article reviews polynya research carried out in the last 2 decades, focusing on presenting a state-of-the-art picture of the physical interactions between polynyas and the atmosphere-sea ice-ocean system. Observational and modeling studies, the surface heat budget, and water mass transformation within these features are addressed. *INDEX TERMS*: 4540 Oceanography: Physical: Ice mechanics and air/sea/ice exchange processes; 4207 Oceanography: General: Arctic and Antarctic oceanography; 3349 Meteorology and Atmospheric Dynamics: Polar meteorology; *KEYWORDS*: polynya.

Citation: Morales Maqueda, M. A., A. J. Willmott, and N. R. T. Biggs (2004), Polynya dynamics: A review of observations and modeling, *Rev. Geophys.*, 42, RG1004, doi:10.1029/2002RG000116.

1. INTRODUCTION

[2] The *World Meteorological Organization* [1970] defines a polynya as “any non-linear shaped opening enclosed in (sea) ice... Sometimes the polynya is limited on one side by the coast and is called a shore (or coastal) polynya (e.g., St. Lawrence Island Polynya) or by landfast ice and is called a flaw polynya (e.g., Laptev Sea Polynya). If it recurs in the same position every year, it is called a recurring polynya.” Polynya sizes range from 100 or 1000 m to 100 km [Smith *et al.*, 1990] or, in terms of their areas, from 10 to 10^5 km² [Barber *et al.*, 2001a].

[3] According to their mechanism of formation and maintenance, polynyas have been traditionally divided into two classes: “sensible heat” polynyas and “latent heat” polynyas. Sensible heat polynyas are thermally driven. They appear as a result of oceanic sensible heat entering the area of polynya formation in amounts large enough to melt any preexisting ice and prevent the growth of new ice. Sensible heat polynyas are therefore locations of low ice production, and their size is determined by the dimensions of the warm water anomaly that generates them.

[4] Latent heat polynyas are mechanically driven and are created in areas where the ice motion is divergent because of the prevailing winds or oceanic currents. Because the water

inside the polynya is normally at the freezing point and heat is lost across the air-sea interface, ice is continually generated in the polynya region and advected away. By analogy with the oceanic sensible heat mechanism that creates sensible heat polynyas, it has been frequently stated that by slowing the freeze-up process, the latent heat released upon ice formation is responsible for keeping latent heat polynyas opened, thereby leading to their name. This argument is, however, misleading, as the latent heat of formation of ice is “used up,” as it were, to maintain the water at the freezing point, and therefore it exerts no control on the freezing rates. The process that actually keeps this type of polynya open is the export of ice. Consequently, latent heat polynyas would be better termed “wind-driven” or “current-driven” polynyas, depending on the cause of the ice divergence. Latent heat polynyas have high ice production rates, and their extent is governed by the balance between the export of ice from the polynya and the production of ice within it.

[5] Polynyas tend to appear recurrently at fixed geographical locations and periods of the year. Regions of upwelling or strong vertical mixing associated with tidal activity in bays, straits, and channels and regions of interaction between currents and topographic features, such as seamounts, are preferred sites for the appearance of sensible heat polynyas. Latent heat polynyas tend to form adjacent to coastlines swept by offshore winds and downwind or down current of landfast ice, glacier tongues, and grounded icebergs.

[6] A problem with the classification of polynyas into sensible and latent heat types is that many polynyas are

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formed by a combination of the two mechanisms (although one of them is normally predominant). Additionally, in spring and summer, polynyas are mainly driven by solar radiation entering the ocean surface and producing sea ice melt in and around the polynya area [Ohshima et al., 1998; Hunke and Ackley, 2001]. The distinction between sensible and latent heat polynyas at this time of the year is therefore not relevant. A pragmatic classification that does not present ambiguities is one based on the location of occurrence: “deep water” versus “shelf water” polynyas. Deep water polynyas are polynyas that form at or beyond the continental shelf break, while polynyas that occur over the shelves are shelf water polynyas (Figure 1). This classification is used in our review.

[7] Polynyas have the potential to play an important climatic role. First, polynyas are sites of strong ocean-to-atmosphere moisture and heat losses (up to several hundreds of W m^{-2}), which lead to a rapid warming of the air column above and downwind of the polynya and therefore to modifications in the mesoscale atmospheric motions [Walter, 1989; Kottmeier and Engelbart, 1992; Dethleff, 1994; Alam and Curry, 1995; Gallée, 1997]. In winter, about 50% of the total atmosphere-ocean heat exchanged over the Arctic Ocean occurs through polynyas and leads [Maykut, 1982]. During summer, leads (linear and narrow fractures within the sea ice cover caused mechanically by shear and divergent motion of the ice) and polynyas admit large quantities of shortwave radiation into the oceanic mixed layer, which ultimately impacts the heat and mass balance of the ice pack and the ocean [Maykut and McPhee, 1995; Maykut and Perovich, 1987]. Second, brine rejection during the formation of frazil ice (spicules or plates of ice suspended in water) within the polynya area increases the salinity of the upper ocean (up to several tenths of 1 part per thousand), which affects the baroclinic circulation in the vicinity of the polynya [Schumacher et al., 1983]. These cold saline waters may be transported by large-scale currents and have considerable impact far away from their source [Markus et al., 1998]. For instance, the Arctic halocline might be maintained by advection of cold saline water formed as a result of sea ice growth over the continental shelves of the Arctic Ocean and the Bering Sea [Aagaard et al., 1981; Cavalieri and Martin, 1994; Winsor and Björk, 2000]. In coastal polynyas, large buoyancy losses can cause downslope oceanic flows of dense, saline, and cold water. This is an important mechanism for ventilating deep and bottom water in both the Southern Ocean [Grumbine, 1991; Emms, 1997; Baines and Condie, 1998; Comiso and Gordon, 1998] and the Arctic Ocean [Schauer, 1995; Schauer and Fahrback, 1999]. Third, polynyas have effects on biogeochemical air-sea fluxes and on tracer transport by the ocean as a result of vertical mixing and convection. For example, polynyas are believed to act as sinks of atmospheric CO_2 because of both physical-chemical processes and biological activity [Yager et al., 1995; Bates et al., 1998; Arrigo et al., 1998, 1999]. In addition, turbulence and freezing within polynyas provide conditions for entrainment of sediments and pollutants

by frazil and anchor ice (underwater ice formed on a shallow ocean bottom or on a submerged object) as well as downslope transport along the ocean floor [Nurnberg et al., 1994; Reimnitz et al., 1994; Sherwood, 2000; Smedsrud, 2004]. Fourth, recurrent polynyas can be likened to “oases” in a polar ocean desert, as they support a diverse range of marine life forms from benthic populations through spring- and summer-blooming phytoplankton and zooplankton to overwintering mammals and birds [Grebmeier and Cooper, 1995; Smith and Gordon, 1997; Stirling, 1997].

[8] Projections of climate models suggest that if global warming occurs, polar regions are likely to experience temperature changes that will be faster and larger than in most other areas of the planet [Everett and Fitzharris, 2001]. Variations in the occurrence and size of polynyas might then be suitable indicators of any ongoing climatic alteration. Elucidating the interactions between polynyas and their environment is important for determining the role of high latitudes in global climate, especially with regard to deep ocean ventilation, and is without doubt one of the major challenges of present-day polar research.

[9] This paper reviews polynya studies carried out in the last 20 years. Polynya field experiments and models have been previously reviewed by Smith et al. [1990]. Compendia on the physics of polynyas, leads, and sea ice are given by Untersteiner [1986], Leppäranta [1998], and Wadhams [2000]. There is a growing body of descriptive research on polynyas encompassing in situ and remote sensing observations of polynya location and seasonality, the atmospheric and oceanic conditions that create and sustain them, their interactions with the boundary layers above and below, and their surface heat, fresh water, and ice budgets. Considerable understanding about the mechanisms of formation and maintenance of polynyas has also been gained through the use of models of diverse complexity, ranging from one-dimensional process models to coupled three-dimensional models of the atmosphere-sea ice-ocean system. All these studies will be highlighted here. Outside the scope of our paper is the description of polynya field experiments and observational tools. Nevertheless, it would be inappropriate not to make reference to the advances in polynya research afforded by developments since the 1970s in the remote sensing of sea ice [e.g., Zwally et al., 1983; Thomas et al., 1989; Gloersen et al., 1992; Johannessen et al., 1998].

2. LOCATION, SEASONALITY, AND VARIABILITY OF POLYNYAS

[10] The following describes many polynyas occurring around the Earth with references to the relevant literature. Readers who are primarily interested in the discussion of the physics of polynyas and their impact on the ocean and atmosphere may prefer to go directly to sections 3, 4, and 5 where polynya processes and impacts are reviewed.

[11] Recurrent polynyas owe their existence to the conjunction of geographical factors (ocean floor topography,

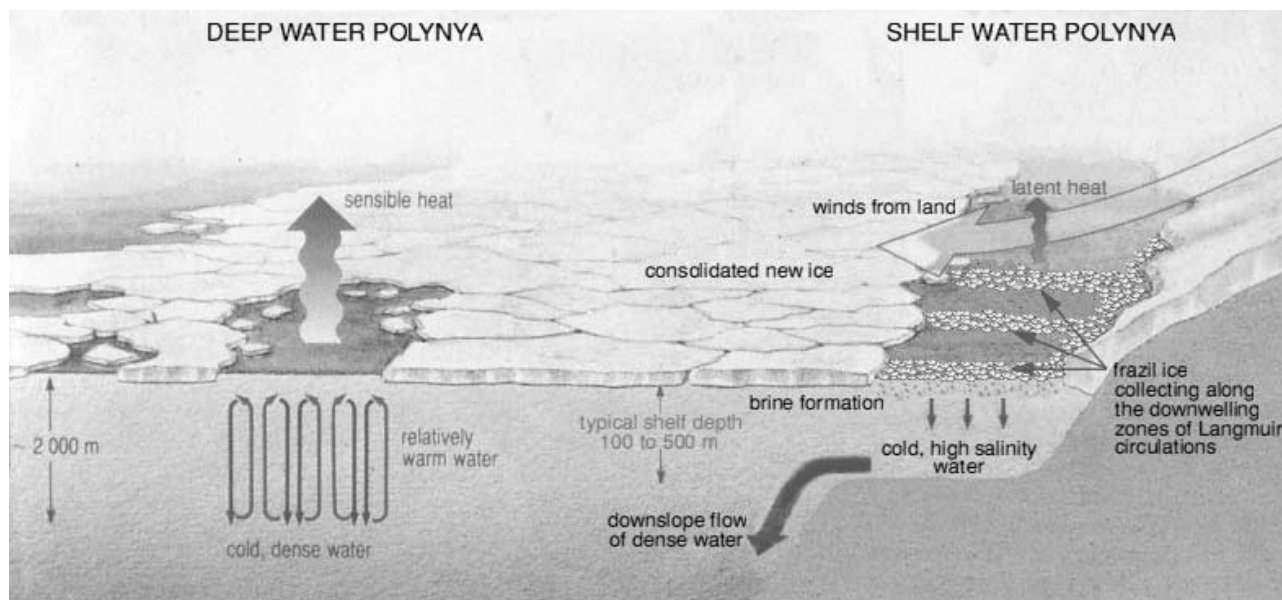


Figure 1. Schematic representation of physical processes taking place in deep water and shelf water polynyas. Deep water polynyas occur beyond the continental shelf break and are frequently created and maintained by a sensible heat mechanism, whereby upwelling or vertical mixing of subsurface water brings about sea ice melting or prevents ice from forming. Strong surface buoyancy losses through cooling can lead to deep water formation. Shelf water polynyas appear over the relatively shallow continental shelves. The majority of these polynyas are wind-driven. Offshore winds push the pack ice away from the coast, exposing the freezing surface waters to the cold atmosphere. Frazil ice is thus formed and herded downwind along Langmuir windrows. Brine-rich, cold water associated with sea ice formation will accumulate over the shelf and eventually flow down the shelf break slope to form deep and bottom water. Redrawn from *Open University Oceanography Course Team* [2001], with permission from Elsevier. See color version of this figure at back of this issue.

shoreline geometry, and coastal orography), atmospheric and oceanic features (strength, orientation and frequency of winds and currents, and distribution of air and water masses), and air-ice-water interactions (interfacial heat, moisture, momentum, and salt fluxes). Conditions favorable for the opening of polynyas occur intermittently within a year, or from year to year, and only in very particular locations. Figures 2 and 3 show all the geographical sites referred to in this paper. For the major polynyas described below, a summary of their types, together with a list of post-1990 references on observational and modeling studies, appears in Tables 1 and 2.

2.1. Deep Water Polynyas in the Southern Ocean

[12] Deep water polynyas are often formed by a sensible heat mechanism. They tend to occur in areas where surface cold, fresh waters are separated from underlying warmer, saltier waters by a weak pycnocline. Enhanced mixing or convection brings the warm water to the surface causing sea ice to melt or preventing it from forming. The evolution of polynyas formed in this way is controlled by their size [Comiso and Gordon, 1987]. The polynya will be terminated if excess lateral ice inflow occurs. For example, assuming that the ice around a polynya of radius 10 km is of depth 0.5 m and moves inward at 0.1 m s^{-1} , the maintenance of the polynya requires the oceanic heat flux to be $\sim 3000 \text{ W m}^{-2}$, which is unfeasible. In contrast,

a polynya of radius $\sim 300 \text{ km}$, such as the Weddell Sea Polynya of 1974–1976 (see section 2.1.1), requires a realistic $\sim 100 \text{ W m}^{-2}$ heat flux for its survival.

2.1.1. Weddell and Maud Rise Polynyas

[13] The giant Weddell Sea Polynya that occurred during the winters of 1974 to 1976 (Figure 4) is often mentioned as the archetypal deep water polynya. With a maximum extent of $\sim 350,000 \text{ km}^2$ [Carsey, 1980] this polynya, which has never reappeared, is the largest ever to be observed.

[14] There is scarce observational evidence upon which to establish the causes of the Weddell Polynya. Early modeling work [Martinson *et al.*, 1981; Motoi *et al.*, 1987] suggested that static instability promoted by cooling and brine rejection over the area of the polynya could have been sufficient to prevent ice growth. More recently, Akitomo *et al.* [1995] and McPhee [2000, 2003] have shown that the eastern Weddell Sea is an area susceptible to thermobaric convection, an instability due to the increase in the thermal expansion rate of seawater with pressure. This propensity to instability would favor the formation of deep water polynyas in the region.

[15] Other studies of the Weddell Polynya have stressed the importance of atmosphere-surface interactions in the development and maintenance of the polynya. According to Parkinson [1983], low surface turbulent heat fluxes associated with weak winds at the center of transient pressure systems over the Weddell region slow sea ice

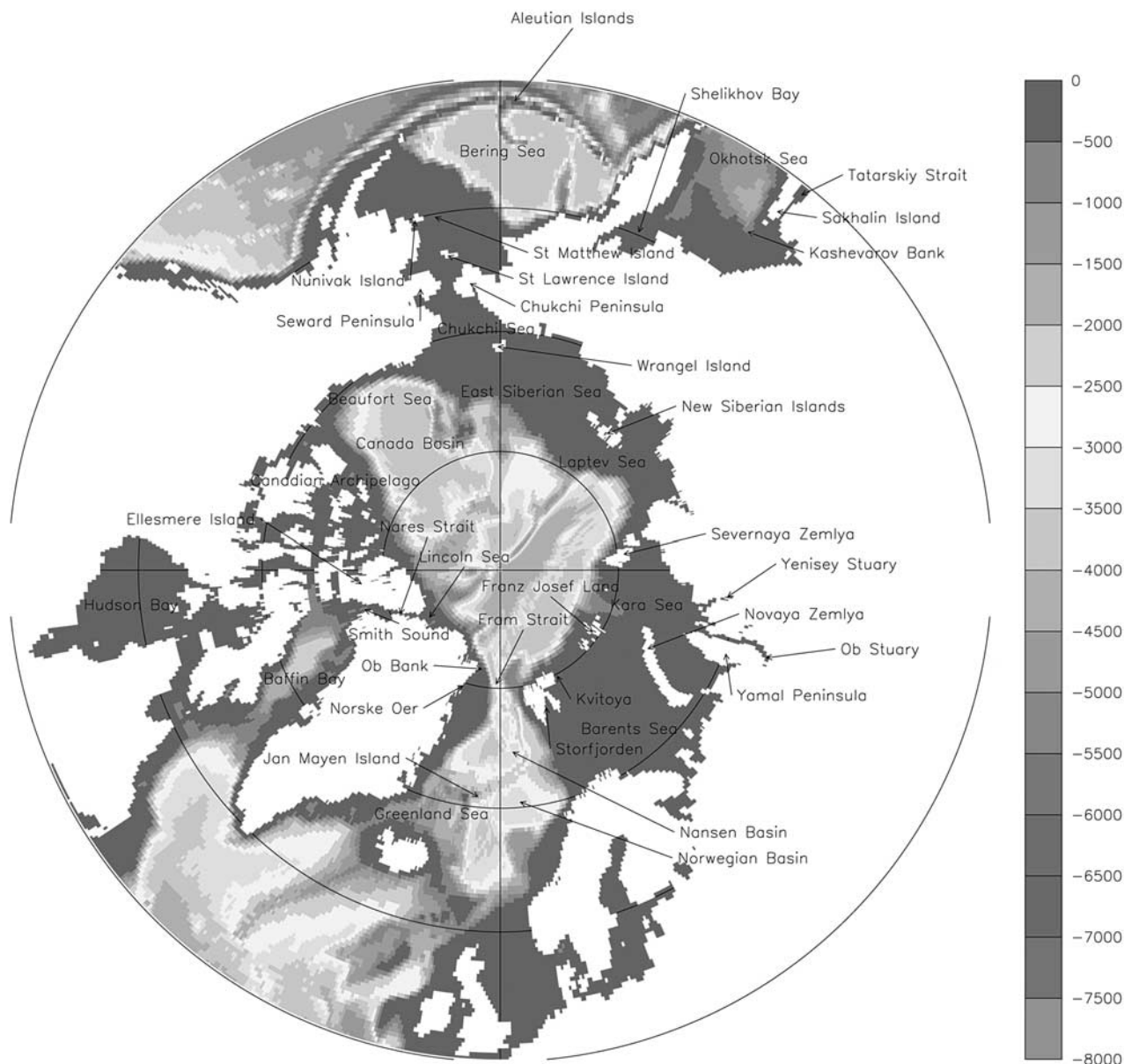


Figure 2. Northern Hemisphere geographical locations referred to in the text, together with bathymetry (in meters). See color version of this figure at back of this issue.

growth, thus making possible the appearance of the polynya. *Parkinson's* [1983] results were obtained with a dynamic thermodynamic sea ice model forced by observed winds, and therefore her simulations do not rule out an oceanic contribution to the polynya opening mechanism. *Parkinson* [1983] suggests that the lack of ocean coupling may explain the shorter than observed lifespan of the simulated polynya. *Lemke et al.* [1990] share the view that the longevity of the Weddell Polynya was sustained by anomalous meteorological or oceanic conditions.

[16] *Goosse and Fichefet* [2001] present a mechanism whereby strong wind-driven ice divergence in autumn creates conditions for hyperproduction of ice, thus destabilizing the water column and initiating the polynya. Variations in surface freshwater fluxes have also been advanced as a candidate for explaining large polynyas in the Weddell

Sea. *Marsland and Wolff* [2001] discuss coupled ice-ocean simulations in which a “Weddell Polynya” appears when the surface freshwater flux falls below a value of 0.35 m yr^{-1} . *Timmermann et al.* [1999] elaborate the theme of atmosphere-ocean interactions and explain the persistence of the polynya as the result of ice divergence associated with thermal wind cyclonic circulation, which, in turn, is stimulated by turbulent heat release over the polynya. Their simulations suggest that in cases where the atmospheric cross-polynya flow is weak ($<1 \text{ m s}^{-1}$), the thermal wind field can prolong the polynya event: A self-preserving low-pressure system develops and promotes repeated occurrence of the polynya.

[17] It does not seem probable that a divergent wind stress field alone can account for the presence of the Weddell Polynya, simply because the rate of sea ice export

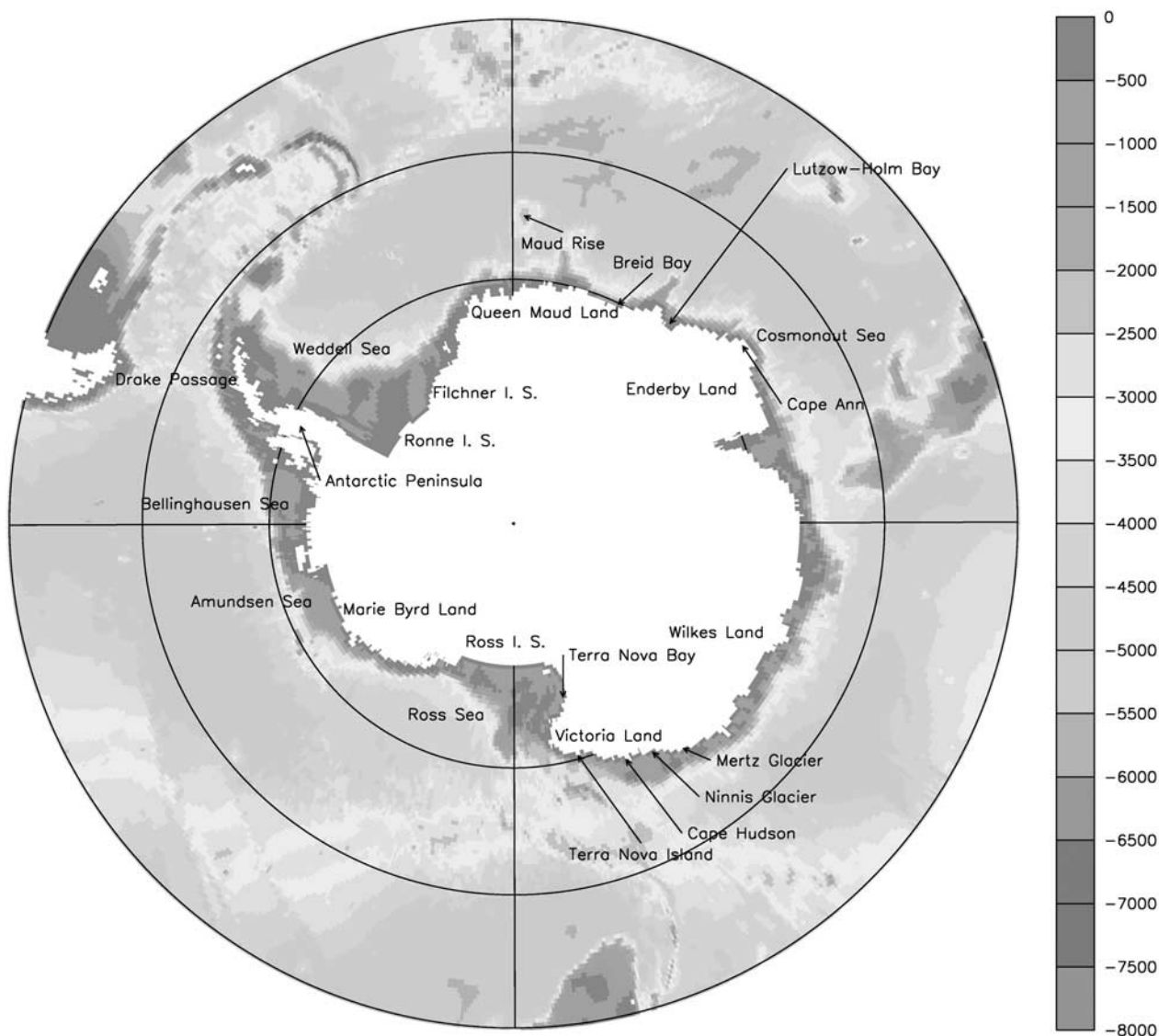


Figure 3. Southern Hemisphere geographical locations referred to in the text, together with bathymetry (in meters). Adélie coast (not shown) is the coastal area west of Victoria Land. See color version of this figure at back of this issue.

required to maintain the polynya in this fashion would require unrealistically large wind speeds. Assuming ice production rates of $\sim 10\text{--}12\text{ m yr}^{-1}$ [Cavaliere and Martin, 1985], unrealistically large wind speeds of the order of 20 m s^{-1} would be required to maintain the polynya by ice advection alone.

