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Arctic Sea Ice Extent and Thickness [and Discussion]

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Arctic sea ice extent and thickness

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Current knowledge on Arctic sea ice extent and thickness variability is reviewed, and we examine whether measurements to date provide evidence for the impact of climate change. The total Arctic ice extent has shown a small but significant reduction of $(2.1 \pm 0.9)\%$ during the period 1978–87, after apparently increasing from a lower level in the early 1970s. However, open water within the pack ice limit has also diminished, so that the reduction of sea ice area is only $(1.8 \pm 1.2)\%$. This stability conceals large interannual variations and trends in individual regions of the Arctic Ocean and sub-Arctic seas, which are out of phase with one another and so have little net impact on the overall hemispheric ice extent. The maximum annual global extent (occurring during the Antarctic winter) shows a more significant decrease of 5% during 1972–87. Ice thickness distribution has been measured by submarine sonar profiling, moored upward sonars, airborne laser profilometry, airborne electromagnetic techniques and drilling. Promising new techniques include: sonar mounted on an AUV or neutrally buoyant float; acoustic tomography or thermometry; and inference from a combination of microwave sensors. In relation to climate change, the most useful measurement has been repeated submarine sonar profiling under identical parts of the Arctic, which offers some evidence of a decline in mean ice thickness in the 1980s compared to the 1970s. The link between mean ice thickness and climatic warming is complex because of the effects of dynamics and deformation. Only fast ice responds primarily to air temperature changes and one can predict thinning of fast ice and extension of the open water season in fast ice areas. Another region of increasingly mild ice conditions is the central Greenland Sea where winter thermohaline convection is triggered by cyclic growth and melt of local young ice. In recent years convection to the bottom has slowed or ceased, possibly related to moderation of ice conditions.

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1. Introduction

In this review we discuss evidence for climate-related changes in Arctic sea ice extent and thickness, and for ice-related changes in convection in the Greenland Sea.

The most important property of Arctic sea ice which affects its response to climate change is the fact that it is in motion, driven mainly by wind stress. Ice opens up under divergent stress, creating leads which rapidly refreeze in winter. Subsequent convergence or shear causes the leads to be crushed to create pressure ridges, semi-permanent features of the ice cover which contain about half of the total Arctic ice volume (Wadhams 1981). The wide distribution of thickness produced by mechanical processes results in ocean-atmosphere heat and moisture fluxes which are highly time- and space-dependent. A simple warming will not necessarily cause the ice cover as a

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whole to become either thinner or less in extent; one has to consider several feedbacks associated with ice deformation and thickness variations.

Palaeoclimatic evidence suggests that some 18 000 years BP, during the most recent glaciation, sea ice extended down to the latitude of Spain and New England (Thiede, this volume). Thus, in contrast to the stable Antarctic Ice Sheet, sea ice does appear to be relatively unstable and responsive to climate change.

In § 2 we consider evidence for sea ice extent variability, then in § 3 we deal with the thickness distribution and how it varies. In § 4 we consider how changes in Greenland Sea ice may be related to a change in the rate of winter convection, and in § 5 we consider the mechanisms by which the extent and thickness of pack ice may respond to climate change.

2. Sea ice extent

(a) *Importance in global change*

Sea ice extent is defined as the ocean surface area enclosed by the ice edge. Within that boundary ice concentration may be less than unity and locally may even be zero, as in the case of persistent winter or spring polynyas such as the Northeast Water off northeastern Greenland or the North Water in Smith Sound. Therefore ice area is less than ice extent.

Sea ice area is related to global change by the same albedo feedback loop as snow cover on land. The albedo of sea ice ranges from 0.9 in winter after fresh snowfall, to 0.4–0.55 in summer when the surface is partly covered by meltwater pools. Since the albedo of open water is about 0.10, a decline in ice area will allow the absorption of more incoming short-wave radiation by the ocean and hence will feed back positively on surface temperature.

A further climatically important effect related to sea ice concentration is that most ocean-atmosphere heat loss occurs through open water or thin ice (Maykut 1986). In the case of open water the effect is not linear, in that when the concentration drops below $\frac{4}{10}$ the heat flux is almost as high as in open water (Worby & Allison 1991). Further, in summer, melting is enhanced in an open icefield because of the large floe perimeter per unit area.

(b) *Measurement*

The high contrast between ice and open water in both the visible and microwave spectral bands means that the advent of satellite observations permitted ice extent to be mapped on a worldwide basis. The first studies used visible or thermal infra-red imagery from radiometers on weather satellites, such as the AVHRR (advanced very high-resolution radiometer) on the NOAA satellites. Estimates of Arctic sea ice extent on an annually averaged basis from 1973 to 1990, for instance, were given in IPCC (1992, figure C13). The inability of such radiometers to see through cloud is a cause of inadequacy in the datasets, extending also to published ice charts which are usually constructed from a range of satellite data. It is interesting, however, that the general trend of ice extent in the 1992 IPCC chart (a Northern Hemisphere recovery in 1975–76 from an initially low value, followed by a slow subsequent decline) was reproduced successfully in predictions of total ice volume in the Arctic generated by Flato (1995) using a model which used daily varying geostrophic winds and monthly varying surface air temperatures to drive the ice.

A more effective means of measuring sea ice area and extent is by the use of

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satellite-borne passive microwave radiometers. These give all-weather day-and-night coverage of the polar regions apart from a small zone around the Poles. Algorithms relate microwave emissivity to ice concentration for single-frequency sensors, and yield additionally the fraction of older (multi-year) ice present in the case of multi-frequency sensors. Passive microwave data are available from 1973 to 1976 from the single-frequency ESMR (electrically scanning microwave radiometer), replaced from 1978 by the SMMR (scanning multichannel microwave radiometer) and in 1987 by the SSM/I (special sensor microwave/imager). Results from ESMR and SMMR have been summarized in atlases (Parkinson *et al.* 1987; Gloersen *et al.* 1992), which also discuss the method of data reduction and ice parameter extraction.

(*c*) *Evidence for changes in extent*

(i) *Hemispheric changes*

Ice extent data from these satellites demonstrated regional positive and negative trends in extent (Parkinson & Cavalieri 1989), with a slight negative overall trend for 1979–86 as compared to a slight positive trend for 1973–6. In a further analysis of the same data, Gloersen & Campbell (1991) corrected for instrumental drift and orbital errors, chose a 15% isopleth as the ice edge position, and used band-limited regression as the most appropriate statistical technique to search for long-term trends in a seasonally cyclic dataset. Their final result was a small but statistically significant reduction of $(2.1 \pm 0.9)\%$ in Arctic ice extent during the 8.8 years of SMMR (1978–87), with a confidence level of 96.5%, and a decrease in open water area within the pack limits of $(3.5 \pm 2.0)\%$ over the same period (confidence level 93.5%). The actual area of Arctic ice cover thus showed a less significant negative trend of $(1.8 \pm 1.2)\%$ with a significance of 88.5% (figure 1). No significant trend occurred in the Antarctic.

Present efforts are devoted to checking this relationship by extending the dataset backwards in time to include ESMR and forward to include SSM/I, but there are difficulties in achieving compatibility between algorithm products from different sensors.

In an earlier analysis (1988) Gloersen and Campbell had found that if Arctic and Antarctic sea ice extent are combined, the resulting global curve shows a significant negative trend in its peak value, amounting to 5% in 13 years. For this study the data were uncorrected and so the trend may be an over-estimate.

Even when ESMR and SSM/I data to the present (1994) can be included, the maximum duration of passive microwave time series will only be 22 years. As with all climatological datasets of limited duration, there is the possibility that an apparent trend detected over this time interval is actually part of a longer-term cycle.