[18] A third line of research hypothesizes that the Weddell Polynya arose from the interaction between the oceanic circulation/stratification and the Maud Rise, a seamount located near the Greenwich Meridian rising from a depth of 5000 m to within 2000 m of the surface. Ou [1991] shows that when the vertical stratification falls below a critical value and, at the same time, the circulation is blocked by a seamount, sufficient surface cooling leads to deep convection and polynya formation. Muench et al. [2001] suggest, in contrast, that upwelling of warm deep

water induced by the interactions between the circulation and the topography is responsible for the polynya appearance. Another theory for polynya generation related to the presence of the seamount is that of Holland [2001a, 2001b], who demonstrates how variations in the oceanic flow past Maud Rise cause shedding of cyclonic eddies that transmit a divergent Ekman stress to the sea ice, leading to the creation of an area of open water. Holland [2001b] predicts that a dipole structure of cyclonic-anticyclonic vorticity is created by interactions of mean currents with Maud Rise and that this cyclone-anticyclone generation process has been detected in hydrographic data by Bersch et al. [1992].

[19] In a modeling study by Beckmann et al. [2001] the interaction between tidal motions, steady ocean currents, and winds leads to the generation of negative sea ice anomalies and positive mixed layer temperature anomalies

TABLE 1. Summary of Polynyas in the Northern Hemisphere^a

| <i>Polynya Location</i> | <i>Observational Studies</i> | <i>Modeling Studies</i> | <i>Remarks</i> |
|----------------------------------|--|--|---|
| Eurasian shelves | <i>Stringer and Groves</i> [1991], <i>Cavalieri and Martin</i> [1994], <i>Dehleff</i> [1994], <i>Pfimmann et al.</i> [1995], and <i>Winsor and Björk</i> [2000] | <i>Winsor and Björk</i> [2000] | shelf water, latent heat (wind driven) |
| Kashevarov Bank (Okhotsk Sea) | <i>Alfilitis and Martin</i> [1987] and <i>Martin et al.</i> [1998] | <i>Polyakov and Martin</i> [2000] | deep water, sensible heat (tidal control) |
| North Water | <i>Levuc et al.</i> [2001] | <i>Mysak and Huang</i> [1992], <i>Darby et al.</i> [1994], and <i>Biggs and Willmott</i> [2001] | shelf water, predominantly latent heat |
| Northeast Water | <i>Schneider and Budéus</i> [1994] and <i>Minnett et al.</i> [1997] | <i>Holland et al.</i> [1995] and <i>Willmott et al.</i> [1997] | shelf water, latent heat (current driven) |
| Okhotsk Sea shelves | <i>Wakatsuchi and Martin</i> [1990], <i>Martin et al.</i> [1998], and <i>Gladyshchev et al.</i> [2000] | | shelf water, latent heat (tidal control) |
| St. Lawrence Island | <i>Kozo et al.</i> [1990], <i>Stringer and Groves</i> [1991], <i>Cavalieri and Martin</i> [1994], <i>Grebmeier and Cooper</i> [1995], <i>Liu et al.</i> [1997], and <i>Lynch et al.</i> [1997] | <i>Lynch et al.</i> [1997], <i>Morales Maqueda and Willmott</i> [2000], and <i>Biggs et al.</i> [2000] | shelf water, latent heat (wind driven) |
| Storfjorden (south of Svalbard) | <i>Schauer and Fahrbach</i> [1999], <i>Haarpaintner</i> [1999], <i>Haarpaintner et al.</i> [2001], and Skogseth and Haugan (unpublished manuscript, 2003) | <i>Haarpaintner et al.</i> [2001] and Skogseth and Haugan (unpublished manuscript, 2003) | shelf water, latent heat (wind driven) |
| Whaler's Bay (north of Svalbard) | <i>Falk-Petersen et al.</i> [2000] | | deep water, sensible heat (upward mixing of Atlantic water) |

^aPre-1990 observational and modeling studies can be found within the references of *Smith et al.* [1990]. Further studies of the North Water can be found in the special issue on the International North Water (NOW) Polynya Study of *Atmosphere-Ocean* (39, 2001) and *Deep Sea Research II* (49(22–23), 2003). Further studies of the Northeast Water can be found in the Leads and Polynyas special section of the *Journal of Geophysical Research* (100(C3), 1995) and in the *Journal of Marine Systems* (10, 1997).

TABLE 2. Summary of Polynyas in the Southern Hemisphere^a

| <i>Polynya Location</i> | <i>Observational Studies</i> | <i>Modeling Studies</i> | <i>Remarks</i> |
|-------------------------------------|---|---|--|
| Cosmonaut Sea | <i>Comiso and Gordon</i> [1996] | <i>Bailey et al.</i> [2001] | deep water, sensible heat (upwelling of circumpolar water?) |
| East Antarctica shelves | <i>Massom et al.</i> [1998] and <i>Ushio et al.</i> [1999] | | shelf water, latent heat (wind driven) |
| Eastern Weddell Sea shelves | <i>Kottmeier and Engelbart</i> [1992] | | shelf water, latent heat (wind driven) |
| Ronne-Filchner Ice Shelf | <i>Ackley et al.</i> [2001], <i>Hunke and Ackley</i> [2001], and <i>Renfrew et al.</i> [2002] | <i>Hunke and Ackley</i> [2001] | shelf water, latent heat (wind driven) |
| Mertz Glacier (East Antarctica) | <i>Bindoff et al.</i> [2001], <i>Lytle et al.</i> [2001], and <i>Roberts et al.</i> [2001] | <i>Wu et al.</i> [2003] | shelf water, latent heat (wind driven) |
| Ross Sea Ice Shelf | <i>Bromwich et al.</i> [1993, 1998], <i>Jeffries and Adolphs</i> [1997], <i>Smith and Gordon</i> [1997], <i>Arrigo et al.</i> [1998], <i>Bates et al.</i> [1998], and <i>Gordon et al.</i> [2000] | <i>Fichefet and Goosse</i> [1999] | shelf water, combined latent heat (wind driven) and sensible heat (upwelling of circumpolar water) |
| Terra Nova Bay | <i>Gallée</i> [1997], <i>Van Woert</i> [1999b], <i>Budillon and Spezie</i> [2000], and <i>Budillon et al.</i> [2000] | <i>Darby et al.</i> [1995], <i>Gallée</i> [1997], and <i>Van Woert</i> [1999b] | shelf water, combined latent heat (wind driven) and sensible heat (upwelling of circumpolar water) |
| Weddell Sea 1974–1976 and Maud Rise | <i>Bersch et al.</i> [1992], <i>Parkinson</i> [1992], <i>McPhee et al.</i> [1999], <i>McPhee</i> [2000], and <i>Muench et al.</i> [2001] | <i>Lenke et al.</i> [1990], <i>Ou</i> [1991], <i>Akitomo et al.</i> [1995], <i>Timmermann et al.</i> [1999], <i>Goosse and Fichefet</i> [2001], <i>Holland</i> [2001a, 2001b], <i>Marstand and Wolff</i> [2001] | deep water, sensible heat |

^aPre-1990 observational and modeling studies can be found within the references of *Smith et al.* [1990]. Further studies of the Mertz Polynya can be found in the recent investigations of the Mertz Polynya and George V Land continental margin, East Antarctica, special issue of *Deep Sea Research II* (50(8–9), 2003).

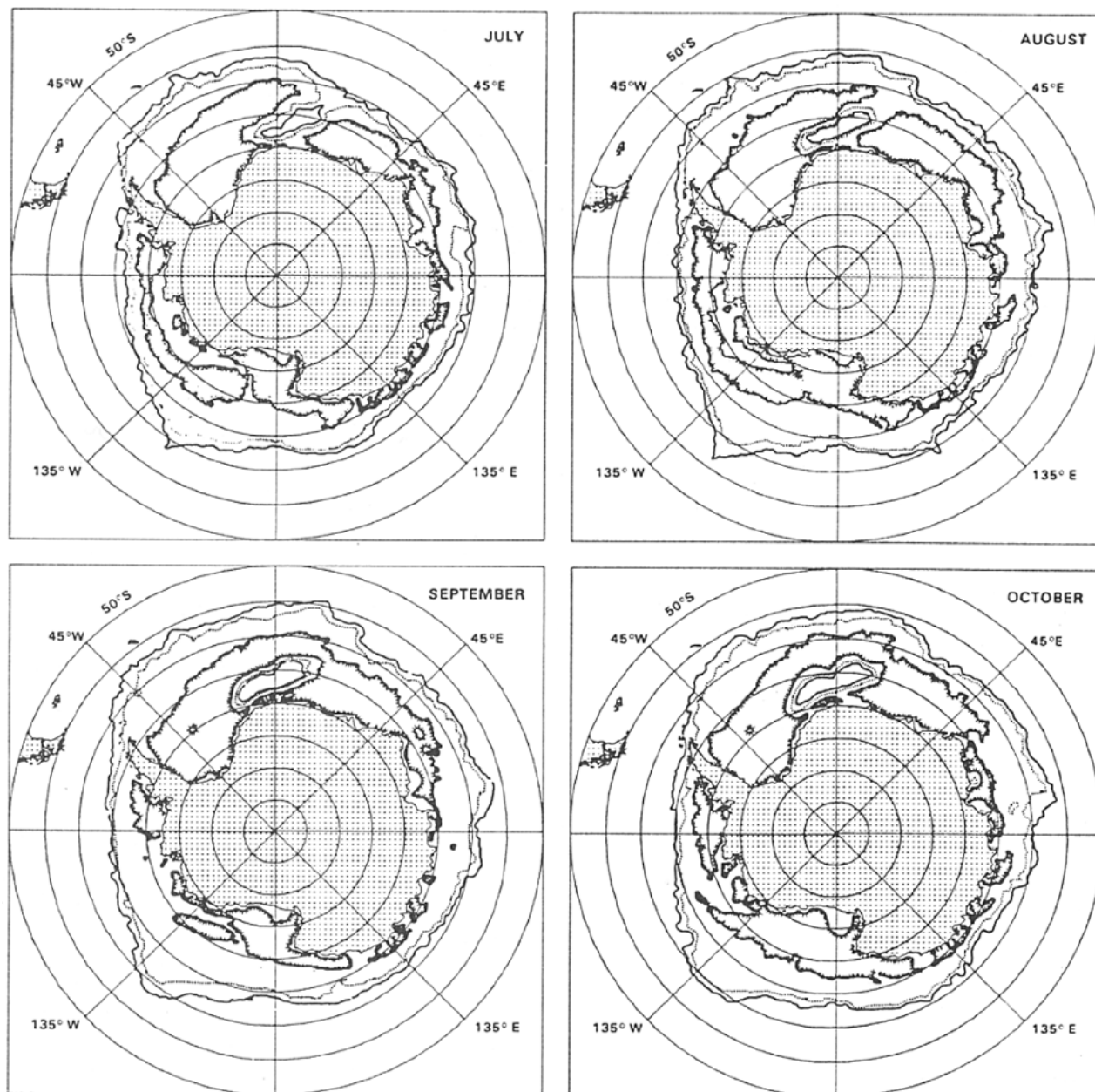


Figure 4. Monthly mean Southern Ocean ice concentration contour from July to October 1974, based on Electrically Scanning Microwave Radiometer brightness temperatures showing the Weddell Sea Polynya just offshore and centered on the Greenwich Meridian [Zwally *et al.*, 1985].

in the region. Tidal flow over Maud Rise enhances upward oceanic heat fluxes, while westward winds and steady ocean currents transport the anomalies downstream.

[20] The mechanisms advanced by *Holland* [2001a, 2001b] and *Beckmann et al.* [2001] do not fully explain the persistence and large scale of the mid-1970s Weddell Polynya, since the simulated sea ice anomalies have a shorter life and are of a smaller scale. Nevertheless, a polynya with characteristics close to those simulated by *Holland* [2001a, 2001b] has been observed over Maud Rise in 1980 [Comiso and Gordon, 1987] and 1994 [Drinkwater, 1996]. The mechanisms that account for the formation and decay of the polynya have not yet been identified, although

its existence has been attributed to storm-induced Ekman pumping and drift-induced turbulent mixing that seem to cause oceanic heat fluxes large enough to reduce ice concentrations. *Comiso and Gordon* [1987] argue that the primary process for the termination of the polynya is wind-driven inflow of ice into the polynya, where it melts, creating a low-salinity cap that suppresses vertical mixing.

2.1.2. Cosmonaut Polynya

[21] Satellite observations have revealed that a recurrent polynya forms east of the Weddell Gyre, offshore of Enderby Land. In this area, known as the Cosmonaut Sea, the water reaches depths of 3000–4000 m. The Cosmonaut Polynya has been observed virtually every year since 1973.

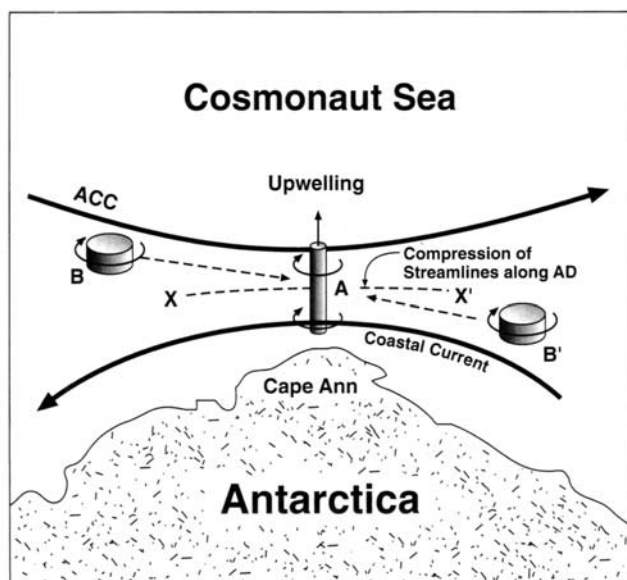


Figure 5. Mechanism for the Cosmonaut Polynya according to Comiso and Gordon [1996]. As water columns B (transported by the Antarctic Circumpolar Current (ACC)) and B' (transported by the Coastal Current) move both toward the polynya formation area A located over the Antarctic Divergence (AD), along XX' they must stretch in order to conserve absolute vorticity. Upwelling of warm deep water ensues. Reprinted from Comiso and Gordon [1996].

Comiso and Gordon [1987, 1996] and Gordon and Comiso [1988] have studied the polynya's morphology, behavior, and variability. The Cosmonaut Polynya is a polynya system with two patterns of formation. The first occurs only in early winter, is located west of the 45°E meridian, and seems to be caused by strong cyclone activity leading, presumably, to upwelling and the formation of an embayment of the ice edge. The second pattern develops farther to the east in both winter and early spring and is preceded by the formation of a coastal, wind-driven polynya. The maximum area of the polynya system oscillates between 35,000 and 115,000 km², with the effective area of open water within the polynya amounting to 30–45% of the total extent. The duration of a polynya event ranges from days to weeks, and the interannual variability of polynya size and timing is considerable. The polynya can occur with about equal probability any time between July and October. Comiso and Gordon [1996] note from an analysis of Special Sensor Microwave Imager (SSM/I) data from 1973 to 1993 that interannual variability of the polynya area about a long-term mean is ~65%.

[22] The occurrence of the western Cosmonaut Polynya seems to be linked to the action of storms that force ice to leave the region. The polynya ends as the center of the cyclones moves eastward, giving way to southerly winds. The eastern Cosmonaut Polynya might have an oceanographic origin. Comiso and Gordon [1996] surmise that the polynya may be caused by upwelling of warm water associated with a stretching of the water column that results

from compression of the westward flowing coastal current by the eastward flowing southern edge of the Antarctic Circumpolar Current (Figure 5, see also section 3). However, modeling and observational data that could support this mechanism still need to be provided. Bailey *et al.* [2001] suggest that the Cosmonaut Polynya does not form through ocean forcing alone. Rather, divergent wind stress would drive sea ice and ocean current divergence, the latter inducing upwelling that would maintain the polynya.

2.2. Deep Water Polynyas in the Arctic Ocean and Subarctic Seas

2.2.1. Kashevarov Bank Polynya

[23] A recurrent polynya has been revealed by satellite imagery over the 100-m-deep Kashevarov Bank, in the Sea of Okhotsk [Alfultis and Martin, 1987; Martin *et al.*, 1998]. In December an embayment of ice-free water appears over the bank. It persists for several weeks before being covered by sea ice or else becomes fully enclosed and is transformed into a polynya. Even when the polynya is not present, the area is often characterized by low ice concentrations (see charts given by Gloersen *et al.* [1992, pp. 46–71]). Tidal sea level oscillations and currents, together with the attending vertical mixing and upwelling, are important in the generation and maintenance of the polynya. Tidal currents of up to 2 m s⁻¹ cause widespread and intense vertical mixing that leads to an almost homogeneous vertical structure in salinity and temperature, bringing heat to the surface and causing polynya formation [Kowalik and Polyakov, 1998]. In addition, tidal residual currents create an anticyclonic ice drift that maintain the polynya [Polyakov and Martin, 2000].

2.2.2. Whaler's Bay Polynya

[24] The Whaler's Bay Polynya forms north of Svalbard, close to the ice edge, over the shelf break of the Eurasian Basin. The polynya is highly variable [Falk-Petersen *et al.*, 2000], and its formation is probably due to strong mixing of cold surface waters with warm (3°–4°C) Atlantic water transported at a depth of 200–300 m by the West Spitsbergen Current [Aagaard *et al.*, 1987].

2.2.3. Odden and Nordbukta

[25] Odden (Norwegian for “The Cape”) is an extension of the East Greenland Ice Stream that forms north of Jan Mayen Island over a deep oceanic region (2000–3500 m) located along the northeastward flowing Jan Mayen Current and that often extends beyond the Greenwich Meridian (see Figure 6). The Jan Mayen Current is cold and transports comparatively fresh water derived from ice melt, which facilitates sea ice formation in the area [Shuchman *et al.*, 1998]. Odden forms almost every year in November–December, reaches its maximum extent in February (250,000 km² on average and up to 330,000 km² sometimes), and normally disappears by May [Shuchman *et al.*, 1998; Wadhams and Comiso, 1999; Comiso *et al.*, 2001]. Associated with Odden and located to the northwest is a bight of relatively warm open water called Nordbukta (“North Bay”).

[26] Comiso *et al.* [2001] hypothesize that frazil ice is formed in the Nordbukta region and then removed from the

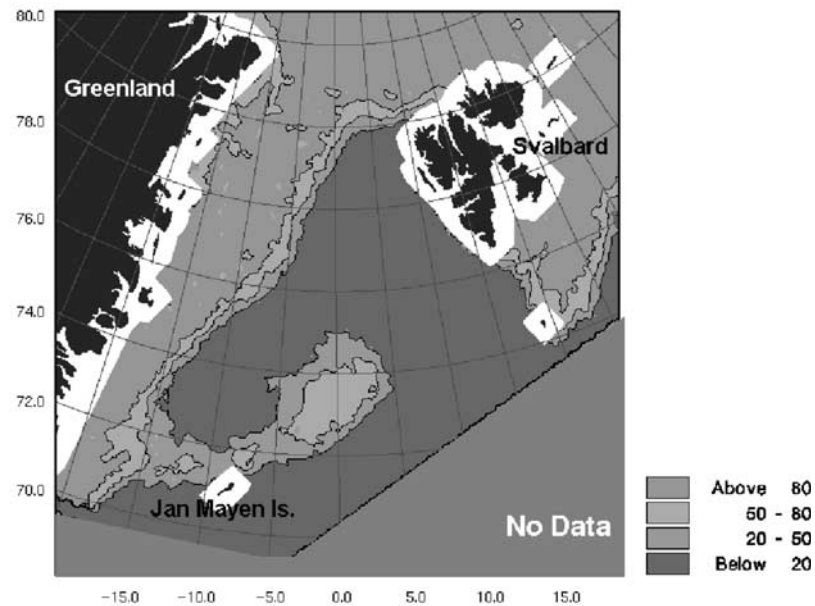


Figure 6. Ice concentration in the northern Greenland Sea from Special Sensor Microwave Imager (SSM/I) data on 16 March 1993. White areas are caused by masking of land. The Odden is the area of relatively low ice concentration stretching northeast from the landfast ice adjacent to the Greenland coast. Image courtesy of M. Lundhaug, Nansen Environmental and Remote Sensing Center, Bergen, Norway.

area and transported over to Odden by westerly winds, very much as ice is blown offshore from a coastal polynya and then collected at the polynya edge. In contrast to coastal polynyas, Odden is not surrounded by ice but by the open ocean and hence is exposed to wave activity and warm surface waters. As a result, Odden ice cover is fragmented and consists mainly of frazil ice and pancakes (circular pieces of ice up to 10 cm thick and 0.3–3 m in diameter). Note therefore that although Odden is frequently connected to the East Greenland Ice Stream, the source of its ice is mostly thermodynamic and hence local. Only on rare occasions (e.g., in 1987 and 1996) is first year or multiyear ice from the East Greenland Current advected during spring and summer into the region, where it melts [Wadhams and Comiso, 1999]. The Odden-Nordbukta system has a marked interannual variability. According to Shuchman *et al.* [1998] and Comiso *et al.* [2001] the two major controls of Odden-Nordbukta are the air temperatures, with lower temperatures leading to larger sea ice extents, and the wind, which influences the size and shape of the feature and the formation of the Nordbukta embayment.