(ii) *Regional changes*

A number of studies on regional changes in the Arctic have shown trends of decadal length (Walsh & Johnson 1979; Smirnov 1980; Mysak & Manak 1989; Parkinson & Cavalieri 1989). The SMMR atlas (Gloersen *et al.* 1992) divided the Arctic into nine regions, each of which was analysed separately for ice extent; ice area; open water area within the pack; and, in some cases during winter, multi-year fraction. Each region showed large interannual variations in each of these quantities, but with some out-of-phase effects between regions. The trends involved were analysed quantitatively by Parkinson and Cavalieri (1989) and again by Gloersen *et al.* (1992, p. 205). They found negative trends in the Sea of Okhotsk ($15\,000\ \mathrm{km^2\ a^{-1}}$), Kara/Barents Seas ($28\,000\ \mathrm{km^2\ a^{-1}}$) and Greenland Sea ($5000\ \mathrm{km^2\ a^{-1}}$); and generally smaller positive

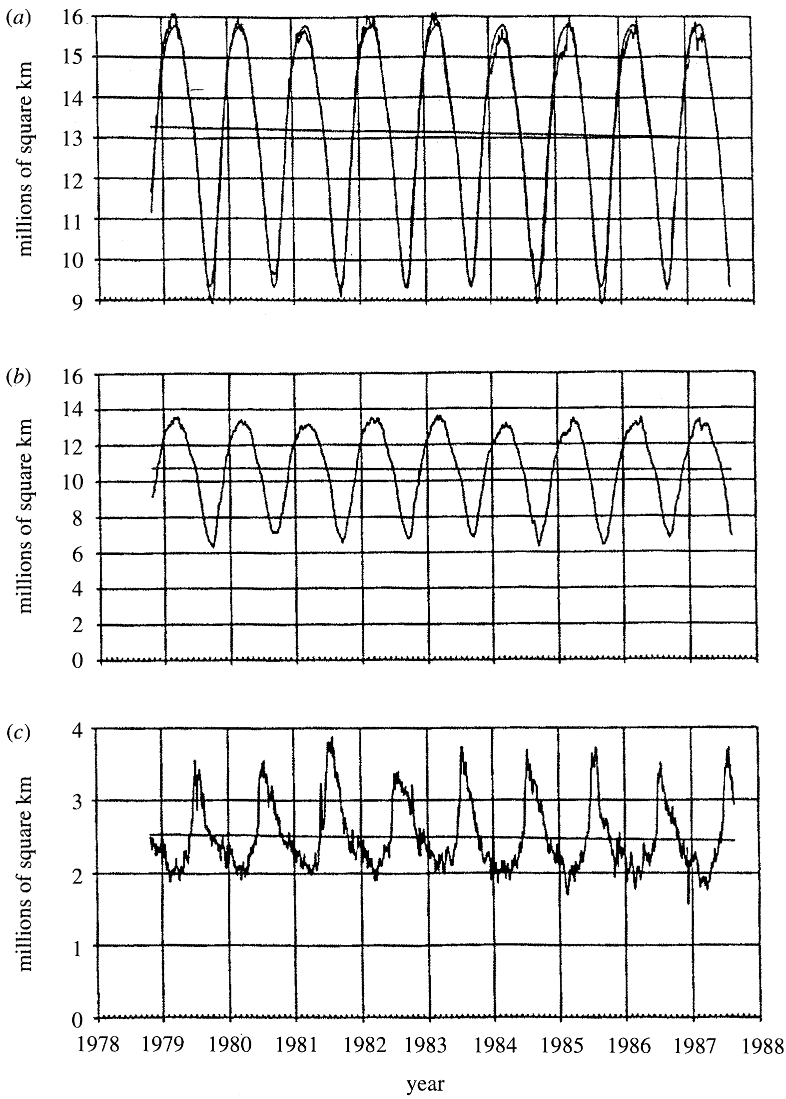

Figure 1. An 8.8-year cycle of Arctic sea ice (*a*) extent and (*b*) area, and (*c*) the open water area within the pack, together with trend lines. (After Gloersen *et al.* 1992.)

trends in Baffin Bay/Davis Strait (19 000 km^2 a^{-1}), Hudson Bay, the Bering Sea, the Gulf of St. Lawrence and the central Arctic Basin (all 4–6000 km^2 a^{-1}), the overall trend being a slightly negative one of 8000 km^2 a^{-1}.