[27] Carsey and Roach [1994] have inferred from remote sensing imagery the occurrence of convection episodes in the area during winter 1989. They attribute the creation of Nordbukta and polynya-like features in Odden ice cover to melting caused by convection.

2.3. Shelf Water Polynyas in the Southern Ocean

[28] Most shelf water polynyas are mechanically driven and fall therefore in the traditional class of latent heat polynyas. There are, however, exceptions. Many polynyas in the Canadian Arctic Archipelago, for example, are caused by surfacing of warm Atlantic water. In other cases, shelf

water polynyas are maintained by a mixed sensible-latent heat mechanism.

[29] Two conditions are required for the formation of a purely mechanically driven shelf water polynya. First, net divergence of the pack ice must occur. Second, surface heat losses need to be large enough to freeze the seawater, so that frazil ice formation can take place. The frazil ice is then removed from the region by the same mechanism that produced the initial ice divergence. The ice divergence is caused by winds and currents and necessitates the presence of an obstacle (e.g., a shoreline, landfast ice, an ice bridge, a glacier tongue, or an iceberg) that prevents the pack ice from invading the polynya from upwind or upstream. These polynyas form primarily along coastlines or in channels obstructed by sea ice, and, accordingly, the water column beneath the polynyas tends to be shallow and well insulated from oceanic sources of heat.

[30] We can distinguish two regions in a mechanically driven polynya (Figure 7): (1) an inner region of open water where frazil ice grows and (2) an outer region surrounded by first-year ice and occupied by new and young ice floes that have formed by accretion of frazil ice. Liu *et al.* [1997] call regions 1 and 2 the “active polynya” and “young ice” regions, respectively.

[31] Shelf water polynyas have been referred to as “ice factories,” because the constant removal of frazil ice results in large ice formation rates. Those polynyas that remain open during spring and summer tend to transform into “ice melt factories” as waters inside the polynya absorb large amounts of solar radiation, leading to melting of ice [Ohshima *et al.*, 1998]. The link between ice decay and solar heating is powerful and is not confined to polynya regions. Nihashi and Ohshima [2001a, 2001b] show that

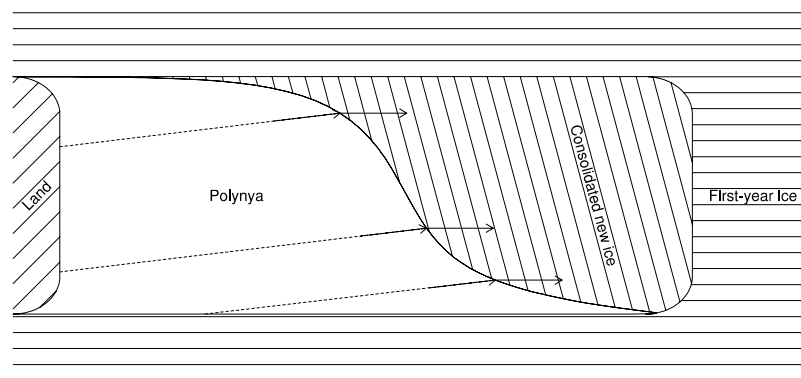


Figure 7. Schematic diagram illustrating a mechanically driven shelf polynya (latent heat polynya). Frazil ice grows in the area of nearly open water adjacent to the coast, the polynya proper, and is transported downwind, as schematically represented by the arrows inside the polynya area. The frazil ice arriving at the polynya edge piles up and forms consolidated new ice, which, in general, will move in a different direction from that of the frazil ice (arrows inside the consolidated new ice area). Adapted from Morales Maqueda and Willmott [2000].

sea ice melting in the Southern Ocean is driven by oceanic solar input, and they speculate that interannual variability of sea ice can lead to changes in the yearly storage of heat in the ocean, which, in turn, will affect the advance and retreat of the ice cover in following seasons and years.

2.3.1. Weddell Sea Polynyas

[32] Shelf water polynyas in the Weddell Sea are primarily wind-driven. Along the eastern flank of the Antarctic Peninsula, winter winds of up to 20 m s^{-1} cause polynyas to open. The oceanic area occupied by these polynyas decreases toward spring and summer, as the offshore component of the winds weakens and ice is pushed against the coast by the Weddell Gyre. Polynya seasonality along the seaward edge of the Ronne-Filchner Ice Shelf is similar to that of polynyas along the Antarctic Peninsula ice shelves, although winter polynya formation is larger there and more polynyas remain open in summer (Figure 8). Typical offshore polynya dimensions are 1–2.5 km [Markus *et al.*, 1998].

[33] The frequency of winter and spring polynyas along the Weddell Sea coast east of the Filchner Ice Shelf is comparable to that encountered in front of the ice shelves [Kottmeier and Engelbart, 1992]. Satellite-derived estimates of open water area adjacent to the coast during winter are 12%, 20%, and 16% for polynyas along the Antarctic Peninsula, the Ronne-Filchner Ice Shelf, and the eastern Weddell Sea, respectively [Comiso and Gordon, 1998]. However, the uncertainty of these figures is large (Markus *et al.* [1998] report values that are 20–40% smaller).

[34] The water-albedo feedback is crucial in controlling polynya area in spring and summer. Heating of the upper ocean by solar radiation prevents frazil ice formation and promotes melting at the polynya edge. At this stage the polynyas cannot reach equilibrium, and they increase in size for as long as winds are directed offshore. An example of such a summer polynya occurred in January–February 1998 off the Ronne Ice Shelf. The polynya extended 500 km north of the shelf occupying an area of $400,000 \text{ km}^2$. The

opening was driven by south-southwesterly winds and sustained by absorption of solar radiation [Ackley *et al.*, 2001; Hunke and Ackley, 2001]. Ackley *et al.* [2001] and Marshall and King [1998] speculate that the anomalous spring-summer event of 1997–1998 could be associated with a circulation pattern induced by the El Niño–Southern Oscillation (ENSO).

[35] Another case of a summer polynya in the Weddell Sea is a polynya that forms off Queen Maud Land. Ramesh Kumar and Sadhuram [1989] have found that solar radiation is the most important component of the polynya's surface heat budget, although it would appear that most of the heat absorbed by the ocean is advected away by tidal currents, resulting in stable sea surface temperatures.

[36] The interannual variability of polynya area in the Weddell Sea is thought to be related to atmospheric mesoscale and synoptic features: storm tracks, cyclones, barrier forced winds, and katabatic winds [Renfrew *et al.*, 2002]. It is unclear whether this variability is generated locally [Venegas and Drinkwater, 2001] or whether it is related to larger-scale modes such as the Antarctic Circumpolar Wave [White and Petersen, 1996]. Grounding of icebergs over shallow bathymetry, where they act as temporary islands, is also a cause for interannual polynya variability [Comiso and Gordon, 1998; Nøst and Østerhus, 1998].

2.3.2. Ross Ice Shelf and Terra Nova Bay Polynyas

[37] With an average (maximum) area of $27,000 \text{ km}^2$ ($50,000 \text{ km}^2$) the Ross Ice Shelf Polynya is the largest polynya to regularly form around Antarctica [Zwally *et al.*, 1985; Gloersen *et al.*, 1992]. It extends along the entire ice shelf, although its maximum width is attained westward of the Date Line. Between May and October the polynya waxes and wanes with a periodicity of roughly 1 month, and even when the polynya region is not ice-free, ice thickness tends to be small in this region (Jeffries and Adolphs [1997] reported 0.2 m in May 1995). The opening of the polynya has been attributed to the influence of

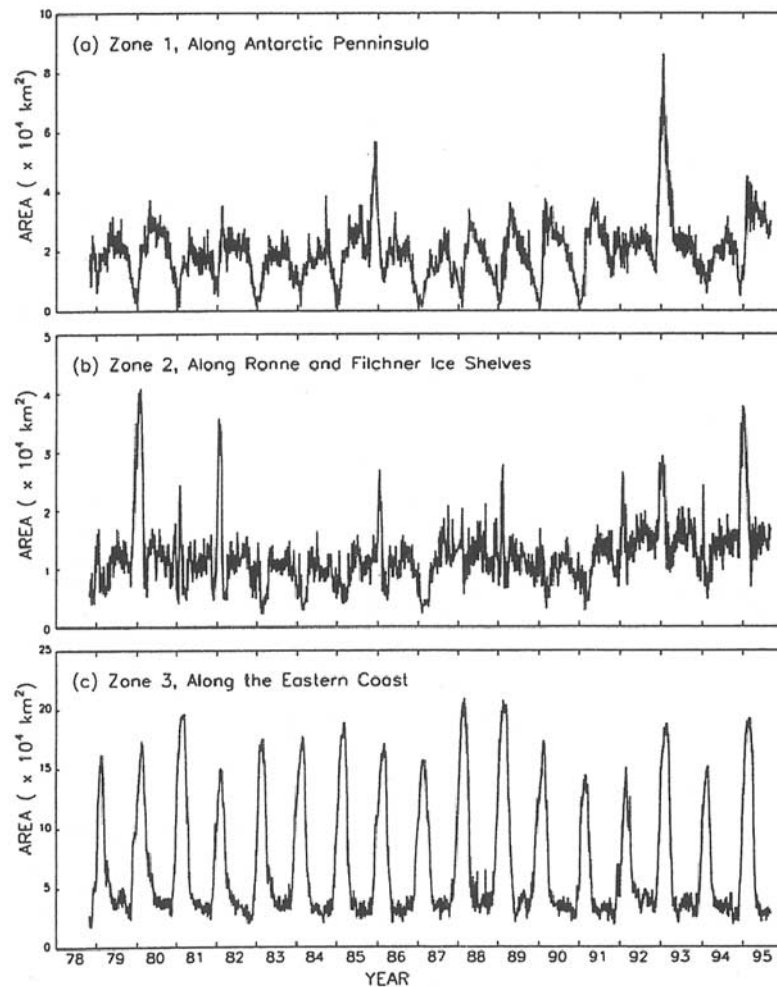


Figure 8. Open water areas in Weddell Sea coastal polynyas derived from SSM/I daily average data by Comiso and Gordon [1998]. (a) Polynyas along the Antarctic Peninsula. (b) Polynyas along the Ronne-Filchner Ice Shelf System. (c) Polynyas along the eastern Weddell Sea coast. Reprinted from Comiso and Gordon [1998].

synoptic weather systems that force ice divergence [Zwally et al., 1985; Jacobs and Comiso, 1989]. Bromwich et al. [1993, 1998] stress rather the role of northwestward katabatic airflow, with velocities $>30 \text{ m s}^{-1}$, in the formation of the polynya. They estimate that for the winters of 1988–1991, about 60% of polynya events are linked to katabatic surges. However, katabatic wind intensification is related to the passage of synoptic cyclones, and hence both weather systems and gravitational winds appear to be important in controlling polynya activity. The polynya is also influenced by upwelling of warm Circumpolar Deep Water onto the continental shelf [Pillsbury and Jacobs, 1985; Jacobs and Comiso, 1989].

[38] Commencing in early spring, the ice cover decays from south to north, and the polynya increases in size (see charts given by Gloersen et al. [1992, pp. 163–177]). By January the polynya edge reaches the ice margin. This rapid expansion is mostly due to warming of the upper ocean caused by absorption of solar radiation in the polynya area (heat gained by the ocean in January amounts to 196 W m^{-2} [Budillon et al., 2000]), although wind forcing

and upwelling probably continue to contribute to enhancing the polynya size [Fichefet and Goosse, 1999].

[39] The Terra Nova Bay Polynya, located between the Drygalski Ice Tongue and the land, is also created and maintained by katabatic winds. The polynya is oriented in a north-south direction, with the winds blowing from the west and the ice tongue preventing sea ice from invading the polynya from the south [Bromwich and Kurtz, 1984; Kurtz and Bromwich, 1983, 1985; Parish and Bromwich, 1989; Bromwich and Geer, 1991]. Van Woert [1999a] shows that sensible and longwave heat fluxes explain up to 50% of the observed variance in polynya area.

[40] The polynya area oscillates with a period of 15–20 days. The average area of open water is 1000 km^2 , and maximum values are ~ 5 times larger. Estimated cumulative ice production over the winter is around 40–60 m of ice ($50\text{--}80 \text{ km}^3$), which represents $\sim 10\%$ of the annual ice production over the Ross Sea continental shelf [Kurtz and Bromwich, 1985; Van Woert, 1999b]. The polynya is important for the modification of High-Salinity Shelf Water (HSSW), as brine plumes created during ice formation

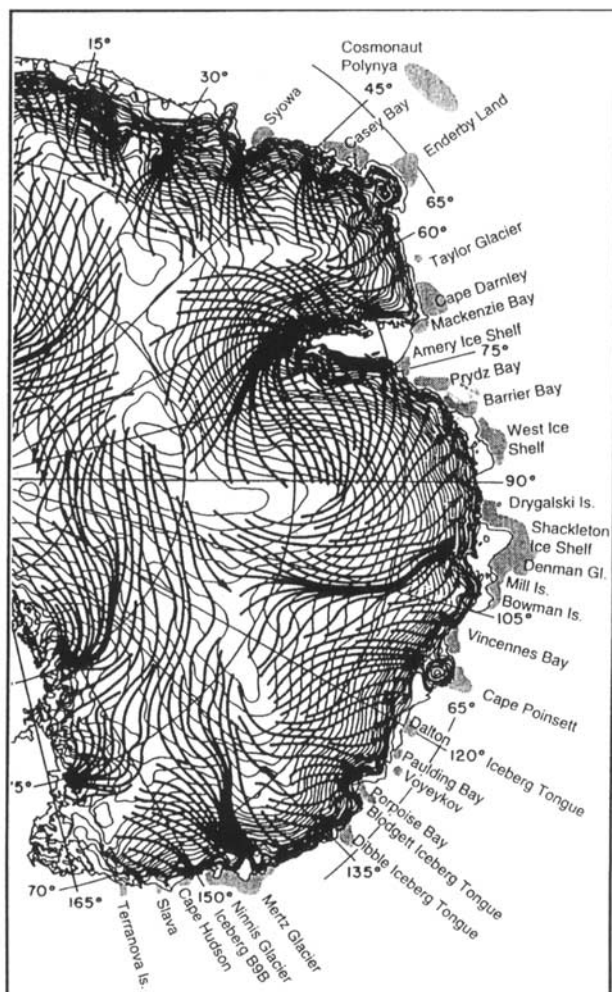


Figure 9. Coastal polynyas along the coast of East Antarctica. The grey shading indicates the approximate maximum extent of each polynya. Over the continent, thin lines are contours of surface elevation, and bold ones are streamlines of near-surface gravity drainage winds. From *Massom et al.* [1998]. Reprinted from the *Annals of Glaciology* with permission of the International Glaciological Society.

salinize the ~ 1 sverdrup (Sv) ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) of HSSW exported from the polynya [*Kurtz and Bromwich*, 1985; *Van Woert*, 1999b; *Budillon and Spezie*, 2000].

2.3.3. East Antarctica Polynyas

[41] *Massom et al.* [1998] used satellite imagery to map the distribution of 28 polynyas along the coast of East Antarctica between 1987 and 1994 (Figure 9). Their analyses revealed that the combination of bathymetry, land relief, and winds is responsible for the occurrence and characteristics of these polynyas. The polynyas appear in areas where north-south oriented obstacles, such as land or landfast ice protrusions, glacier tongues, or icebergs grounded over shallow banks and shoals, prevent the advection of sea ice into the polynya area from the east. Katabatic winds [*Adolphs and Wendler*, 1995; *Wendler et al.*, 1997] and synoptic winds provide the mechanical forcing to remove ice from the polynyas. In all cases,

polynya maximum areal extents occur in winter (June–October). Their average size oscillates between 1000 km^2 (Cape Hudson Polynya) and $23,000 \text{ km}^2$ (Mertz Glacier Polynya). Small polynyas tend to appear only occasionally during the year or not at all, while the larger polynyas are more stable and highly recurrent. Because the continental shelf around East Antarctica is not wide, the landfast ice margin is sometimes located over or beyond the shelf break. Oceanic mixing in coastal polynyas adjacent to the edge of the landfast ice can therefore connect surface and deep waters via turbulent entrainment. This is the case for the winter polynyas that appear in Breid and Lützw-Holm Bays off Enderby Land, where winter mixed layer depths of 350 m have been observed [*Ushio et al.*, 1999].

[42] In spring and summer these polynyas become regions of enhanced melt within the ice cover, so that the ice pack in East Antarctica decays simultaneously from the north and from the south. *Massom et al.* [1998] report two examples of this process occurring at the latitudes of Terra Nova Island and Mertz and Ninnis Glaciers. By December most coastal polynyas have merged with the open waters of the ice edge.

2.4. Shelf Water Polynyas in the Arctic and Subarctic Seas

2.4.1. Polynyas Over the Arctic Ocean Shelves

[43] Active and mostly wind-driven coastal polynyas occur every winter over all the seas peripheral to the central Arctic [*Gloersen et al.*, 1992]. Although these polynyas are generally small (their overall mean width is $\sim 6 \text{ km}$ [*Winsor and Björk*, 2000]), they are an important source of new ice growth and dense shelf water contributing to Arctic intermediate and deep water production [*Swift et al.*, 1997]. *Martin and Cavalieri* [1989], *Cavalieri and Martin* [1994], and *Winsor and Björk* [2000] have studied the seasonality and interannual variability of these polynyas with the goal of determining their contribution to the maintenance of the cold halocline layer of the Arctic Ocean (see section 3).

2.4.1.1. Barents Sea Polynyas: Storfjorden Polynya

[44] A persistent polynya has been observed over several winters in Storfjorden, a large bay situated in the south of the Svalbard archipelago [*Haarpaintner*, 1999; *Haarpaintner et al.*, 2001]. The polynya opens under northerly winds, and strong tidal currents probably help widen the polynya by breaking off ice floes from the landfast ice area that delineates its northern boundary. Remote sensing images for 1997/1998 (Figure 10) indicate that the polynya can be active over periods of weeks. At its largest extent it encompasses an area of 6000 km^2 of open water and new ice.

[45] The polynya is probably the source of brine-enriched waters flowing from Storfjorden toward the western shelf edge of the Barents Sea and into the Nansen and Norwegian Basins [*Schauer*, 1995]. Lee polynyas are also found west and southwest of Kvitøya and Franz Josef Land [*Falk-Petersen et al.*, 2000].

2.4.1.2. Kara Sea Polynyas

[46] A flaw polynya forms in winter seaward of the landfast ice that covers the estuaries of the rivers Ob' and

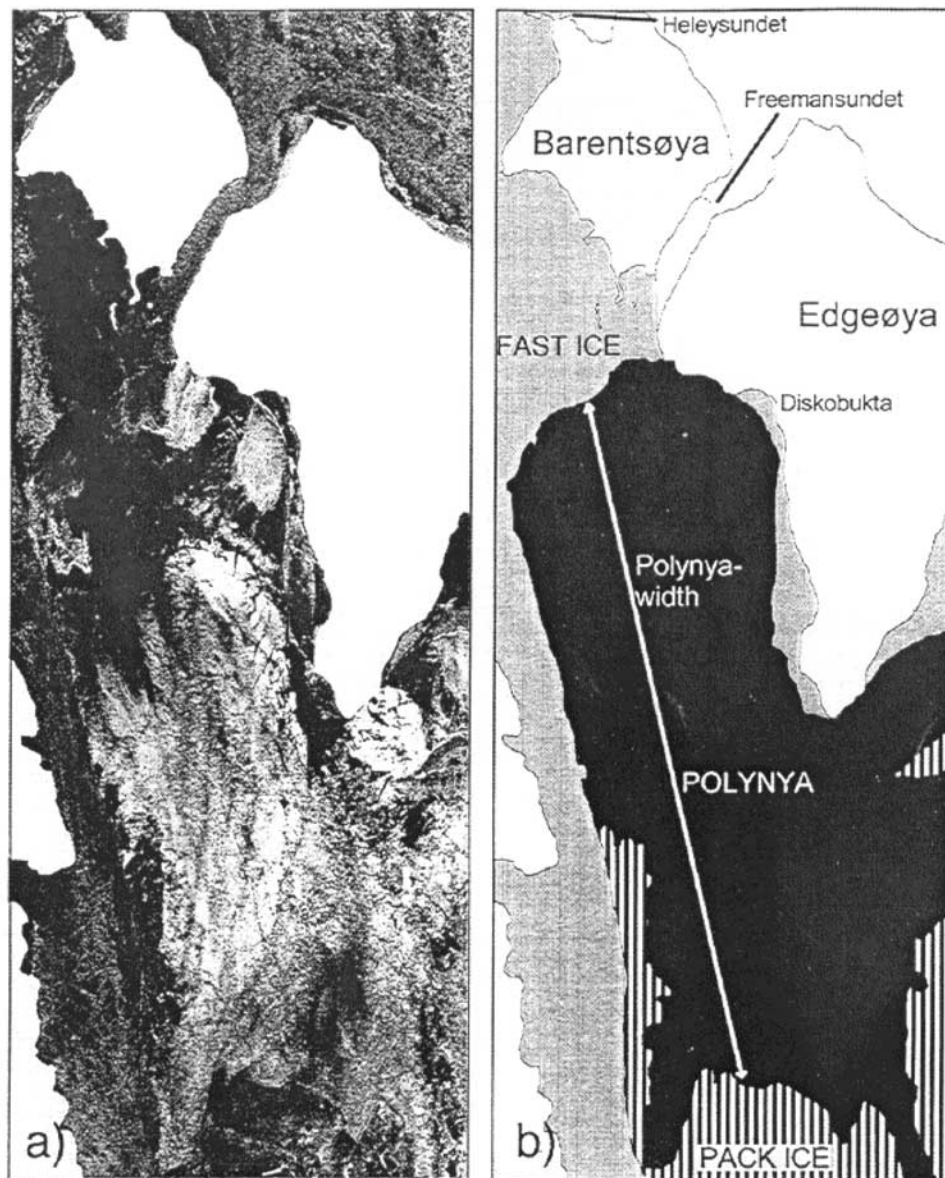


Figure 10. European Space Agency Remote Sensing Satellite (ERS-2) synthetic aperture radar estimates of the winter 1997/1998 partition of Storfjorden into fast ice, polynya, and pack ice normalized to an area of 10,000 km². Reprinted from *Haarpaintner et al.* [2001].

Yenisey. The polynya recurs in 80% of winters and extends in 65% of the cases west of the Yamal Peninsula [Buzov, 1991]. This zone is swept by offshore winds and is believed to produce most of the ice formed over the shelf (up to 1.5–2 m [Pavlov and Pfirmann, 1995]). An open water region coinciding with the location of the winter polynya is observed in summer and may represent increased melting due to the lower albedo in the remnant of the flaw polynya [Pfirmann et al., 1995].

2.4.1.3. Laptev Sea Polynyas

[47] The winter ice regime of the 500-km-wide Laptev Sea shelf is controlled by offshore Siberian winds that generate a quasi-perennial flaw polynya extending almost 2000 km along a shelf of landfast sea ice several hundreds of kilometers wide. The polynya width oscillates between 10 and 100 km. *Reimnitz et al.* [1994] speculate that this

polynya might result in the Laptev Sea being the major contributor to sea ice reformation in the whole Arctic, although the ice production estimates of *Winsor and Björk* [2000] suggest that Beaufort Sea polynyas might generate almost twice as much ice. With summer warming the Laptev Sea polynya turns into an area of high heat gain, resulting in the retreat of the ice edge to a higher latitude and greater distance (500 km) from the mainland than in any other area of the Arctic [*Reimnitz et al.*, 1994].

2.4.1.4. Chukchi Sea Polynyas

[48] *Stringer and Groves* [1991] have identified 10 recurrent polynyas in the Chukchi Sea. Eight of them are coastal polynyas, while the remaining two form in the open ocean over shoals. The prevailing winds in the vicinity of the coastal polynyas are from the north and east, and polynyas opening in the lee of coasts under these wind

regimes are frequent and persistent. Polynyas opening off north facing coasts appear only sporadically. Winds and currents might be equally important in explaining generation and extent of polynyas in the area. Polynyas in the Chukchi Sea have been identified by *Winsor and Björk* [2000] as contributors to deep water formation in the Arctic. Cold salty eddies have been observed below the Canada Basin mixed layer that might have been formed in association with polynyas along the Alaskan Chukchi Sea coast [*Muench et al.*, 2000].

2.4.1.5. Canadian Arctic Archipelago Polynyas

[49] A large system of polynyas forms in the shallow straits, bays, sounds, and channels of the Canadian Archipelago. Arctic surface waters flow southeastward, and their temperature and salinity are modified as they proceed toward Baffin Bay. Most polynyas result from tide-induced mixing in the channels that brings warm Atlantic water to the surface [*Melling et al.*, 1984]. A polynya appearing at the northern end of Nares Strait, over the Lincoln Sea, has, however, a hybrid origin [*Kozo*, 1991]. It occurs in autumn to the lee of a sea ice arch that forms immediately north of the strait. Newly formed ice is advected southward through the still open Nares Strait. Once the strait closes, open water areas, nevertheless, remain because of tidally induced upward mixing of warm water.

2.4.2. Bering Sea Polynyas

[50] Characteristic ocean currents on the eastern Bering Sea continental shelf are slow ($\leq 0.02 \text{ m s}^{-1}$ [*Kinder and Schumacher*, 1981]). Consequently, sea ice drift and coastal polynya formation in this region are dominated by wind systems [*Kozo et al.*, 1990]. Thus formation of polynyas adjacent to the generally south facing coasts of the St. Matthew, St. Lawrence, and Nunivak Islands and along the east-west oriented coastlines of the Chukchi and Seward Peninsulas takes place mainly as a result of north-northeasterly winds that push the thin ice cover of the Bering Sea southward [*Overland*, 1985]. Polynyas also form occasionally on the north facing coasts under southerly winds. These latter polynyas tend to be large because winds blowing from the south bring warmer and cloudier air masses over the region, thus reducing sea ice growth. In agreement with a mainly wind generation mechanism, when the north coast polynyas are open, the south coast polynyas are either not present or very small [*Stringer and Groves*, 1991].

[51] The polynya that opens along the southern coast of St. Lawrence Island has received particular attention [*Pease*, 1987; *Walter*, 1989; *Kozo et al.*, 1990; *Stringer and Groves*, 1991; *Liu et al.*, 1997]. The St. Lawrence Island Polynya (SLIP) is highly active. *Winsor and Björk* [2000] estimate a frequency of up to 30 polynya events per season, with a mean duration of 3 days and a maximum of 21 days. Owing to the form of the coastline the SLIP is often divided into two subregions that are separated by the 171°W meridian [*Cavalieri and Martin*, 1994]. The SLIP tends to be narrow ($\sim 10\text{--}20 \text{ km}$), but it can extend up to 270 km alongshore. Estimates of annual mean ice production in the SLIP oscillate between 22 and 32 km^3 [*Schumacher et al.*, 1983; *Cavalieri and Martin*, 1994; *Morales Maqueda and*

Willmott, 2000]. *Pease* [1980] calculated that the ice grown in the SLIP is enough to regenerate the entire Bering Sea ice cover 2–6 times per season.

2.4.3. Okhotsk Sea Shelf Polynyas and Polynyas Over the Sea of Japan

[52] Coastal polynyas occur along the eastern coast of Sakhalin Island, on the northwest and northern shelves of the Okhotsk Sea, and in Shelikhov Bay. These polynyas are caused by the north-northeasterly winds created by the Aleutian Low and their associated cold air temperatures (below -15°C). The polynya in Shelikhov Bay might also be controlled by tidal motions and mixing, as large tides (up to 13 m) occur in that region [*Kowalik and Polyakov*, 1998]. *Martin et al.* [1998] have determined the mean characteristics of the polynyas for the years 1990–1995. Taken together, the average open water area amounts to $25,500 \text{ km}^2$, and the area of surrounding thin ice totals $70,000 \text{ km}^2$. The polynya located along the northwest shelf has large interannual variability, since small changes in the almost alongshore geostrophic wind direction can lead to large variations in polynya width. Water modified in the polynyas contributes to a cold, low-salinity intermediate water mass in the Okhotsk Sea and a similar one in the Pacific [*Gladyshev et al.*, 2000].

[53] Northerly cold winds lead to low ice concentrations and enhanced ice production in the Tatarskiy Strait, Sea of Japan, between Sakhalin Island and the mainland [*Martin et al.*, 1992]. Each winter, storms create particularly strong northerly winds and low temperatures that lead to the formation of a transient polynya. The polynya is estimated to produce about 25% of the total amount of ice grown in a season and contributes to the generation of Sea of Japan bottom water. Interdecadal variability in the frequency of the storms suggests that the polynya might exhibit variations on similar timescales.

2.4.4. North Water

[54] The North Water Polynya (NOW) appears in northern Baffin Bay during winter and spring months. This polynya has recently been the focus of a Canadian-led field study conducted between 1997 and 1999. The physical processes group within the project has extensively reported their results [*Barber et al.*, 2001b; *Ingram et al.*, 2002]. Three processes are involved in the formation of the NOW. First, an ice bridge forms across the northern Smith Sound (southern Nares Strait) connecting Ellesmere Island and Greenland and prevents ice from the north entering Baffin Bay [*Dunbar and Dunbar*, 1972; *Dunbar*, 1981; *Mundy and Barber*, 2001]. Second, northerly winds channeled parallel to Smith Sound [*Muench*, 1971; *Ito*, 1982; *Barber et al.*, 2001c] and a southward flow of cold water from the Arctic [*Melling et al.*, 2001] mechanically remove ice from over the NOW area at a rate of around $4 \text{ km}^3 \text{ d}^{-1}$ or $600 \text{ km}^2 \text{ d}^{-1}$ [*Ito and Muller*, 1982]. Downwind fetch in the NOW can exceed 100 km at times [*Smith et al.*, 1990]. Third, oceanic upwelling contributes to sustaining low ice concentrations [*Steffen*, 1985, 1986, 1991], although studies on the formation of Baffin Bay bottom waters suggest that sensible heat accounts for a small amount of the heat loss to the

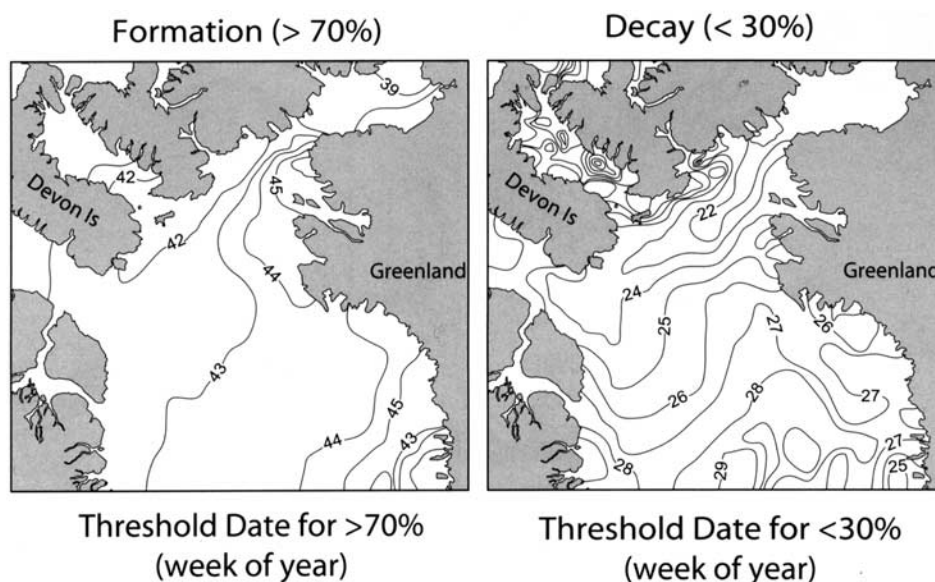


Figure 11. Average ice formation and decay in the North Water Polynya for 1979–1996. (left) Week of the year in which ice concentration becomes $>70\%$. (right) Week of the year in which ice concentration becomes $<30\%$. Note the late ice formation date along the western coast of Smith Sound (see Figure 1). Reprinted from Barber *et al.* [2001c] with kind permission of the Canadian Meteorological and Oceanographic Society.

atmosphere in winter and spring [Bourke and Paquette, 1989]. Mundy and Barber [2001] and Ingram *et al.* [2002] conclude a latent heat mechanism for the NOW, with the exception of the part of the polynya adjacent to the west Greenland coast, where advection of warm waters from the south and higher air temperatures lead to thinner winter ice.

[55] The NOW occurs in a fluctuating manner throughout winter and spring, driven by modifications in the Smith Sound ice bridge and the passage of synoptic meteorological systems. The polynya generally contains a considerable amount of ice, most of it consisting of thin young ice and thin-to-medium first-year ice [Steffen, 1986; Mundy and Barber, 2001]. Typical ice coverage in the NOW region is 80% between the second half of November and January, dropping to 60% in February, and rising again to $\sim 90\%$ in March; with the onset of ice melt the concentration decays steadily to values under 10% in July, at which stage the opening polynya merges with the reduced ice cover of Baffin Bay to create a predominantly open ocean area during August and September (Figure 11). Ice concentration starts increasing again in October [Gloersen *et al.*, 1992].

[56] Interannual variability in the NOW is large. A “bridge dipole” operates in Nares Strait and Smith Sound: When ice anomalies increase (decrease) in the former, they decrease (increase) in the latter. Between 1979 and 1996 this dipole has tended to operate more frequently in a negative (positive) mode in Nares Strait (Smith Sound), which suggests that the duration of the ice bridge has decreased annually over that period [Barber *et al.*, 2001c]. This raises the question of whether the stability of the bridge is related to the distribution of the ice floes (i.e., the percentages of first and multiyear ice) entering Kane Basin, which, in turn, could be related to the large-scale Arctic

Ocean circulation. Two distinct paths of the Transpolar Drift Stream on decadal and longer timescales are shown to be related to the large-scale atmospheric circulation in high northern latitudes [Polyakov and Johnson, 2000; Mysak, 2001]. In one path configuration the Transpolar Drift Stream directly transports sea ice and relatively fresh water from the Laptev Sea to Fram Strait. In the second path configuration the Transpolar Drift Stream is curved toward the Beaufort Sea before exiting the Arctic Ocean [Mysak, 2001].

2.4.5. Northeast Water

[57] The Northeast Water (NEW) Polynya appears over the northeast Greenland shelf between 77°N and 81°N . It is the northernmost of several polynyas along the east Greenland coast. A German-led study project on the NEW was carried out during three field seasons (1991–1993), and its results have been reported in the *Journal of Geophysical Research* (100, 1995) and the *Journal of Marine Systems* (10, 1997).

[58] Minnett *et al.* [1997] and Schneider and Budéus [1997] summarize the knowledge regarding the formation and seasonality of the NEW. The polynya owes its presence to two ice features: the Ob’ Bank Ice Shelf to the north and the Norske Øer landfast sea ice shelf to the south. The Ob’ Bank, made preponderantly of grounded icebergs and massive pieces of sea ice (floebbergs), prevents sea ice drifting southward through Fram Strait with the East Greenland Current from entering the polynya. The floating Norske Øer Shelf allows water transported by the northward flowing Northeast Greenland Coastal Current to flow into the polynya but not sea ice. Thus the region downstream of the barrier is protected from ice import while current-driven ice export continues [Schneider and Budéus, 1995; Johnson

TABLE 3. Annual Ice and Dense Water Production in Selected Polynyas

| <i>Polynya(s)</i> | <i>Annual Ice Production, m³ yr⁻¹</i> | <i>Dense Water Production, Sv</i> | <i>Source</i> |
|-----------------------|---|-----------------------------------|--|
| All Arctic | 6.8 × 10 ^{11a} 2.7–3.3 × 10 ¹¹ | 0.7–1.2 0.2 | <i>Cavalieri and Martin</i> [1994] <i>Winsor and Björk</i> [2000] |
| Okhotsk Sea shelves | 2 × 10 ¹¹ | 0.2–0.4 | <i>Martin et al.</i> [1998] |
| Southern Weddell Sea | 0.7–1.5 × 10 ¹¹ | | <i>Renfrew et al.</i> [2002] |
| Storfjorden | | 0.05 | <i>Schauer</i> [1995] |
| | 2.12–3.88 × 10 ¹⁰ | | <i>Skogseth and Haugan</i> (unpublished manuscript, 2003) |
| | | 0.11 | <i>Jungclaus et al.</i> [1995] |
| Terra Nova Bay | 5–8 × 10 ¹⁰ | 1 | <i>Kurtz and Bromwich</i> [1985], <i>Van Woert</i> [1999b], and <i>Budillon and Spezie</i> [2000] |
| Weddell Sea 1974–1976 | | 1.6–3.2 | <i>Gordon</i> [1982] |
| | | 1 | <i>Martinson et al.</i> [1981] |

^aValue is for Alaskan, Siberian, and Canadian coastal polynyas only.

and *Niebauer*, 1995]. Satellite imagery suggests that the area between the Henrik Krøyer Islands and Ob' Bank is characterized in late winter by a relatively thin ice cover, which could be interpreted as the fingerprint of a latent heat polynya created by the stormy outburst of offshore northeasterly winds [*Schneider and Budéus*, 1995; *Minnett*, 1995].

[59] In early summer, open water areas start developing over the NEW induced by high winds and a positive air-sea heat flux, which is controlled by shortwave and long-wave radiation, with summer mean net values of 105 and –24.5 W m⁻², respectively [*Minnett*, 1995]. The polynya stays open from May to September, with an area between 59,000 and 120,000 km². *Bohm et al.* [1997] have estimated that the heat budget explains up to 86% of the polynya interannual variability. Enhanced evaporation in summer leads to increased snow precipitation over the adjacent coastline [*Schneider and Budéus*, 1997].

3. WATER MASS TRANSFORMATION PROCESSES

[60] In the polar seas, water mass transformation occurs primarily in leads and polynyas. The three key water mass transformation processes are (1) surface heat and moisture loss, (2) brine input into the upper ocean associated with sea ice production during the freezing season, and (3) freshwater input into the upper ocean during the melt season.

[61] Polynyas on the continental shelves are sites of large sea ice formation. Episodic breakouts of cold polar air [*Backhaus et al.*, 1997; *Bromwich et al.*, 1998] generate polynyas in the lee of islands and ice tongues. On the continental shelves of Antarctica the dense water formed in polynyas ultimately flows onto the continental slope, mixing with ambient fluid as it does so [*Baines and Condie*, 1998]. Prior to reaching the slope, the dense water sometimes sinks beneath ice shelves [*Nicholls and Makinson*, 1998], thereby modifying the water mass characteristics. If the density of the mixed fluid on the slope becomes equal to the ambient density of the ocean at a given depth, it will leave the slope, forming Antarctic Intermediate Water. If the mixed fluid reaches the bottom of the slope, it becomes Antarctic Bottom Water (AABW), the densest water mass of the ocean (excluding some of the waters

formed in the Greenland, Iceland, and Norwegian Seas, which, nevertheless, do not spread to the global ocean, as they are constrained by the relatively shallow sills of the Greenland-Shetland passageway).

[62] The dense water formed on the Arctic continental shelves is thought to be responsible, in part, for maintaining the cold halocline layer [*Aagaard et al.*, 1981; *Martin and Cavalieri*, 1989; *Cavalieri and Martin*, 1994; *Winsor and Björk*, 2000]. This section contains a selection of studies on water mass transformation and polynyas. A wealth of literature exists on the subject, and the reader is advised that the exposition is not exhaustive. Direct measurements of the physical oceanography within polynyas are sparse, and consequently estimates of dense water production rates are often based on modeling studies that utilize whatever data are available. Table 3 presents annual mean sea ice production and dense water production rates for several Arctic and Antarctic polynyas. It is clear that at the beginning of the 21st century our knowledge of these quantities is still minimal.

3.1. Deep Water Polynyas in the Southern Ocean

3.1.1. Weddell and Maud Rise Polynyas

[63] Convective overturning within the Weddell Polynya of the mid-1970s produced deep water temperatures in 1977 typically 0.8°C colder than in 1973 (prior to the polynya formation) to a depth of 2500 m [*Gordon*, 1982; *Gordon and Comiso*, 1988]. A model estimate for the rate of dense water production within the Weddell Polynya is 1 Sv [*Martinson et al.*, 1981; *Marshall and Schott*, 1999] in comparison with the range of values 1.6–3.2 Sv estimated from observations by *Gordon* [1982]. *Ou* [1991] and *Alverson and Owens* [1996] theorize that trapping of water in Taylor columns directly above Maud Rise could precondition the region for open ocean convection. Intense surface cooling and brine input due to ice production reduce the temperature and increase the salinity of the shallow mixed layer to the point where the stable stratification is destroyed. At this point, vertical mixing brings warm salty water to the surface, thereby maintaining the Weddell Polynya.

[64] This picture of the formation and maintenance of the Weddell Polynya has two notable deficiencies. First, the area of the polynya was greatly in excess of the typical

area of open ocean convection cells [Killworth, 1979; Marshall and Schott, 1999]. It is possible that numerous convection cells (or “chimneys”) were located within the polynya. However, Marshall and Schott [1999] show that in the Labrador and Greenland Seas these cells are separated by regions of downwelling fluid. Perhaps the heat brought to the surface by mixing is subsequently diffused into the ambient fluid, thereby suppressing ice production over a region larger than the area of a chimney. Second, the location of the Weddell Polynya did not remain fixed; each winter it formed westward from its location in the previous winter. How did this phenomenon come about? There is speculation that water masses can to a large degree retain their stratification from one winter to the next, perhaps because of the long diffusive timescale of salinity. Thus, at the onset of spring the weakly stratified waters within the Weddell Polynya could be advected westward by the gyre circulation at the same time retaining their stratification. At the onset of the next winter a polynya is likely to form westward of its location in the previous winter. To date, no modeling study has correctly simulated the area and drift of the Weddell Polynya.

3.1.2. Cosmonaut Polynya

[65] The formation of the eastern Cosmonaut Polynya near Cape Ann at least once during the July to late August period is postulated by Comiso and Gordon [1996] to be regulated by the interaction of the Antarctic Circumpolar Current (ACC) with the westward flowing coastal current. In the neighborhood of Cape Ann the offshore width of the coastal current decreases as it is constrained to flow westward between the cape and the eastward ACC. The decrease in the coastal current width enhances the magnitude of the lateral current shear. Assuming that columns of fluid in the upper (mixed) layer of the coastal current conserve potential vorticity, the increase in magnitude of the lateral shear will lead to deepening of the fluid columns. Entrainment of warm deep water into the upper ocean then accompanies this deepening (Figure 5). However, the fact that the deep water in the Cosmonaut Sea is not anomalously cold indicates that convection does not occur in the polynya.

[66] The boundary separating the westward flowing coastal current from the eastward flowing ACC is called the Antarctic Divergence (AD). Along the AD, warm salty water is upwelled to the surface driven by the divergent surface wind stress field. Comiso and Gordon [1996] observe that the AD has its closest approach to Antarctica at Cape Ann, and this will tend to reduce sea ice production rates. Further, they speculate that cyclonic eddies within the AD create local upwelling, thereby forming highly transient polynyas.

3.2. Deep Water Polynyas in the Arctic Ocean and Subarctic Seas: Odden-Nordbukta

[67] The Iceland, Greenland, and Norwegian Seas are regions where open ocean convection creates deep water. The triggering of convection is sensitive to processes that determine the buoyancy of the upper ocean, such as surface cooling and the creation or melt of sea ice.