Attempts have been made to associate periodicities in ice severity with periodic phenomena occurring in other parts of the world. Ono (1993) reported an apparent correlation between ice area in the Baffin Bay – Labrador Sea region as analysed by Mysak & Manak (1989), and indices of severity of the El Niño Southern Oscillation (ENSO). Figure 2 shows the Niño-3 sea surface temperature anomaly (anomaly in mean SST in the quadrangle 4° N, 4° S, 150° W, 90° W, a positive value indicating an El Niño year); the southern oscillation index (difference in sea level atmospheric pressure between Tahiti and Darwin, a negative value indicating atmospheric forcing linked to El Niño) and the ice area anomaly. There is an apparent correlation

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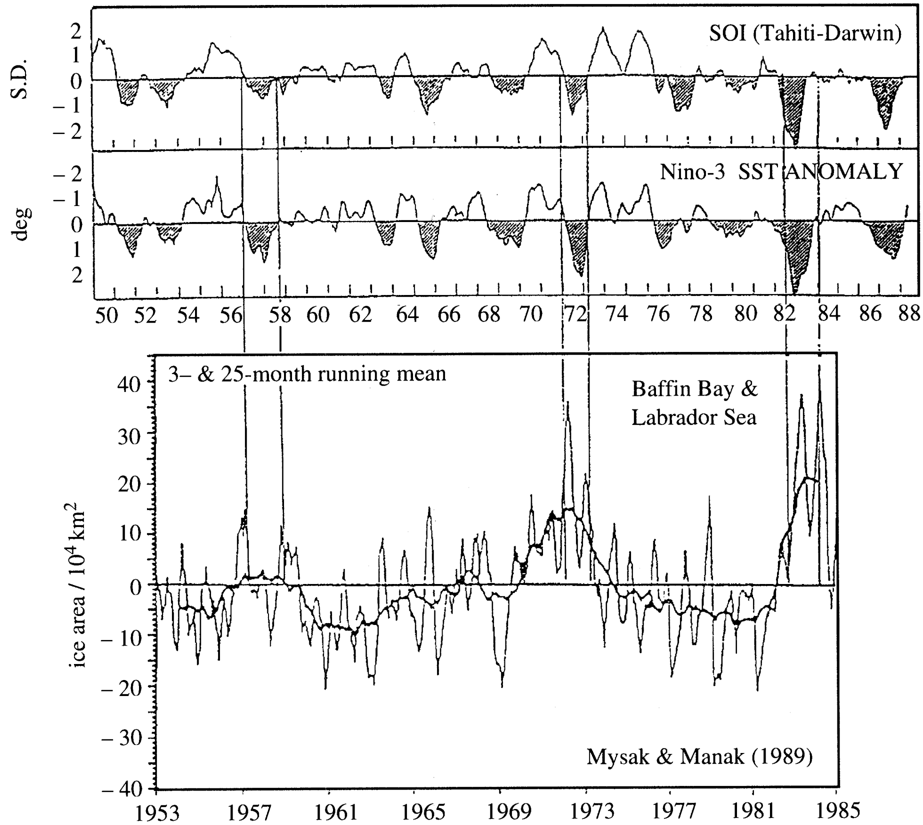

Figure 2. Ice extent anomaly in the Baffin Bay–Labrador Sea sector of the Arctic, compared with two indicators of the ENSO cycle. (After Ono 1993.)

involving three major El Niño events, but a causal link between tropical Pacific and northwestern Atlantic processes remains obscure. A more direct link might be expected with Bering Sea ice extent, and indeed if we compare figure 2 with fig. 3.4.2 of Gloersen *et al.* (1992) we find that El Niño events of 1982–3 and 1987 correspond with anomalously low winter ice extents in the Bering Sea. It is premature to comment on whether these correlations indicate association.