[68] From the point of view of predicting the upper ocean stratification it is crucial to determine whether the frazil and pancake ice that make up Odden is formed locally or whether it is advected into the area from the East Greenland Current. In the former case, there may be no net freshwater flux in the Odden region; brine input during the freezing season is mitigated by freshwater input in the melt season [Brandon and Wadhams, 1999]. In the latter case a net contribution of fresh water to the gyre is likely to occur, enhancing vertical stability and inhibiting convection.

[69] The ice-free Nordbukta is a region of convective activity and dense water production [Visbeck et al., 1995]. In early winter, ice forms in the bay and is advected to the southwest by northerly winds. Brine release induces “haline convection,” assisted by the ice export from the Nordbukta, reducing the insulation of the surface waters from the atmosphere. As the mixed layer deepens, it entrains warm intermediate Atlantic water. Local ice production is suppressed as warm water is mixed to the surface and the Nordbukta becomes established. At this time, Greenland Sea Deep Water is formed by “thermal convection” in the Nordbukta. Convection in the Odden-Nordbukta region has been reported to occur in several years during the late 1970s and throughout the 1980s [Shuchman et al., 1998].

3.3. Shelf Water Polynyas in the Southern Ocean

3.3.1. Coastal Polynyas in the Southern Weddell Sea

[70] The Ronne-Filchner Ice Shelf is one of the most active areas of the Weddell Sea for coastal polynyas [Markus et al., 1998; Comiso and Gordon, 1998]. Strong episodic offshore wind outbursts open polynyas adjacent to the ice shelf front. The presence of abundant frazil ice in sea ice cores from the Weddell Sea [Weeks and Ackley, 1986; Eicken and Lange, 1989] suggests that an important portion of Weddell ice originates in coastal polynyas. Comiso and Gordon [1998] note the coherence between peak values of coastal polynya areas and Weddell ice extent, which would indicate a strong role for polynyas in the production of sea ice. Indeed, while coastal polynyas in the southern Weddell Sea represent only 0.2% of the ice-covered area of the Weddell Sea [Markus et al., 1998], they account for between 2.5% and 9% of the total Weddell ice production [Renfrew et al., 2002]. The large heat losses over these polynyas (winter monthly average of 400 W m^{-2}) create ice production rates as large as $\sim 10\text{--}12 \text{ m yr}^{-1}$, which is a factor of 10 greater than ice production rates in the central Weddell Sea [Cavaliere and Martin, 1985; Zwally et al., 1985; Markus et al., 1998; Renfrew et al., 2002]. Renfrew et al. [2002] note that the total annual ice production within these polynyas has pronounced interannual variability, ranging from $0.71 \times 10^{11} \text{ m}^3$ in 1994 to $1.11 \times 10^{11} \text{ m}^3$ in 1995. However, given the uncertainties in both the observational data and modeling tools employed by Renfrew et al. [2002], the figures above must be taken with caution.

[71] Together with the Ross Sea [Koshlyakov and Tarakanov, 2003] and the area off Adélie Coast [Gordon

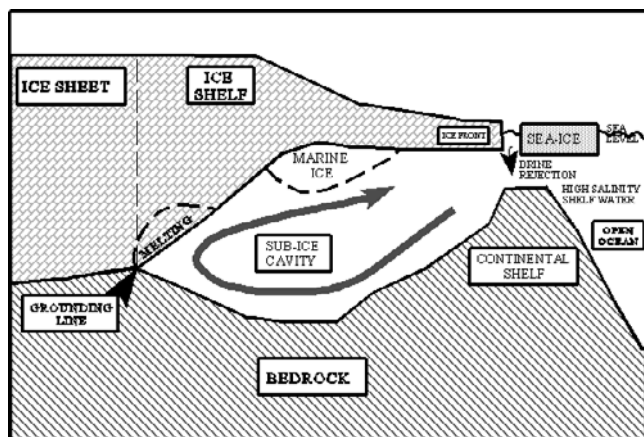


Figure 12. Schematic of circulation under an ice shelf. High-Salinity Shelf Water flows into the sub-ice cavity along the seabed and undergoes mixing and cooling through ice melting near the grounding line. The resulting less dense Ice Shelf Water (ISW) rises along the base of the ice shelf, and some of it freezes, forming marine ice at the base of the ice shelf. The remaining ISW emerges from the cavity and contributes to the formation of bottom water. Image courtesy of D. M. Holland (http://fish.cims.nyu.edu/~holland/project_oisi/motivation/ice_shelf_pump.gif).

and Tchernia, 1972; Rintoul, 1998; Fukamachi et al., 2000], the Weddell Sea is one of the few known regions where AABW is produced. Broecker et al. [1998] have estimated from chlorofluorocarbon inventories that deep waters formed along the Weddell Sea coast, presumably in polynyas, are 1–5 Sv, i.e., not more than ~30% of ventilated deep water flux in the whole Southern Ocean. At most sites where AABW is formed, coastal polynyas play a role in the production process [Foldvik and Gammelsrød, 1988]. In the Southern Weddell Sea, HSSW is formed in the polynyas adjacent to the Filchner-Ronne ice front. Intense cooling of the ocean and brine input during sea ice formation are responsible for the creation of HSSW [Nicholls and Makinson, 1998]. The fraction of HSSW that flows beneath the ice shelves is subsequently modified [Williams et al., 1998] (see Figure 12) by the exchange of heat, fresh water, and momentum at the ice shelf base, forming Ice Shelf Water (ISW). The primary outcome of the interaction of HSSW with the underside of the ice shelves is the production of a colder and fresher water mass [Baines and Condie, 1998].

[72] Seasonal and interannual variability of sea ice formation within the polynyas adjacent to the ice shelf front influences the water mass properties in the deep cavities beneath the ice shelves [Nicholls and Makinson, 1998]. For example, a reduction of ice formation leads to a reduction in the flux of warm water entering the cavities. As a consequence, ISW occupies a greater volume of the cavity. Thus warmer conditions at the ice shelf front lead to colder conditions beneath the ice shelf [Nicholls, 1997; Nicholls and Makinson, 1998].

[73] In the eastern Weddell Sea (30°W to 20°E), there is no evidence to support downslope flows of dense water,

formed originally in coastal polynyas [Baines and Condie, 1998]. The absence of these flows is attributed to the narrowness of the continental shelf [Fahrbach et al., 1994]. Coastal polynyas formed in this region would extend over the deep water. Brine associated with ice production is unlikely to accumulate because of the presence of the coastal current at the shelf edge flushing away the water within the polynyas. In addition, circumpolar deep water is more likely to intrude upon the continental slope, promoting melting.

[74] Although the offshore katabatic wind events are considered to be primarily responsible for the formation of polynyas adjacent to the Filchner-Ronne ice front, the impact of tidal currents cannot be neglected [Robertson et al., 1998; Makinson and Nicholls, 1999] in the maintenance of these features. Ice divergence created by tidal currents produces open water regions, thereby increasing the upper ocean salinity in response to sea ice formation [Kowalik and Proshutinsky, 1994]. Tides in the Weddell Sea also play a role in the production of AABW by increasing benthic shear stresses and mixing rates. The former process modifies the ocean circulation, while the latter process changes the water properties of the dense water that flows off the shelves.

3.3.2. Ross Ice Shelf and Terra Nova Bay Polynyas

[75] In a manner similar to the Weddell Sea polynyas, polynyas in the Ross Sea are formation sites of HSSW. Because HSSW evolves into other water masses (e.g., AABW), temporal changes in HSSW are likely to be observed in the characteristics or volumes of bottom water emanating from the Ross Sea. Since the 1960s, HSSW has freshened by 0.15 practical salinity units (psu) [Jacobs and Giulivi, 1998]. Sea ice extent in the Ross Sea oscillates with a 4- to 5-year period. The resulting interannual cycles of sea ice can be correlated with regional air temperature, shelf water salinities, and meridional winds. Since the 1950s, air temperature has increased at a rate of 0.25°C per decade; salinities on the shelf freshened by 0.03 psu per decade. Changes in the shape and location of polynyas might be responsible for these trends. Brine production in coastal polynyas is likely to vary with the dimensions of upstream ice tongues, icebergs, and ice fronts. The steady northward advance of the western side of the Ross Ice Shelf front may have gradually reduced the size of the Ross Sea Polynya. Some coastline changes will also move their adjacent low-salinity coastal currents, and this could affect upwelling and sensible heat sources maintaining polynyas. In addition, polynya development is often associated with katabatic surges. The interannual variability of the katabatic winds, possibly related to ENSO events [Bromwich et al., 1998], hence creates year to year variability in the dense water production within these polynyas.

3.3.3. Mertz Polynya

[76] The Adélie coast has been identified as a production site of AABW [Rintoul, 1998]. This region is not, a priori, a site where we would expect the production of AABW to occur. In contrast with the Weddell and Ross Seas, there are no broad shelves bordered by land to the west that assist in the accumulation of brine-enriched shelf waters. Indeed, the

continental shelf is relatively narrow. However, a major factor in the formation of AABW in this region is the presence of the Mertz Polynya, which forms in the lee of the Mertz Glacier Ice Tongue. Large amounts of brine are thought to be produced in this polynya. In addition, the salinity of the shelf waters is increased by intrusions of Circumpolar Deep Water, which probably help as well in maintaining the polynya. The Mertz Polynya has recently been the object of an ambitious international field experiment, results of which have been published in *Deep Sea Research II* (50(8–9), 2003).

3.4. Shelf Water Polynyas in the Arctic Ocean and Subarctic Seas

3.4.1. Polynyas Over the Arctic Ocean Shelves

[77] An important subsurface water mass in the Arctic is the cold halocline layer (CHL), characterized by temperature close to freezing and salinities that typically increase from 32.5 psu at 50 m to 34 psu at 150 m. The CHL thermally insulates the surface cold, fresh Arctic layer from the deep warm, saline Atlantic waters. The extent of the CHL has a major impact on sea ice. During the 1990s the CHL retreated from the Eurasian Basin into the Makarov Basin. *Steele and Boyd* [1998] conjecture that the retreat of the CHL was due to a shift in ice drift and upper ocean circulation forced by the strengthening of the Eurasian low-pressure cell. These changes are believed to occur on a roughly decadal timescale [*Proshutinsky and Johnson*, 1997; *Mysak*, 2001].

[78] Three possible mechanisms for maintaining the CHL are discussed by *Winsor and Björk* [2000]: first, lateral advection of cold, saline water formed on continental shelves by ice growth in the seasonal pack and polynyas; second, cooling and freshening of water from the Atlantic by ice melting [*Steele et al.*, 1995; *Steele and Boyd*, 1998]; and, third, formation of halocline water in the Eurasian Basin when Atlantic water is transformed by melting ice and atmospheric cooling in regions where deep convection operates.

[79] We will focus on dense water production in polynyas over the continental shelves of the Arctic Sea and its role in maintaining the CHL and ventilating the water beneath the CHL in the deep Arctic Basin. The combination of a perennial halocline and winter sea ice cover in the central Arctic prevents deep-reaching convection. However, slope convection of dense water masses produced on the shelves leads to significant ventilation of intermediate and deep water masses beneath the halocline [*Schauer*, 1995; *Schauer and Fahrbach*, 1999]. It is recognized that polynyas and leads on the shelves are important sites of dense water production, although the lack of hydrographic data from these sites prevents reliable estimates of the annual production rate of dense water.

[80] Attempts to quantify the annual dense water production from all the Arctic polynyas using remotely sensed sea ice statistics have been carried out by *Martin and Cavalieri* [1989], *Cavalieri and Martin* [1994], and *Winsor and Björk* [2000]. This exercise is fraught with problems such as lack

of in situ temperature and salinity data within the Arctic polynyas; difficulties in estimating the aggregate polynya areas during the freezing season from passive microwave satellite data [*Van Woert*, 1999a]; and low reliability and poor spatial coverage of meteorological data that are required to estimate ice production rates.

[81] *Cavalieri and Martin* [1994] estimate the average annual production of dense water within the polynyas of the western Arctic to be 0.5 ± 0.2 Sv. Interannual variability of this production rate of dense water is 40% of the annual average. The dense water production (with a salinity of 32.85 psu) for all the polynyas in the Arctic is estimated to be in the range of 0.7–1.2 Sv, with the Kara, Barents, and Chukchi Polynyas supplying the largest part and Chukchi Polynyas possibly contributing to deep water formation (Figure 13) [*Winsor and Björk*, 2000]. According to *Aagaard et al.* [1981] and *Björk* [1989, 1990], 1–2 Sv of dense water are required to maintain the CHL, and it appears that a significant fraction of this amount can be supplied by water mass transformation within Arctic polynyas. *Winsor and Björk* [2000] identify the Kara, Barents, and Chukchi Polynyas as the major sources of dense water, with the last being the only candidate for deep water formation.

[82] It is likely that the width of the shelf over which a polynya forms is a crucial parameter for determining whether the site is one of deep water production [*Schauer and Fahrbach*, 1999]. Although high-salinity shelf water is efficiently produced in the Barents and Laptev Seas [*Golovin*, 2002], with density comparable to that of the deep Arctic Basin, the distance of these dense water sources from the shelf edge is relatively large. As a consequence, as the dense water flows toward the shelf edge, there is considerable mixing with the ambient shelf waters, thereby reducing the density of the former. When the dense waters cascade down the slope, they do not reach the deep basin. *Backhaus et al.* [1997] also consider the role of sediments in possibly enhancing slope convection in the Arctic Ocean.

[83] The Storfjorden coastal polynya in the Svalbard Archipelago has been identified as a site where dense shelf water plumes are formed [*Schauer and Fahrbach*, 1999; *Haarpaintner et al.*, 2001; R. Skogseth and P. M. Haugan, Ice freezing and brine production in Storfjorden from four winters of satellite and in situ observations, unpublished manuscript, 2003, hereinafter referred to as Skogseth and Haugan, unpublished manuscript, 2003] through entrainment of Atlantic water. These plumes can double their volume en route to the shelf edge where they subsequently sink down to ventilate the deep Norwegian Sea. The water mass characteristics of the plumes vary from year to year [*Schauer and Fahrbach*, 1999; Skogseth and Haugan, unpublished manuscript, 2003], probably because of interannual variability of ice production in Storfjorden. Skogseth and Haugan (unpublished manuscript, 2003) propose that the strength of the southwesterly winds from the North Atlantic is an important factor in the ice production rate. When the southwesterly winds are strong, less Arctic ice is advected into the Barents Sea, resulting in higher surface

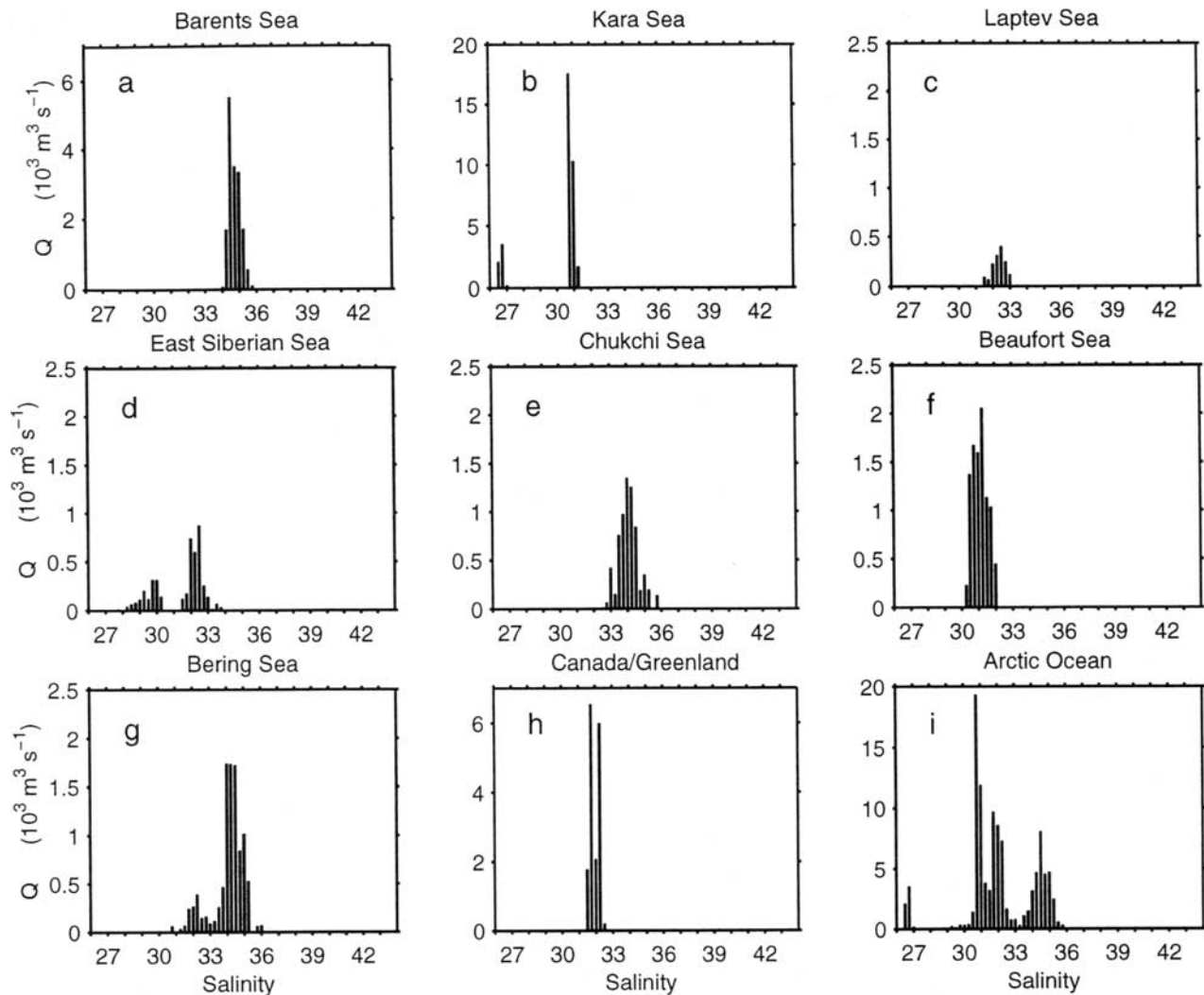


Figure 13. Estimates of flow rate distribution as a function of salinity for waters produced in polynyas in (a–h) different Arctic areas and (i) the entire Arctic Ocean. Polynya surface waters are assumed to be replenished by a constant surface current of 2 cm s^{-1} normal to the coast [Winsor and Björk, 2000]. Reprinted from Winsor and Björk [2000].

salinity in Storfjorden. It is possible that the North Atlantic Oscillation (NAO) could be implicated in the interannual variability of this region. Strong northerly (southerly) winds over Storfjorden are perhaps associated with a low (high) NAO index.

3.4.2. Bering Sea Polynyas

[84] Salt fluxes due to ice formation in polynyas over the shallow northern Bering Sea shelf modify water masses in the area. Since the circulation over this part of the Bering Sea is directed into the Arctic [Schumacher et al., 1983; Overland and Roach, 1987], the dense water thus formed is likely to contribute to the Arctic halocline [Aagaard et al., 1981].

[85] Typical ice production rates in the St. Lawrence Island Polynya are 0.1 m d^{-1} , equating to an aggregate winter ice production of the order of 5 m. Brine input into the shelf waters, associated with sea ice production, modifies the cross-shelf pressure gradient [Schumacher et al., 1983]. As a consequence, flow reversals of the shelf water

are observed during the winter from the mean weak east to west flowing current, with typical speeds of 0.035 m s^{-1} . These flow reversal events could also be partly wind-forced.

[86] If efficient vertical mixing over the shallow shelf of St. Lawrence Island (30–40 m) takes place, freshwater input during the melting season would be large enough to counterbalance the brine input during the winter. In practice, the counterbalance does not operate because the strong northward flow (current speeds $\sim 0.15 \text{ m s}^{-1}$) to the northwest of St. Lawrence Island is capable of transporting salinity-enriched shelf waters into the Arctic before the onset of the melt season. It is therefore understandable that Cavalieri and Martin [1994] included the dense water formed within the SLIP in the total estimate of the dense water formed within polynyas in the western Arctic.

3.4.3. Okhotsk Sea Shelf Polynyas

[87] The North Pacific Intermediate Water, an isopycnal layer characterized by a salinity minimum within the North Pacific subtropical gyre, is ventilated mainly from the

Okhotsk Sea and the Kuril Straits [Talley, 1997]. During the passage of North Pacific waters through the Okhotsk Sea, water mass modification takes place mainly in the coastal polynyas [Wakatsuchi and Martin, 1990; Martin et al., 1998]. Martin et al. [1998] estimate the dense water production to be in the range of 0.2–0.4 Sv, most of which takes place on the northern shelf of the Okhotsk Sea. The lower bound for this production rate is a significant fraction of the 0.5 Sv of dense water estimated by Yasuda [1997] to be required to maintain the North Pacific Intermediate Water. Large interannual variability of sea ice cover and dense water production in the Okhotsk Sea is likely to be related to interannual variability in the intensity of the Aleutian low [Martin et al., 1998].

3.4.4. North Water

[88] The waters passing through the Canadian Arctic Archipelago into Baffin and Hudson Bays from the Arctic are modified by buoyancy fluxes associated with ice growth and melt and land drainage [Ingram and Prinsenbergh, 1998]. Ultimately, these waters will reach the Labrador Sea and affect convective activity in the area [Marshall and Schott, 1999].

[89] Within the archipelago region is the annually recurring North Water Polynya (see section 2). The pycnocline within the NOW is generally too deep to be eroded by wind-generated turbulence [Melling et al., 2001], thus excluding oceanic sensible heat as a mechanism for the polynya maintenance, with the exception of one region. Along the Greenland coast the surface mixed layer depth generally attains a minimum. Potentially, the contributions from oceanic sensible heat in maintaining the NOW are likely to occur here, as the stratification of the mixed layer base is weak and the equatorward winds are upwelling favorable. Nevertheless, Melling et al. [2001] conclude that the warmest waters within the West Greenland Current remain subsurface. Vertical mixing of the warm subsurface waters into the mixed layer does not exceed 18 m d^{-1} [Melling et al., 2001], equivalent to a maximum heat flux of $300\text{--}400 \text{ W m}^{-2}$. Although this sensible heat flux sounds impressive, it is overshadowed by the magnitude of the latent heat liberated to the atmosphere. At best, oceanic sensible heat delays the onset of ice formation along the Greenland coast, but it is insufficient to promote ice melt. On the western side of the NOW (adjacent to Ellesmere Island) the stratification is stronger, and the entrainment rate of subsurface water into the mixed layer is correspondingly smaller than on the eastern side.