(iii) *Changes in the length of the ice season*

Using the 1979–86 SMMR dataset, Parkinson (1992) studied the length of the ice season, defined as the number of days in a calendar year during which a given grid element was covered by ice of concentration at least 30%. A least squares fit to a linear trend was obtained for each grid element. The results showed a consistent, spatially coherent pattern. During the 8-year period the season became significantly shorter off the north coast of Russia, and in the Greenland Sea, the Barents Sea and the Sea of Okhotsk, by often as much as 8 d a^{-1}, and significantly longer in the Gulf of St Lawrence, Labrador Sea, Hudson Bay, Beaufort Sea and eastern Bering Sea, by up to 16 d a^{-1}. This reinforced conclusions reached in the regional trend analysis, and demonstrated how these trends actually represent a shift in ice climatology during the SMMR period towards greater severity in the North American Arctic and greater mildness on the Eurasian side.

(d) Summary

The total Arctic ice extent has shown a small but statistically significant reduction of $(2.1 \pm 0.9)\%$ during the period 1978–87. Open water within the pack ice limits has also diminished, so that the reduction of ice area is only $(1.8 \pm 1.2)\%$. The underlying stability conceals large interannual variations and possible longer period trends in individual regions, some of which are completely isolated from an Arctic Ocean source (e.g. Sea of Okhotsk, Gulf of St. Lawrence). These trends have sufficient out-of-phase components with one another to result in only a minor overall trend in the hemispheric ice extent. When Arctic and Antarctic data are combined, the maximum annual global extent (occurring during the Antarctic winter) shows a more significant decrease of 5% during the same period.

3. Sea ice thickness

(a) The importance of ice thickness distribution

In considering ice thickness, it is important to deal with the probability density function $g(h)$ rather than just the mean thickness, for the following reasons:

(i) $g(h)$ determines the ocean-atmosphere heat exchange, with thin ice dominating;

(ii) together with the ice velocity, it gives mass flux;

(iii) its downstream evolution gives the melt rate, i.e. the fresh water flux;

(iv) its shape is a measure of the degree of deformation of the ice cover;

(v) if multi-year fraction is also known, $g(h)$ can be used to estimate ice strength and other statistically definable mechanical properties of the ice cover;

(vi) its variability is a test of model outputs;

(vii) its long term trend indicates the climatic response; and

(viii) in the central Greenland Sea the thickness achieved by young ice is associated with the magnitude of winter convection.

In addition to $g(h)$ it is also valuable to measure the ice bottom shape, requiring under-ice profiling rather than simply sampling the thickness at fixed intervals as is done with moored sonar. The advantages of this procedure are:

(i) shape is a determining factor for the aerodynamic and hydrodynamic drag coefficients;

(ii) the deepest pressure ridges are responsible for generating internal waves which may cause a significant internal wave drag;

(iii) seabed scour by the deepest ridges defines the fast ice limit on shelves and the extent of the stamukhi zone (Reimnitz *et al.* 1994);

(iv) ridges are an important component in the force exerted by an icefield on offshore structures;

(v) the scattering of underwater sound by ridges defines the range to which acoustic transmission can be accomplished, since upward refraction leads to repeated surface reflection; and

(vi) ridged ice provides additional habitats for sea ice biota.

The ice thickness distribution evolves with time through three processes: thermodynamic growth (or decay) which causes thin ice to grow thicker and thick ice to ablate; divergence of the ice cover, a source of open water and a sink of ice-covered area; and ridge formation, which moves ice from thin to thick categories in a way which is parameterized in models by a so-called ice thickness redistributor.

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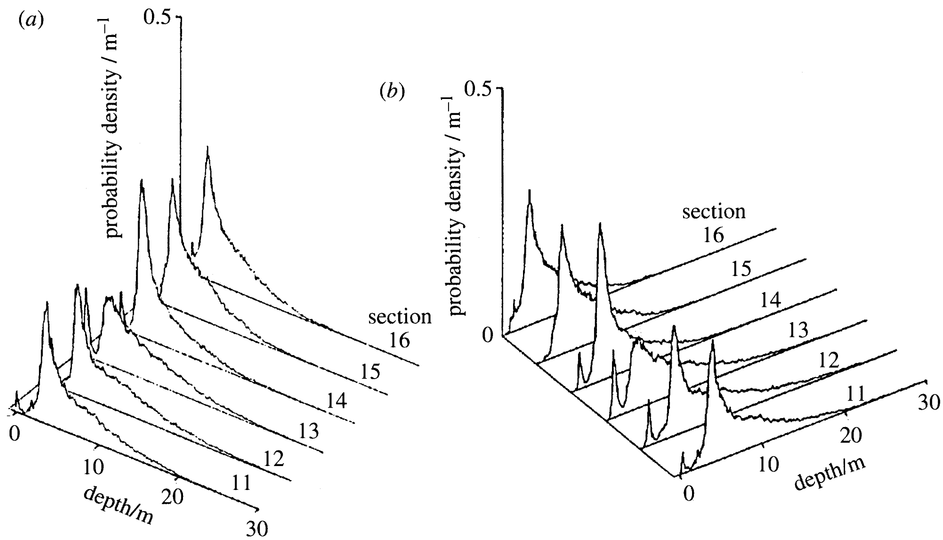

Figure 3. Some typical probability density functions of ice draft derived from successive 100 km submarine sonar profiles in the Arctic Basin at 85–90° N, 70° W. The two projections (*a*) and (*b*) emphasize the thick and thin ice respectively. (After Wadhams 1981.)

A typical ice thickness distribution (figure 3) has the following features: one or more peaks in the range 0–1.5 m due to ice in refrozen lead systems of different ages; a peak at about 2 m and another at about 3 m (sometimes the two are merged) representing undeformed first- and multi-year ice; and a tail, representing ridged ice, which shows an excellent fit to a negative exponential distribution (Wadhams 1981). If one views ridge-building as a Gibbs process, in which the work done translates into the potential energy acquired by the deformed ice, then a negative exponential form is the expected shape for this part of $g(h)$ (B. Kerman 1994, personal communication). The thickness distribution has been approximated by forms such as a gamma function (Goff 1995) or a sum of lognormals (Hughes 1991) for stochastic simulation purposes.

The ice surface topography has other statistical characteristics whose variability may have climatic implications. Pressure ridge keel depths and sail heights are observed to fit a negative exponential distribution (Wadhams 1981); pressure ridge spacings fit a log-normal (Wadhams & Davy 1986); and lead widths fit a power law of exponent about −1.45 for leads less than 100 m wide and −2.50 for wider leads (Wadhams 1992).

(*b*) *How ice thickness is measured*

(i) *Present techniques*

Methods of measuring ice thickness in the Arctic have been discussed by Wadhams (1994*a*) and Wadhams & Comiso (1992). Five methods are in common use. In decreasing order of total data quantity they are: submarine sonar profiling; moored upward sonar; airborne laser profilometry; airborne electromagnetic techniques; and drilling.

Submarine sonar can obtain synoptic data on thickness and under-ice topography rapidly and accurately, and most information on the distribution of $g(h)$ over the Arctic comes from upward-looking submarine sonar profiles. The addition of sidescan or swath sounding sonar adds information on ice type and two-dimensional bottom

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topography (Wadhams 1988). However, repeated profiling over identical tracks to test for climate-related trends, or systematic grid surveys, are both difficult with military submarines since ice profiling is an addendum to their operational tasks.

Moored upward sonar gives long-term information from a single point. It is invaluable for assessing the time variation of ice flux through critical regions such as Fram Strait. However, the cost and difficulty of deployment and recovery preclude its general use on a systematic measurement grid over the central Arctic Basin.

Airborne laser profilometry yields a freeboard distribution which can be converted to draft distribution if the mean density of ice plus overlying snow is known (Wadhams *et al.* 1991). Seasonal and regional validation of snow depths is needed before this otherwise rapid and efficient technique can be used for basinwide surveys.