3.4.5. Northeast Water

[90] The pronounced upper ocean stratification within the NEW prevents production of bottom water [Budéus and Schneider, 1995]. An upper ocean water mass with temperature below 0°C and salinity below 34.4 psu, referred to as Polar Water (PW), is formed locally. The depth distribution of the PW is determined by a geostrophic balance with the anticyclonic gyre over the shelf. Thus the PW layer has a maximum depth (typically 140 m) at the center of the NEW. Budéus and Schneider [1995] speculate that the PW salinity is controlled in the long term by a balance between summer

freshwater inputs, winter brine-release inputs, and mixing with adjacent water masses. Below 150 m, water mass properties within the trough system traversing the continental shelf are controlled by salinities and temperatures at the trough entrances.

4. SURFACE HEAT AND MOISTURE FLUXES OVER POLYNYAS AND ATMOSPHERE-POLYNYA INTERACTIONS

[91] The existence of leads and polynyas within the ice cover greatly affects the air-ice-ocean interactions. During winter the turbulent heat loss to the atmosphere and the rate of ice growth depend strongly on ice thickness. For example, sensible and latent heat fluxes are up to 2 orders of magnitude larger over a refreezing lead or polynya than over the surrounding pack ice, which is usually snow-covered [Maykut, 1978, 1982; Ledley, 1988]. In the central Arctic, roughly as much ice is produced in winter in leads and polynyas (1% of the total pack ice area) as under perennial ice. In summer several times more shortwave radiation enters the upper ocean through leads and polynyas than through the ice, and this affects the mass balance of the ice cover and the oceanic heat balance [Maykut and Perovich, 1987; Maykut and McPhee, 1995]. Leads and polynyas also modify the surface freshwater and momentum balances since they allow a fraction of the falling snow to reach the ocean, enhance the average evaporation rates, and put the mixed layer in direct contact with the surface winds.

[92] Over open water or thin ice ($<10 \text{ cm}$ thick, so that ice thermal inertia can be neglected) the surface heat balance Q_{net} (positive if the water column gains heat) can be decomposed as

$$Q_s + Q_l + Q_{\text{lw}} + (1 - \alpha)Q_{\text{sw}} + Q_p = -(Q_o + L_f P_f) = Q_{\text{net}}, \quad (1)$$

where Q_s and Q_l are the turbulent sensible and latent heat fluxes, respectively, Q_{lw} is the net longwave radiation, α is the surface albedo, Q_{sw} is the downwelling shortwave radiation, Q_p is the heat flux associated with precipitation, Q_o is the upward oceanic sensible heat flux, L_f ($\sim 0.33 \times 10^6 \text{ J kg}^{-1}$) is the latent heat of formation of ice, and P_f is the sea ice production rate. Note that P_f differs from zero only if $(Q_{\text{net}} + Q_o)$ is negative and the ocean temperature is equal to or less than the freezing point T_f . Below we examine the relative importance of the different processes represented on the left-hand side of equation (1) in the evolution of a polynya. Table 4 displays a compilation of net turbulent and radiative heat fluxes over selected polynyas reported in the literature.

[93] If an initially ice-free ocean reaches the freezing temperature, supercooling and ice formation start. Nucleation of ice takes place on solid impurities and on snow crystals deposited on the surface [Weeks and Ackley, 1986], the ice particles thus created being mixed down to depths of several meters because of vertical turbulence [Martin and Kauffman, 1981]. However, the rate of ice production is at

TABLE 4. Net Turbulent and Radiative Heat Fluxes in Selected Polynyas^a

| Polynya(s) | Observation Period | Area, 1000 km ² | Turbulent Flux, W m ⁻² | Averaged Net Heat Flux | Radiative Flux, W m ⁻² | Source |
|----------------------------------|-----------------------------------|-------------------------------|--------------------------------------|---------------------------|--------------------------------------|---|
| Bennet Island | 18 February 1983 | 0.3–0.375 | –458 | | –105 | <i>Dethleff</i> [1994] |
| Dundas Island | March 1980 | 1 | –270 | | –60 | <i>Topham et al.</i> [1983] |
| | | | –270 | | | <i>den Hartog et al.</i> [1983] |
| Okhotsk Sea coast (southwest) | December 1978–1982 | 25 ^b | –471 | | –94 | <i>Alfultis and Martin</i> [1987] |
| Okhotsk Sea coast | January–March 1990–1995 | 25.4 ^b | | –256 | | <i>Martin et al.</i> [1998] |
| Queen Maud Land coast | December 1986 to February 1987 | ... | 3 | | 142 | <i>Ramesh Kumar and Sadhuram</i> [1989] |
| Ronne Ice Shelf | winter 1992–1998 | 4.893 | –272 | | –47 | <i>Renfrew et al.</i> [2002] |
| Ronne Ice Shelf | summer 1992–1998 | 36.20 | –1 | | 160 | <i>Renfrew et al.</i> [2002] |
| St. Lawrence Island | winter 1981 | ... | | –535 | | <i>Schumacher et al.</i> [1983] |
| St. Lawrence Island | February 1982–1983 | ... | –192/–412 ^c | | –79/91 ^c | <i>Pease</i> [1987] |
| Northeast Water | July–August 1992 | 10–18 | –7.1 ^d | | 80.9 ^d | <i>Minnett</i> [1995] |
| Northeast Water | spring 1993 | ... | –31 | | 195 | <i>Schneider and Budéus</i> [1995] |
| Storfjorden | 7 March 1998 | 6 | –300 | | –100 | <i>Haarpaintner et al.</i> [2001] |
| Terra Nova B. | May–September | 1 | –790 | | –38 | <i>Kurtz and Bromwich</i> [1985] |
| Weddell Sea coast (60°W–20°E) | May–October 1992 | 15 | –100/–300 ^e | | 50/–100 ^e | <i>Markus et al.</i> [1998] |

^aNegative values indicate heat loss from the polynya to the atmosphere.

^bValue does not include area of thin ice.

^cValues are for the 1982 and 1983 events, respectively.

^dValues are weighted by observed ice cover.

first small because of the tiny ice area exposed to the supercooled water. The supercooling therefore increases to temperatures of 0.05°–0.1°C below T_f until the volumetric concentration of ice in the water column becomes large enough so that more heat is released by freezing than is lost at the surface. The water temperature then rises to the freezing point, and an equilibrium is eventually reached, i.e., $L_f P_f = Q_{net} + Q_o$ [Omstedt and Svensson, 1984]. In relatively calm weather (surface wind speeds less than $\sim 5 \text{ m s}^{-1}$ [Pease, 1987]), ice reaching the surface will form a continuous thin skim of ice (congelation ice), which will later break into pans; under windier conditions the surface frazil ice will be herded downwind into a mixture of coagulating ice and salt water termed grease ice. Although the presence of new ice substantially reduces the atmosphere–ocean exchanges, significant fluxes still occur over the partly frozen oceanic area thus formed. An accurate knowledge of ice production rates and the accompanying surface fluxes is essential for the determination of the mass balance of sea ice and the energy budget of the atmosphere. Unfortunately, concurrent in situ observations of ice growth and surface exchanges over polynyas do not yet exist.

[94] From a meteorological point of view the moisture and turbulent fluxes resulting from air–sea interactions over a polynya during cold weather periods is paramount. Turbulent heat losses are the single most important component of the surface heat balance over winter polynyas, contributing up to 75% of the total heat loss [Winsor and Björk, 2000]. Values of sensible and latent heat fluxes over leads can exceed 400 W m^{-2} and 130 W m^{-2} , respectively [Andreas et al., 1979]. In contrast, sensible heat fluxes over mature central Arctic ice hover around 30 W m^{-2} (they are downward because of the strong winter polar inversion), while latent heat fluxes are negligible [Maykut, 1982]. The turbulent latent heat flux is normally several times smaller,

with typical values for the Bowen ratio (Q_s/Q_l) reported in the literature oscillating between 2 and 4 [e.g., den Hartog et al., 1983; Pease, 1987; Renfrew et al., 2002].

[95] Since (in a polynya) the ocean surface will normally be close to the freezing point during winter, variations in Q_s are almost exclusively associated with changes in air temperature and wind speed. Likewise, because the air specific humidity is very small at the low winter air temperatures, the surface specific humidity q_s dominates the humidity balance [Andreas and Cash, 1999]. However, q_s is equal or close to the saturation value at the freezing point, and hence wind speed is the main controlling factor of the latent heat fluxes. Pease [1987] notes that the saturation specific humidity for ice is slightly smaller than the saturation specific humidity for water (by up to 3 g kg^{-1}), so that cold air that is saturated with respect to ice upwind of a polynya becomes undersaturated with respect to the water over the polynya and finally becomes supersaturated with respect to the ice downwind of the polynya.

[96] Both sensible and latent heat fluxes depend in addition on fetch from the upwind polynya edge [Andreas and Murphy, 1986]. The reason for this is that a thermal internal boundary layer (IBL) is formed over the warm surface of a lead or polynya [Andreas et al., 1979; Topham et al., 1983; Andreas and Murphy, 1986; Walter, 1989; Glendening and Burk, 1992; Serreze et al., 1992; Andreas and Cash, 1999; Renfrew and Moore, 1999]. As the near-surface air crosses the polynya, it is gradually warmed and moistened, leading to a reduction in turbulent heat fluxes (of the order of 20% over fetches of tens of kilometers [Renfrew et al., 2002]) and a rapid growth of the height of the thermal IBL with fetch. Exchanges at the IBL upper boundary are weak, and therefore the air above tends to retain the characteristics it had before the polynya opened. In addition, for fetches smaller than $\sim 100 \text{ m}$, turbulent

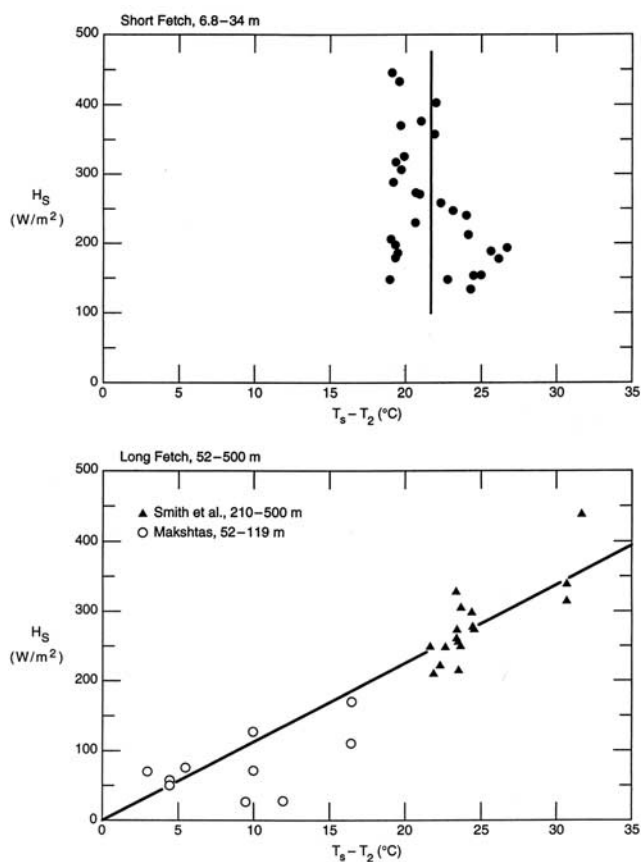


Figure 14. Sensible heat flux H_s and surface minus air temperature at 2 m ($T_s - T_2$) data for (top) short fetches from *Andreas et al.* [1979] and (bottom) long fetches from *Smith et al.* [1983] and *Makshatas* [1991]. Reprinted from *Andreas and Cash* [1999].

transfers are mediated by mixed forced and free convection, since the heat input from the surface is not enough to destabilize the atmospheric column during the short period it takes for the air to cross the warm polynya. In contrast, for large fetches, convective elements are given the opportunity to grow, and free convection then dominates the heat transfer [*Andreas and Cash*, 1999] (see Figure 14).

[97] As warm moisture-laden air rising from a polynya mixes upward in the IBL, it cools and recondenses, releasing the latent heat recently lost by the ocean surface. This process leads to fog layers over and downwind of the polynya that can rise tens to hundreds of meters [*Smith et al.*, 1983; *Walter*, 1989]. Upper tropospheric (~ 7 km) buoyant atmospheric plumes observed in the vicinity of Bennett Island (East Siberian Sea) in the winter of 1983 have been identified by *Dethleff* [1994] as originating from a large polynya in the area. Heat release to the atmosphere was estimated to be up to 683 W m^{-2} and would have caused strong turbulent updrafts and plume formation. Plumes, presumably produced by the same mechanism over Arctic leads, have been reported by *Schnell et al.* [1989]. Formation of moist convective plumes over Southern Ocean polynyas has been simulated by *Dare and Atkinson* [1999, 2000] with a nonhydrostatic turbulent kinetic energy model. The plumes extended from the surface to a height of 600–

1000 m above and downwind of the polynya area and were associated with the presence of convective clouds at the base of the elevated inversion over the polynya. It is worth mentioning that convective clouds and plumes are a feature frequently observed in satellite images of polynyas. *Fett et al.* [1997] present an instance of remote sensing detection of cloud development over the Laptev Sea Flaw Polynya. These authors used an algorithm based on visible data to establish the presence of open and refrozen water in the polynya area. The concurrent observation of cloud plumes, as inferred from midinfrared and infrared imagery, provided indirect evidence for verification and validation of their polynya detection technique. Plumes can exist even in the refreezing process, until the new ice becomes sufficiently thick so that moisture fluxes become negligible.

[98] The net longwave surface absorption can be decomposed into an upward emission from the ocean surface Q_{lwo} and an absorbed downward radiation from the atmosphere Q_{lwa} . Over a polynya that is entirely ice-free the surface temperature is near the freezing point, and therefore the range of variation of the outgoing longwave radiation is small. A typical value for Q_{lwo} is 298 W m^{-2} for a surface temperature $T_s = T_f = -1.8^\circ\text{C}$. However, if frazil ice accumulates inside the polynya, the emissivity of the surface can rise from ~ 0.96 (the emissivity of seawater) to ~ 0.99 (the emissivity of ice) and, for the same surface temperature as before, Q_{lwo} will increase to 304 W m^{-2} . It is instructive to compare these values for the longwave emission with those over consolidated sea ice. The winter mean temperatures of the sea ice cover in the Arctic and Antarctic are around -25°C and -10°C , respectively [*Comiso*, 1994]. The infrared emissions associated with these temperatures are 213 W m^{-2} and 269 W m^{-2} , respectively, namely, $30\text{--}85 \text{ W m}^{-2}$ weaker than the typical upward longwave radiation over open water. The absorbed downward longwave emission from the atmosphere is more variable than the upward longwave component, as the former depends not only on air temperature but also on water vapor content and cloudiness distribution up to heights of at least 30 km. Differences in Q_{lwa} between overcast and clear conditions are around 40 W m^{-2} for air temperatures within the range from -20° to 0°C (see Figure 15). The impact of clouds on Q_{lwa} has, however, some other important ramifications. Clouds play a leading role in determining not only the longwave but also the shortwave radiative transfer in the atmosphere, and cloud properties, such as area, height, optical thickness, water vapor content, and size and shape of cloud particles, are strongly coupled to sea ice and open water conditions and to the surface and planetary albedos. The coupling has two competing manifestations.

[99] 1. On the one hand, it leads to a positive feedback known as the sea ice/cloud/longwave radiation feedback, which can be outlined as follows. The presence of clouds leads to an increase in the downward longwave radiation that slows down freezing rates and can cause ice ablation, enhanced surface moisture fluxes, and cloud formation, thus supporting a positive feedback (let us mention that

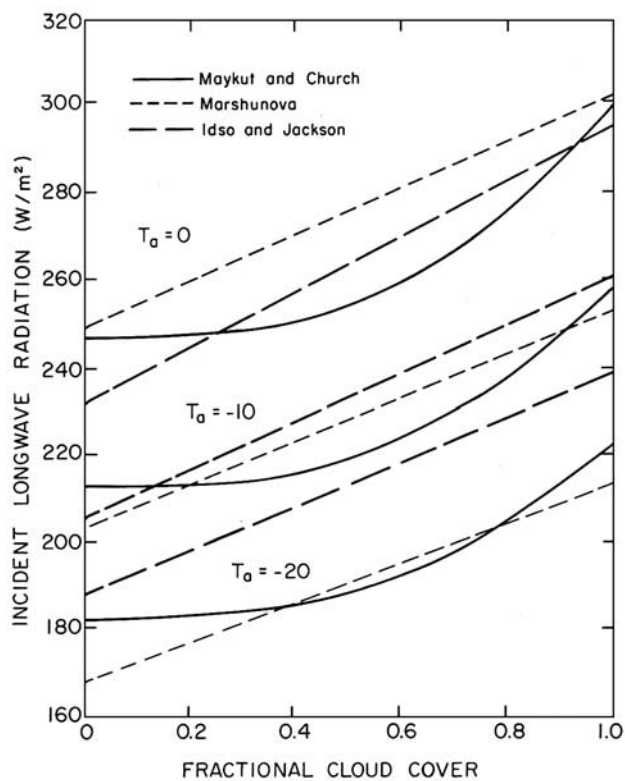


Figure 15. Comparison of the effects of cloudiness and air temperature ($^{\circ}\text{C}$) on incoming longwave radiation as predicted by parameterizations developed for polar conditions. Reprinted from *Maykut* [1986], with kind permission of Kluwer Academic Publishers.

an important source of downwelling longwave radiation in the Arctic during winter seems to be lower tropospheric ice crystal clouds, known as “diamond dust” [Curry and Ebert, 1992]). Conversely, the opening of a polynya, irrespective of its mechanism of generation, promotes local cloud formation, thereby moderating surface heat losses and fostering the expansion of the polynya.

[100] 2. On the other hand, clouds also contribute to the positive ice/cloud/shortwave radiation feedback. Owing to their high albedo, clouds formed over a polynya partly screen the surface from the incoming shortwave radiation, which leads to cooling and hinders polynya growth. Modeling studies indicate that in the Arctic the longwave radiation feedback is dominant, so that clouds have a net warming effect over the course of the year [Curry and Ebert, 1992; Uttal et al., 2002]. This is in contrast to oceanic areas at lower latitudes where clouds seem to cool the surface [Ho et al., 2002]. The difference is explained by the fact that solar heating is absent during a long part of the year and that the sea ice albedo is very high. Note, however, that the situation is different over a spring-summer polynya, where the surface albedo is significantly smaller. Indeed, a study on the effects of clouds on the net radiative forcing of the NOW polynya during spring and summer indicate that the cloud forcing of the surface radiation at this time of the year is mostly negative (i.e., clouds cool the surface when the Sun is in the sky [Hanafin and Minnett, 2001]).

[101] The net shortwave radiation is the dominant component of the surface heat budget in the summer period, at least during the part of the day when the Sun is high enough above the horizon. In the Arctic, for example, Q_{sw} attains maximum values of $\sim 320 \text{ W m}^{-2}$, with clear-sky peaks of up to $\sim 390 \text{ W m}^{-2}$ [Marshunova, 1966; Hanesiak et al., 2001]. The shortwave flux absorbed through polynyas and leads largely contributes to hindering springtime ice growth [Perovich and Richter-Menge, 2000] and driving summer melting. However, part of this heat can be stored in the oceanic pycnocline and can be released only in fall, thus retarding the growth of the ice cover [Perovich and Maykut, 1990]. The cloud cover largely controls the surface shortwave radiation budget. Minnett [1995] reports variations in the July–August midday insolation over the NOW Polynya from 500 to $<200 \text{ W m}^{-2}$ depending on whether the sky is clear or whether clouds or fog are present. Note that for clear skies the ocean surface albedo α depends on the zenith angle. Albedos as high as 0.33 occur when the Sun is close to the horizon, while at noon the albedo can descend to ~ 0.04 [Payne, 1972; Coakley, 1979].

[102] The effects on the heat flux associated with precipitation into polynyas have not been investigated in any detail. However, this component is likely to be very small in general. As an illustration, consider a quite large snow precipitation rate equivalent to the rate of 1 m of water per year. The latent heat needed to remove this snow amounts to only $\sim 10 \text{ W m}^{-2}$.

5. MODELING OF POLYNYAS

[103] Coupled sea ice, ocean general circulation models (OGCMs) do not have enough resolution, in general, to represent polynyas, and thus it is reasonable to enquire whether such models support realistic water masses and thermohaline circulation. This raises the question of whether it is possible to parameterize the water mass transformation processes that take place within polynyas in OGCMs. In order to develop a parameterization of this type it is vital to be able to predict how the area of a given polynya will evolve in time.

[104] A polynya model requires accurate inputs (e.g., wind stress field, cloud cover, surface ocean currents, and ocean temperature distribution) and realizations of the processes involved (e.g., frazil ice drift and pileup). Depending on the complexity of the model, some of the inputs and processes may be assumed to be unimportant and may be ignored or at best crudely parameterized. The quality of the output fields (e.g., polynya area and dense water production rates) is clearly related to the sophistication of the model adopted. Nevertheless, simple models can be of great use and can shed light on the key mechanisms.

[105] We now discuss in turn polynya flux models, models consisting of a dynamic-thermodynamic sea ice model coupled to an ocean and/or atmosphere model, and boundary layer models. Section 5.4 deals with the parameterization

of water mass transformation and ice production within OGCMs that do not resolve polynyas.

5.1. Flux Models

[106] Flux models for polynyas were first formulated by Pease [1987], embracing an idea of Lebedev [1968] that wind-generated coastal polynyas attain a maximum size. Lebedev proposed that this maximum size is determined by a balance between ice production within the polynya and the flux of ice out of the polynya.

[107] Polynya flux models idealize ice dynamics in a number of ways. Within the polynya interior, frazil ice is assumed to grow at a rate determined from the surface heat budget. The frazil ice is then herded downwind by the action of the surface currents and/or the wind stress and “piles up” as consolidated new ice at the offshore edge of the polynya. Pease [1987] describes a model for the opening of a coastal polynya in which alongshore variations in the forcing fields and the coastal boundary are neglected (a one-dimensional model). In the long-time limit the polynya approaches the equilibrium width X_{eqm} . In the special case when the offshore frazil and consolidated ice velocities, u and U , respectively, and the frazil ice production rate F are all constant, $X_{\text{eqm}} = HU/F$, where H is the thickness of consolidated new ice at the polynya edge, also assumed to be constant.

[108] Typically, U is taken as 3% of the wind speed, and $H = 0.1$ m. The frazil ice production rate F is calculated in terms of the radiative and turbulent heat fluxes at the ocean surface [Pease, 1987; Simonsen and Haugan, 1996]. Above a coastal polynya the typical surface air temperature is -20°C , the sea surface temperature is at the freezing point (-1.8°C), and $F \approx 0.34 \text{ m d}^{-1}$. In this case the equilibrium width is then $X_{\text{eqm}} \sim 18$ km, and the polynya takes about 1 day to open. Polynya size in the Pease [1987] model is a strong function of air temperature, but it is only weakly dependent on wind speed, since an increase in the latter increases both the ice advection rate and the frazil ice production rate, and the two effects cancel one another.

[109] A weakness of the Pease model is that frazil ice that forms within the polynya is assumed to instantaneously pile up against the downwind polynya edge. To overcome this deficiency, Ou [1988] develops a model incorporating a finite drift rate for frazil ice within the polynya, together with time-varying winds and air temperature. An equation for the polynya width is readily derived in the Ou [1988] model by noting that Lebedev’s [1968] ice balance hypothesis holds for all times when considered in the reference frame moving with the polynya edge itself. An alternative derivation of the polynya width equation is presented by Biggs et al. [2000].

[110] Mysak and Huang [1992] coupled a one-dimensional Pease-type model to a reduced gravity, coastal upwelling model in order to study the formation of the NOW in northern Baffin Bay. They concluded that the upwelling of comparatively warm water along the west coast of Greenland provides an upward heat flux that melts the ice near the coast and causes the polynya to have greater

meridional extent adjacent to Greenland than at its western edge at Ellesmere Island, as observed [Ito, 1982].

[111] Darby et al. [1994] extend the Mysak and Huang [1992] model by coupling a two-dimensional steady state polynya model to a nonlinear reduced gravity upper ocean model of northern Baffin Bay. The location of the polynya edge is determined by a balance between the components, normal to the edge, of the frazil ice flux reaching the polynya edge and the consolidated ice flux leaving the edge. A more realistic diverging channel geometry is also adopted, which better reproduces the northern reaches of Baffin Bay. It is found that, with the exception of late spring, the sensible heat flux associated with the upwelling along the Greenland coast plays no role in the maintenance of the NOW. Interestingly, the physical oceanography component of the NOW program [Barber et al., 2001a; Ingram et al., 2002] reached the same conclusions as this modeling study.

[112] Darby et al. [1995] and Willmott et al. [1997] investigate the fundamental behavior of the two-dimensional steady polynya model introduced by Darby et al. [1994]. Darby et al. [1995] derive analytical solutions for the steady state polynya edge adjacent to a variety of idealized coastal geometries. These solutions allow identification of an alongshore length scale (L_a), over which the polynya adjusts to its asymptotic width; perturbations in the coastline of length scale $\ll L_a$ are not reproduced in the shape of the polynya edge. By assuming a free-drift frazil ice momentum balance, Willmott et al. [1997] model the impact of surface ocean currents on the frazil ice trajectories.

[113] The Pease [1987] model has been used in a number of polynya modeling studies because of its simple formulation. Chapman [1999] uses a Pease-type model to calculate polynya size and surface buoyancy flux from air temperature and wind speed data in order to study the production of dense water below an idealized coastal polynya (Figure 16). The work extends previous studies [Gawarkiewicz and Chapman, 1995; Chapman and Gawarkiewicz, 1997] wherein the polynya is assumed steady. Subsequently, Chapman [2000] again adopts the steady state polynya assumption and investigates the influence of an along-shelf current on dense water production. The volume of dense water formed is given by Chapman [1999] to be a strong function of wind speed but a weak function of polynya width and atmospheric temperature. The ocean is shown to respond to the buoyancy forcing on timescales longer than those associated with polynya opening and closing (>10 days compared with 1–2 days), suggesting that reasonable estimates of dense water production may be obtained from steady seasonally averaged forcing. Chapman [2000] demonstrates that in the presence of an along-shelf current the basic ocean response is unchanged, and the total volume of dense water formed is not appreciably altered, although small-scale eddies are formed which carry the dense water from the polynya region.

[114] Markus and Burns [1995] compare the area of a polynya in the southern Weddell Sea estimated from an

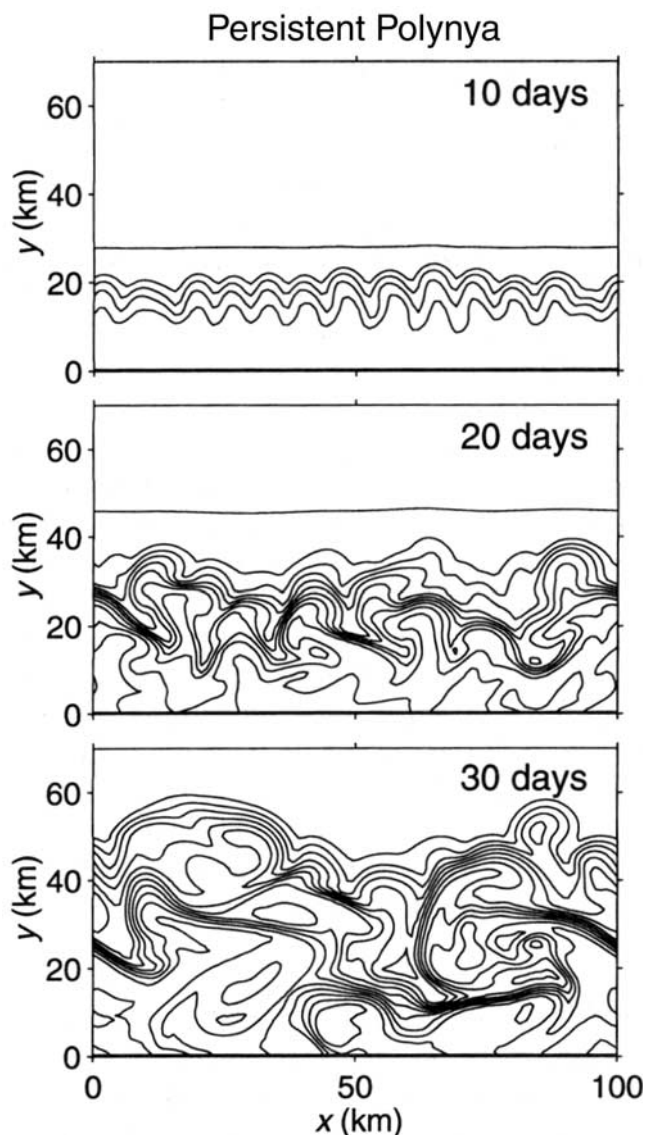


Figure 16. Density anomaly at ocean bottom given persistent periodic forcing. Contours are $0.075\text{--}1.2$ by 0.075 kg m^{-3} with the minimum contour farthest offshore (the coastline is at $y = 0$ km). Polynya width is 65, 25, and 70 km after 10, 20, and 30 days, respectively. Reprinted from Chapman [1999].

analysis of passive microwave data with the area calculated using the one-dimensional Pease [1987] polynya model. Variants of the Pease [1987] model have also been used to model the Terra Nova Bay Polynya [Van Woert, 1999b]. Van Woert [1999b] compares simulations of three variants of a one-dimensional flux model. The first and second models use the unsteady and steady state versions of the Pease [1987] model, respectively, while the third model positions the polynya edge at the point where the frazil ice depth exceeds a specified depth. No significant differences in the modeled mean polynya width were observed between the three models, suggesting that fluctuations in the extent of the Terra Nova Bay Polynya are usually in near equilibrium with the imposed forcing. It is also shown that taking account of the offshore decay of both wind speed and air

temperature results in a more realistic estimate of open water fraction.

[115] Haarpaintner *et al.* [2001] and Skogseth and Haugan (unpublished manuscript, 2003) study the Storfjorden polynya during the winter of 1997–1998. Once again, a one-dimensional Pease [1987] model is used to interpolate between polynya width measurements estimated from satellite images. The impact of surface ocean currents on the ice motion is neglected in both studies. However, Haarpaintner *et al.* [2001] and Skogseth and Haugan (unpublished manuscript, 2003) take into account the continuous ice growth that takes place below the consolidated ice-covered region when estimating the total growth and the associated brine production. The authors estimate that $\sim 40\text{ km}^3$ of ice is produced in the region over the winter, of which half originates in the open water region; this is sufficient to increase the salinity of the Storfjorden waters by 1.2 psu.

[116] Morales Maqueda and Willmott [2000] present a two-dimensional unsteady flux model of the opening of a wind-driven coastal polynya. The theory is an extension of previous one-dimensional unsteady and two-dimensional steady theories. Given surface winds, heat fluxes, and coastline geometry, the frazil ice depth distribution is determined, and the relative magnitudes of frazil ice and consolidated new ice fluxes normal to the polynya edge govern its evolution. The alongshore length scale L_a defined by Darby *et al.* [1995] is rigorously derived employing a perturbation method. The opening of a polynya adjacent to a prototype island, represented by a straight coastal barrier of finite length, is studied in detail by Morales Maqueda and Willmott [2000]. They found that if the length of the coastal barrier is comparable to L_a , then the timescale for the polynya to reach equilibrium can be very different from that obtained from a purely one-dimensional formulation. A simulation of the SLIP is also presented, using realistic coastal geometry (Figure 17). The model reproduces some of the salient features of the SLIP, in particular the splitting of the polynya into two subpolynyas during the winter months.

[117] The prescription of a constant value for the collection thickness of frazil ice at the polynya edge H , a practice adopted by all studies so far described, can lead to frazil ice depths at the polynya edge h_C exceeding H , which is unrealistic. Is there any a priori reason to expect H to be constant along the edge of a coastal polynya? The value of H partly governs the opening time and the area of a coastal polynya. In view of these facts it would be desirable to determine H from measurements. Routine field campaigns to coastal polynyas will not become a reality, and unfortunately, H is unlikely to be determined by remote sensing techniques. All these factors have motivated Biggs *et al.* [2000] to formulate a parameterization for H that does not contain any poorly determined variables. The Biggs *et al.* [2000] parameterization for H is based upon the work of Martin and Kauffman [1981] and Bauer and Martin [1983]. In a two-dimensional context the parameterization takes the form

$$H = h_C + c|(\mathbf{u} - \mathbf{U}) \cdot \mathbf{n}|^2, \quad (2)$$

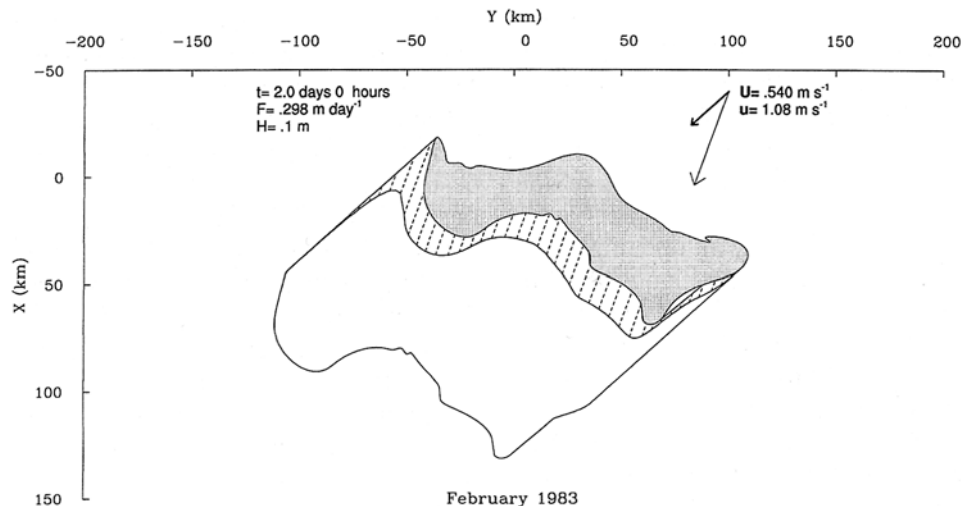


Figure 17. Simulation of the St. Lawrence Island Polynya on 16–18 February 1983, pictured 2 days after commencing opening from zero initial area. The polynya has reached 99% of its steady state area. The bold (thin) vector represents the consolidated new ice (frazil ice) velocity. Adapted from *Morales Maqueda and Willmott* [2000].

where \mathbf{u} and \mathbf{U} are the velocities of frazil ice and consolidated new ice, respectively, c is a constant determined from the densities of sea ice and water with the value $\sim 0.665 \text{ m}^{-1} \text{ s}^2$, and \mathbf{n} is a unit vector perpendicular to the polynya edge. H is thus coupled to the unknown polynya edge and must be determined as part of the problem. The flux equations incorporating the new parameterization for H are robust, as it is clear that $H > h_c$.

[118] The one-dimensional nonconstant H polynya edge evolves to its asymptotic width more slowly than the constant H edge. More striking discrepancies occur between the two classes of solution when the effects of varying the air temperature and offshore wind speed on the one-dimensional asymptotic width and opening time are examined. In contrast to the constant H polynya edge solution, which is a weak function of wind speed, the nonconstant H edge grows quadratically with increasing wind speed. As a consequence, the opening time increases linearly with growing wind speed, compared to a corresponding linear decrease in opening time for the constant H polynya edge solution.

[119] In light of the noted advantages of employing the nonconstant H parameterization, *Biggs and Willmott* [2001] incorporate it into the NOW model of *Darby et al.* [1994]. *Biggs and Willmott* [2001] address how sensitive the area of the NOW is to variations in wind stress orientation, equatorward oceanic volume transport through Nares Strait, and the frazil ice-water drag coefficient. In addition to the treatment of H , principal modifications to the *Darby et al.* [1994] model are the inclusion of a more realistic wind stress and frazil ice trajectories. Sensible heat is again shown to play little or no part in the maintenance of the NOW; while in order to reproduce the observed tongue of open water adjacent to the west Greenland coast, the new parameterization of H

is required. The polynya edge incorporating constant H does not do so.

[120] The parameterization of equation (2) is not robust in the unsteady problem because there are parameter regimes (e.g., frazil and consolidated ice drift orientations) for which the characteristic curves of the first-order partial differential equation governing the evolution of the polynya edge do not intersect the entire edge. To overcome this problem, the right-hand side of equation (2) could be modified by the addition of a small (typically 5 cm) constant depth term, associated with the contribution to frazil ice pileup at the polynya edge due to wave radiation stress [*Martin and Kauffman*, 1981]. This has been achieved by *Biggs and Willmott* [2004].

[121] Alternative empirical parameterizations for H as a function of wind velocity and polynya width [*Alam and Curry*, 1998] and wind velocity only [*Winsor and Björk*, 2000] have been proposed. However, there is currently discussion on whether the use of a constant H might be better after all given the uncertainties associated with wind and waves effects.

[122] Flux models can be adapted to describe the closure of a coastal polynya. Coastal polynya closure will occur when (1) ice is driven onshore or (2) in situ ice growth takes place within the polynya during a period of slack winds. Mechanism 1 is documented by *Winsor and Björk* [2000] for a polynya adjacent to the north coast of Svalbard, while mechanism 2 operates within the polynya system in the Weddell Sea during periods of slack winds [*Markus and Burns*, 1995].

[123] *Tear et al.* [2003] present one-dimensional flux models for the closure of a coastal polynya when the ice is drifting onshore. Two processes contribute to polynya closure: the onshore advection of pack ice and the pileup of frazil ice at the coastal boundary. For most parameter

regimes the closing timescale is shorter than the opening timescale. In these cases it is found that the heat released from the polynya to the atmosphere during opening exceeds that during closing. However, when the ice drift rate onshore (closing) is significantly smaller than the offshore (opening) drift rate, the closing time exceeds the opening time. Parameter regimes are also identified by *Tear et al.* [2003] for which the pileup of frazil ice at the coast plays a minor role in the closing process; in these cases the closing time is approximately equal to the advective time associated with the polynya edge moving onshore.

[124] *Biggs et al.* [2004] have extended the one-dimensional polynya closure model of *Tear et al.* [2003] to two dimensions and have examined the closure of a polynya formed in the lee of an idealized island consisting of a straight line barrier of finite length. For most parameter regimes (i.e., ice drift orientations and frazil ice production rates), polynya opening times are found to again exceed closing times. In addition, the closing time for the two-dimensional polynya is for all cases greater than the corresponding one-dimensional polynya closure time of *Tear et al.* [2003], because of persistent two-dimensional lead-like features which develop near the barrier ends.

5.2. Coupled Ice-Ocean and Ice-Atmosphere Models

[125] The adoption of coupled sea ice-ocean-atmosphere models for the simulation of polynyas has both advantages and disadvantages over flux theory. While it is clear that the inclusion of a dynamic-thermodynamic sea ice model (DTSIM) will better describe the consolidated ice pack, the ability of such models to represent frazil ice is at present severely limited. *Kantha* [1995] allows frazil ice to grow in the water column and immediately accrete at the underside of thicker ice above, but no provision is made for surface regions occupied solely by frazil ice. In a study of air-sea interaction over the Terra Nova Bay, *Gallée* [1997] includes a rudimentary parameterization of frazil ice based on the *Ou* [1988] model. *Gallée* [1997] suggests that the unexpectedly strong modeled heat losses from the polynya surface are a consequence of the idealized frazil ice model and concludes that a better representation of frazil ice is required.

[126] A more immediate advantage of DTSIMs is their more complete treatment of a number of physical processes. For example, consolidated new ice motion is specified a priori in flux models, while DTSIMs determine the motion of the entire sea ice field. Further, thermodynamical processes are described explicitly, and feedback between thermodynamics and dynamics is represented. Thermodynamical processes enter flux theory only via the ice production rate, which is often taken as uniform and usually neglected within the consolidated new ice zone. A further factor to be taken into account is the relative computational effort required for the two approaches. Flux theory is simple to use, and the computer time required for calculations is negligible. On the other hand, DTSIMs are complex to implement and computationally expensive. As a consequence of this difference in computational effort a theory based on polynya flux models offers

an attractive method of parameterization of polynyas within nonpolynya-resolving large-scale coupled ice-ocean models (see section 5.4).

[127] The work of *Bjornsson et al.* [2001] provides a connection between these two approaches. Here a DTSIM forced by a uniform wind stress is spun up to a steady state in a rectangular channel, closed at one end. *Bjornsson et al.* [2001] demonstrate that if the ice is driven against one of the channel side walls by the applied wind stress, a consolidated ice pack region forms downwind of a zone of strong ice velocity convergence. The consolidated ice region is characterized by a reduction in ice velocity and an increase in ice concentration and depth compared with the upstream frazil ice region. In these circumstances a clear polynya edge is obtained; that is, a region emerges in which there are large gradients in ice thickness and concentration. The position of the polynya edge so obtained is then compared to a polynya edge solution found from the flux model of *Darby et al.* [1995]. The flux model and DTSIM results show admirable agreement.

[128] Of particular note, however, are the results for a wind stress orientated along the channel in the direction of the open end. In this case, ice drift is along channel, and there are no regions of convergence. Consequently, no clear polynya edge is produced. This result has far reaching consequences for simulations of polynyas that feature no obvious regions of ice convergence, such as those formed in the lee of an island in the open ocean (e.g., the SLIP). Using the present generation of DTSIMs, contours of simulated ice thickness and concentration downwind of the island will not display the large gradients characteristic of a polynya edge.

[129] An opening of the SLIP in February 1992 is modeled by *Lynch et al.* [1997], who use a DTSIM coupled to a model of the atmosphere. In this study the issue of determining a clear polynya edge is not addressed, and open water is defined as regions covered by sea ice <30 cm thick. When the wind is strong and from the north, the simulated SLIP opening event shows good agreement with satellite data. The introduction of an air-ice drag coefficient dependent on the stability of the atmosphere provides a more accurate simulation. Further improvement is achieved by adding average surface currents to the forcing on the ice drift.

[130] An additional Arctic polynya, the NEW, is simulated by *Holland et al.* [1995]. Ice flowing south along the northeast Greenland coastline is squeezed and constricted by Fram Strait and forms an effective barrier to further southward flow. Doubts have been cast on previously proposed mechanisms for the NEW's formation. Upwelling does not appear to play a role, while the interaction of the Northeast Greenland Coastal Current (NEGC) with an ice barrier to the south, as hypothesized by *Schneider and Budéus* [1994], does not seem crucial, as no NEGC is simulated in the *Holland et al.* [1995] study. However, more recent observational studies of the NEW [*Minnett et al.*, 1997; *Schneider and Budéus*, 1997] suggest that the formation of the NEW in the *Holland et al.* [1995] study

was fortuitous and that the northward flowing NEGCC does, indeed, play an important role.

[131] The application of two- and three-dimensional numerical models, based on the primitive equations for ocean circulation, to the formation of dense water within a coastal polynya and its subsequent collapse and spread have been reviewed by *Baines and Condie* [1998]. These studies include those by Chapman and coworkers discussed in section 5.1, the simulations of small-scale convection and ice formation within polynyas by *Backhaus et al.* [1997] and *Kämpf and Backhaus* [1998], and an investigation of the Terra Nova Bay polynya by *Buffoni et al.* [2002]. The majority of the numerical experiments by *Backhaus et al.* [1997] used a two-dimensional (vertical slice) nonhydrostatic ocean model to study water mass formation due to episodic polar air outbursts over a polynya. However, they make no attempt to model the polynya area in response to the polar air outbursts but instead assume the scale of the polynya is large compared to the model domain. *Backhaus et al.* [1997] demonstrate that cold air outbursts contribute to the formation of dense shelf bottom water, provided the entire water column has been cooled to the freezing point. Ice patterns due to the interaction with oceanic convection are found to be similar in the model simulations to those observed in remotely sensed imagery [*Kämpf and Backhaus*, 1998]. *Buffoni et al.* [2002] also employ a vertical slice ocean model but prescribe a fixed polynya width. Simulated salinity and velocity distributions are, however, consistent with observations, and HSSW export is in reasonable agreement with the estimated 1 Sv of *Van Woert* [1999a].

[132] The work of *Gallée* [1997] contains a simulation of the Terra Nova Bay Polynya using an atmosphere-polynya model, the polynya portion of which is essentially that of *Ou* [1988]. Simple dynamics and thermodynamics are also included to describe the consolidated pack. The modeled heat fluxes are very strong (up to 1348 W m^{-2}), which is believed to be due to the idealized representation of frazil ice. The heat fluxes warm the katabatic airstream above the polynya, and an ice breeze is generated, which reinforces the wind coming off the ice sheet. Interestingly, a rapid decay of the offshore component of the atmospheric flow occurs at a distance commensurate with the Rossby radius of deformation, although as this far exceeds the polynya width, no effect is felt by the polynya itself.

[133] Notable among other studies of polynya evolution and maintenance that make use of DTSIMs are those addressing the formation and maintenance of the Weddell Polynya of the mid-1970s. The models used in these studies range from either regional or global stand-alone DTSIMs [*Parkinson*, 1983] through coupled ocean mixed layer models-DTSIMs-atmospheric boundary layer models [*Timmermann et al.*, 1999] to coupled DTSIMs-OGCMs [*Goosse and Fichefet*, 2001; *Marsland and Wolff*, 2001; *Holland*, 2001a, 2001b; *Beckmann et al.*, 2001]. To our knowledge, no investigation of the Weddell Polynya has yet been carried out with a DTSIM coupled to a primitive equation atmosphere-ocean model. Despite the large

number of modeling studies on the Weddell Polynya the ultimate causes of this extraordinary sea ice feature remain unclear.

[134] The formation mechanisms of many polynyas present in the Southern Ocean off the coast of Antarctica have both a latent and sensible heat component. The investigations of *Fichefet and Goosse* [1999] and *Goosse and Fichefet* [2001] of the Ross Sea spring polynya and of the Maud Rise polynya, respectively, are cases in point. The climatological monthly wind stresses imposed in both regions are such that a sea ice divergence is induced. In addition, an inflow of relatively warm water from the ACC melts the ice present and restricts further ice growth. The polynyas predicted by the model form at times and locations that are slightly incorrect, almost certainly as a consequence of the coarse model resolution.

[135] In contrast, the coastal polynya which opened in the summer of 1997–1998 in the southern Weddell Sea, and studied by *Hunke and Ackley* [2001], is a purely wind-driven feature, at least initially, due to an anomalous wind stress pattern for that period. Solar heating of the mixed layer (the open water-albedo feedback process) then plays a crucial role later in its maintenance. Upwelling from the deep ocean was shown by *Hunke and Ackley* [2001] to be insignificant in the polynya's maintenance.

5.3. Boundary Layer Models

[136] So far, models of the convective internal boundary layer (CIBL) over a polynya have not been coupled to a prognostic polynya model; polynya area is, instead, simply specified. Nevertheless, surface flux estimates from these models show good agreement with observations, so the CIBL approach offers a means to parameterize heat and moisture injection into the lower atmosphere.

[137] Models for the CIBL profile have been developed by *Lo* [1986], *Pinto et al.* [1995], and *Renfrew and King* [2000]. *Lo* [1986] includes heat and vapor equations, together with the momentum, mass, and turbulent energy ones. Numerical results display good agreement with experimental data from a polynya located off Dundas Island in the Canadian Arctic. The model of *Pinto et al.* [1995] is one-dimensional in the vertical but includes a detailed description of radiative transfer, a second-order turbulence closure scheme, and bulk cloud microphysics. A notable result is the exponential decay of CIBL height with time under stable conditions. The model of *Renfrew and King* [2000] is steady state and two-dimensional and assumes a balance between advection and turbulent heat flux convergence. Microphysical and radiative processes are ignored, but the omission only proves significant in a simulation containing cloud cover, in which discrepancies with observed data increase with fetch. In other case studies, modeled boundary layer heights and surface heat fluxes show good agreement with those observed.

5.4. Polynya Parameterization Within OGCMs

[138] The majority of polynyas have a horizontal extent far smaller than the resolution of the current generation of

OGCMs. Consequences of polynya processes are thus clearly not adequately reproduced in such models, and hence no proper account is taken of either the heat and moisture injected into the atmosphere above polynyas or of the substantial deep water production that occurs in the open water regions, both in the Arctic [Winsor and Björk, 2000] and in the marginal seas of Antarctica [Grumbine, 1991]. A procedure for parameterizing polynyas within large-scale coupled sea ice-OGCMs (which we assume for completeness to be coupled to some sort of atmosphere model) might be derived by extension of the work of Bjornsson *et al.* [2001]. Within each grid cell and at each time step it is first determined whether conditions are favorable for polynya formation, for example, for coastal polynyas whether the wind stress is oriented offshore, whether the air temperature is sufficiently low, or whether the upwelling and oceanic sensible heat flux are small enough. If these criteria are met, then a steady state polynya flux solution can be calculated from fields interpolated from the surrounding grid points. Van Woert [1999b] finds that simulations of the Terra Nova Bay Polynya with steady and unsteady one-dimensional flux models do not differ significantly, so it might reasonably be hoped that a steady state model would suffice. Frazil ice will be transported by the surface ocean currents that will need to be diagnosed. The “pileup” thickness of frazil ice at the polynya edge can either be prescribed or parameterized as given by Biggs *et al.* [2000]. The velocity of consolidated new ice is to be taken from the averaged OGCM-derived ice drift or from climatological data. From the results of the flux model, ice production, brine release, and the resulting surface buoyancy flux can be estimated [Winsor and Björk, 2000], as can the injection of heat and moisture into the lower atmosphere.

[139] Within our parameterization scheme, dense water production can be calculated from the results of the flux model following Chapman [1999]. Encouragingly, Chapman finds that the volume of dense water formed is a comparatively weak function of polynya width, so that the approximate polynya edge determined by the flux model may prove sufficiently accurate for reasonable dense water production estimates. A possible alternative to the approach of Chapman [1999] is the adoption of integrated axisymmetric plume models [Paluszkiwicz and Romea, 1997; Thorkildsen and Hagan, 1999], although these models might not be entirely relevant for shallow shelf convection.

[140] Clearly, the effect of polynyas in the lower atmosphere can be wide reaching, and a small number of these features can dominate winter regional budgets. These effects should be derived from the polynya flux model and fed back into the OGCM. A candidate for such a parameterization is the model of Glendening [1995] that describes the vertical distribution of the horizontally integrated heat flux into the atmosphere above an Arctic lead, based upon large-eddy simulations of its convective plume, and discusses a parameterization of this distribution in large-scale models. Similar models are described by Alam and Curry [1995] and Pinto *et al.* [1995]. Glendening [1995] demonstrates that except

under weak cross-lead wind regimes the horizontally integrated heat flux fits an exponentially decaying profile. It is estimated that the height at which 86% of the heat is expected to remain is typically smaller than the upper height of the lowest layer for coarse vertical-resolution climate models.

6. CODA

[141] It is likely that the dramatic increase in polynya studies that occurred during the 1990s will continue apace. Bearing in mind that fields such as surface temperature in the polar regions show pronounced sensitivity in climate change simulations, it is pertinent to address the following questions. First, how will climate change impact the frequency of polynya events, their area, the rate of dense water production, and the transfer of gases across the air-sea interface within these features? Second, will the diverse ecosystems supported within polynyas be sustained in future climatic regimes? A related question is whether present-day human inhabitants (e.g., the Inuit) who rely on the ecosystems supported by polynyas can be sustained in future climatic regimes.

[142] Underpinning research aimed at answering the above questions is the requirement for long-term monitoring of key variables in “key polynyas.” From a logistical point of view it is almost certain that a monitoring program of this type would focus on the Arctic. The NOW is the best documented of all Arctic polynyas, and for this reason alone it should continue to be observed. There is relatively little data from the polynyas over the Eurasian continental shelf. Yet polynyas in these seas are important sites for dense water production. Thus we would advance the idea that a major polynya in one of these seas should become the focus of a long-term monitoring program.

[143] A concentrated effort should be undertaken to address whether the omission of polynya processes in nonpolynya-resolving climate models really matters. Specifically, how important is it to model dense water production and the ocean-to-atmosphere heat flux within polynyas in order to be able to realistically simulate the global thermohaline circulation and atmospheric circulation and hydrological cycle? Polynya flux models provide a promising method of parameterizing the aforementioned quantities in nonpolynya-resolving OGCMs/climate models. However, a drawback with all flux models in the literature is that they do not include an ice concentration field. Instead, flux models characterize frazil and consolidated ice types by their depth distributions. In practice, polynyas are identified as regions of low ice concentration, and therefore caution is required when comparing flux model polynya results with observations. Further research on parameterizations for the collection depth of frazil ice is also urgently required.

[144] Many deficiencies exist in the way sea ice is modeled. One such deficiency concerns the lack of feedback between ice motion, brine input into the ocean, and

ocean pressure gradients. Brine fluxes modify the cross-shelf pressure gradient, which, in turn, modifies the shelf circulation and hence the ice motion. To address this problem rigorously, attention must be given to the way in which brine drains from sea ice. When sea ice forms, the salt-enriched water remains within its crystalline structure until the layer of ice exceeds a certain thickness [Wetlaufer et al., 1997, 2000]. Subsequently, salt-enriched water drains out, and this means that the buoyancy forcing of the shelf waters is highly episodic.

[145] In the Antarctic a number of interesting polynya problems remain to be solved. For example, will the major ice shelves remain intact in a warmer climatic regime or not? This, in turn, requires knowledge of the under-ice shelf circulation in changing climatic regimes, and this problem is related to the existence of Antarctic coastal polynyas. Take the Ronne-Filchner Ice Shelf System as an example. A fraction of the dense water formed in the polynya system adjacent to the ice shelf front sinks beneath the ice shelves. Thus processes within this polynya system drive the under-ice shelf circulation. During the next decade or so, combined observational and modeling programs (e.g., the U.K. Autosub Under Ice programme [Clarke, 2003]) will address the problem of the stability of the Antarctic ice shelves.

[146] Another unsolved problem is determining the processes controlling the formation and quasi-steady extent of deep water polynyas (e.g., Weddell and Cosmonaut Polynyas). We note from sections 2, 3, and 5 that many aspects of this problem have been addressed in process studies. What is lacking is a process model for deep water polynyas capturing the key physics operating in a coupled atmosphere-ocean-sea ice system. Indeed, it may well be that to simulate the evolution of any “large” polynya a three-component atmosphere-ocean-sea ice model will be required. We are confident that during the next decade further pieces in the jigsaw puzzle of polynya processes will fall into place.

[147] **ACKNOWLEDGMENTS.** The comments of M. Van Woert, T. Torgersen, D. M. Holland, P. M. Haugan, and an anonymous reviewer are gratefully acknowledged. NRTB acknowledges support of the NERC via a research grant awarded to AJW. MAMM is supported by the German Ocean/CLIVAR Research Program.

[148] Thomas Torgersen is the Editor responsible for this paper. He thanks three technical reviewers and one cross-disciplinary reviewer.

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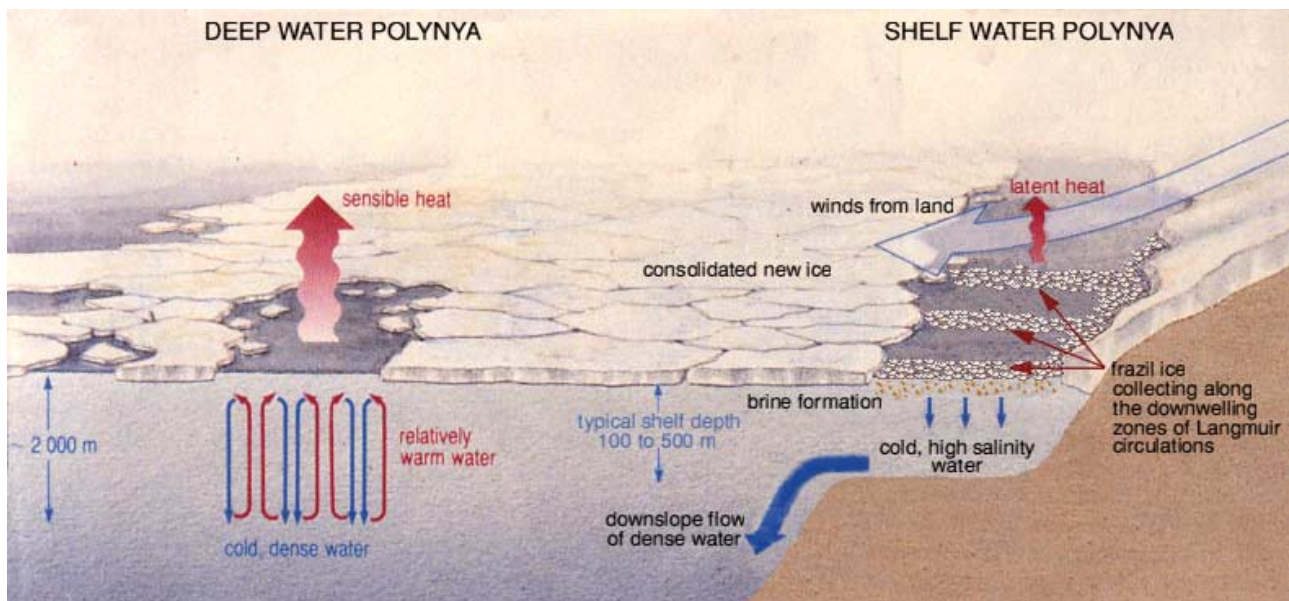


Figure 1. Schematic representation of physical processes taking place in deep water and shelf water polynyas. Deep water polynyas occur beyond the continental shelf break and are frequently created and maintained by a sensible heat mechanism, whereby upwelling or vertical mixing of subsurface water brings about sea ice melting or prevents ice from forming. Strong surface buoyancy losses through cooling can lead to deep water formation. Shelf water polynyas appear over the relatively shallow continental shelves. The majority of these polynyas are wind-driven. Offshore winds push the pack ice away from the coast, exposing the freezing surface waters to the cold atmosphere. Frazil ice is thus formed and herded downwind along Langmuir windrows. Brine-rich, cold water associated with sea ice formation will accumulate over the shelf and eventually flow down the shelf break slope to form deep and bottom water. Redrawn from *Open University Oceanography Course Team* [2001], with permission from Elsevier.

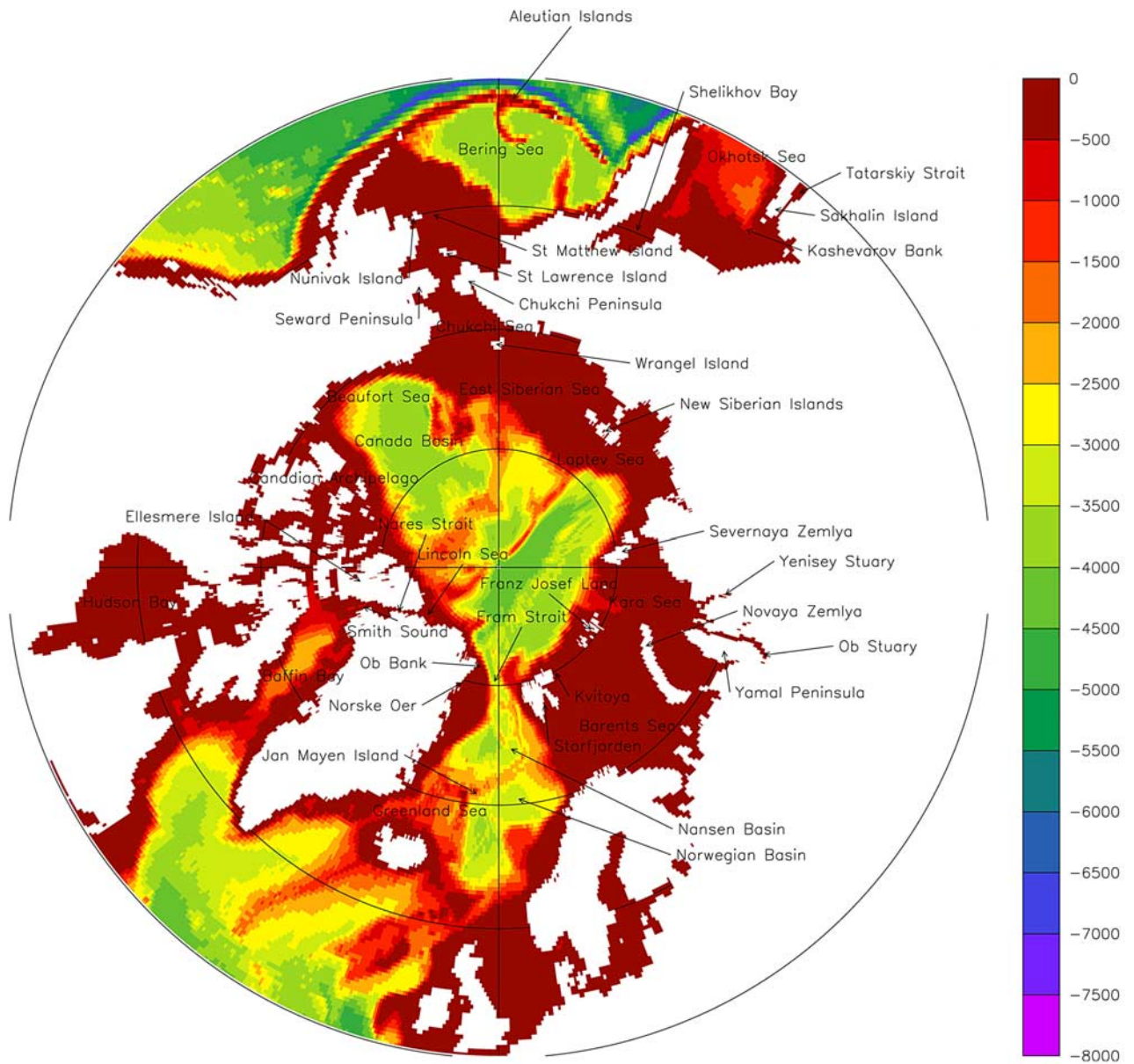


Figure 2. Northern Hemisphere geographical locations referred to in the text, together with bathymetry (in meters).

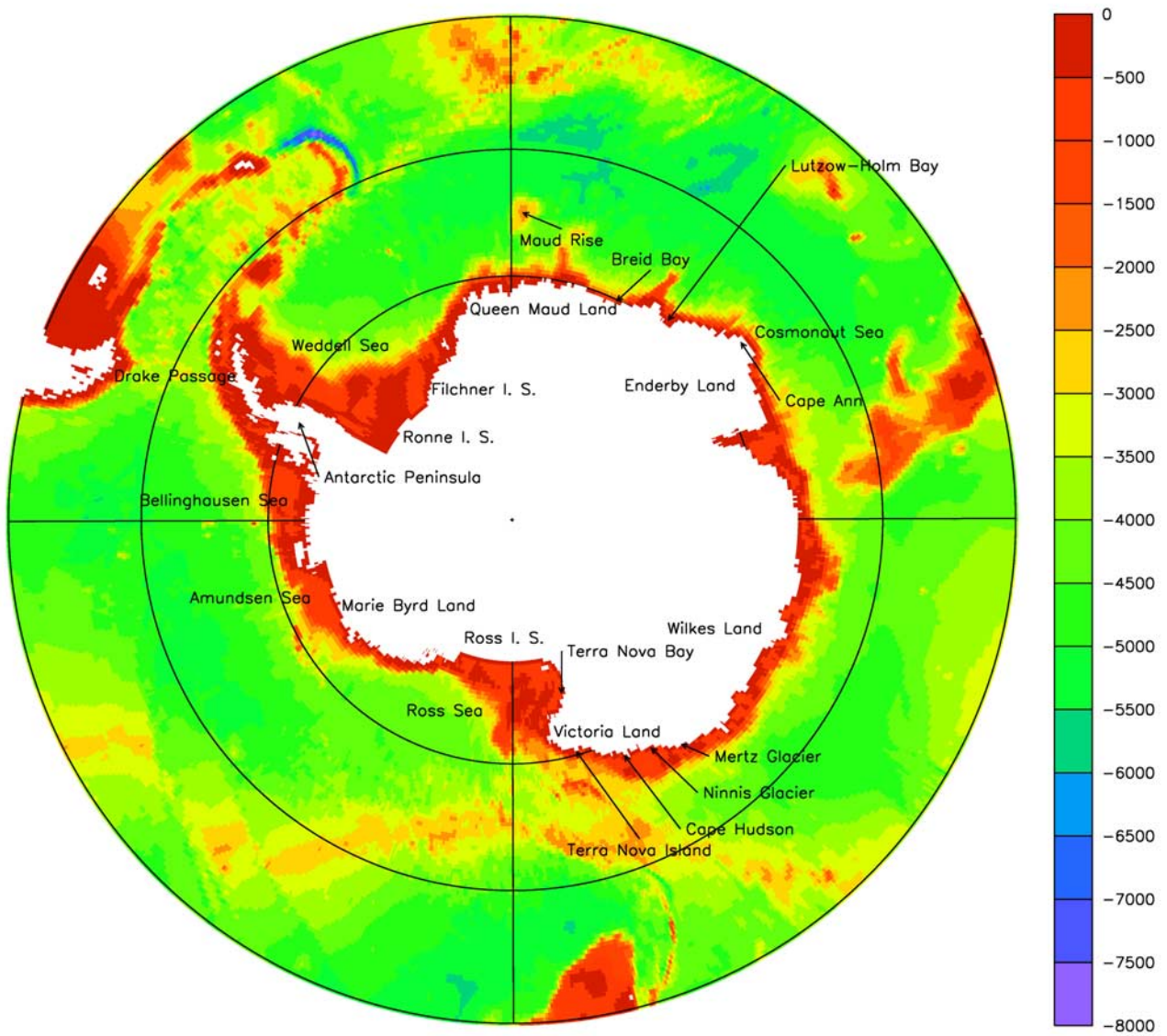


Figure 3. Southern Hemisphere geographical locations referred to in the text, together with bathymetry (in meters). Adélie coast (not shown) is the coastal area west of Victoria Land.