Airborne electromagnetic techniques involve generating and sensing eddy currents under ice by VLF (10–50 kHz) EM induction from a coil towed behind a helicopter, with a simultaneous laser to give range to the snow surface. The technique has been reviewed by Rossiter and Holladay (1994). The wide footprint involves loss of resolution of individual ridges and a need to fly very low. Recently a system has been mounted in a Twin Otter by the Geological Survey of Finland, giving greater range.

Drilling is the most accurate, but slowest, technique, the ultimate validation for all others.

(ii) *Future techniques*

Synoptic, repeated ice thickness surveys over the Arctic Basin are essential for detecting a climatic signal in ice thickness. Novel techniques may be needed, such as:

(i) Sonar surveys by a dedicated submarine, through temporary reassignment from the military, full conversion to a civilian research vessel (the so-called 'white submarine' concept), or by use of a commercially designed small submarine.

(ii) Mounting sonars on an unmanned vehicle. For short-range surveys this could be a cable-controlled ROV (remotely operated vehicle), and for mesoscale and basinwide surveys an AUV (autonomous underwater vehicle). Vehicles with 300 km range are in service (Tonge 1992), and vehicles of longer range are under development.

(iii) Mounting sonar on a neutrally buoyant float, to construct $g(h)$ over, say, a week's drift, the data being transmitted acoustically to a readout station. This requires only a modest extension of the existing technology of under-ice SOFAR and RAFOS floats.

(iv) The use of acoustic techniques. It has been shown that travel time changes for an acoustic path are reduced by the presence of an ice cover, by an amount approximately proportional to the ice thickness (Guoliang & Wadhams 1989; Jin *et al.* 1993). In long range acoustic propagation experiments this can be used to give a single mean value for ice thickness along a path. In spring 1994 the first transmission across the Arctic Basin from Svalbard to the Beaufort Sea was successfully accomplished.

(v) Increased efforts to obtain empirical correlations between ice thickness and the output of satellite sensors such as passive and active microwave or altimeter. Already a positive correlation between SAR backscatter and ice thickness has been demonstrated (Wadhams & Comiso 1992). Further advance requires extensive validation.

(vi) Deriving ice thickness as a by-product of another measurement. For instance, long distance swell propagation in ice is subject to a slow attenuation due to creep (Wadhams 1973). The attenuation rate is frequency- and thickness-dependent. In principle one could obtain spectra of flexure from sets of strainmeters or tiltmeters across the Arctic, and derive a mean value for ice thickness from the attenuation rate.

An Arctic mapping strategy might be to combine repeated under-ice sonar profiles over a grid covering the Basin, with moored upward sonar measurements spanning key choke points for ice transport, i.e. Fram Strait, the Svalbard–Franz Josef Land gap and a small number of specimen points within the Trans Polar Drift Stream and Beaufort Gyre. Sonar moorings could be combined with current meters and sediment traps. If submarines are not available for the task, an alternative would be a combination of long-range AUVs, airborne laser surveys and (for regional measurements) airborne EM. At the same time, the validation of satellite sensors can be developed. Ice thickness mapping is being addressed by the Arctic Climate System Study (ACSYS) of WCRP (WCRP 1994).

(*c*) *Present knowledge of Arctic ice thickness*

Submarine sonar profiling shows that over the Arctic Basin there is a gradation in mean ice thickness from the Russian Arctic, across the Pole and towards the coasts of north Greenland and the Canadian Arctic Archipelago, where the highest mean thicknesses of some 7–8 m are observed (LeSchack 1980; Wadhams 1981, 1992; Bourke & McLaren 1992). These overall variations are in accord with the predictions of numerical models (Hibler 1979, 1980) which take account of ice dynamics and deformation as well as thermodynamics. Large scale maps of the general distribution of mean thickness have been generated by LeSchack (1980), Bourke & Garrett (1987) (figure 4) and Bourke & McLaren (1992). The data used for the Bourke and Garrett map did not include open water, so these maps are over-estimates of the mean ice draft. The update by Bourke & McLaren (1992) included contour maps of standard deviation of draft, and mean pressure ridge frequencies and drafts for summer and winter, based on 12 submarine cruises.

Measurements in sub-Arctic seas show that the ice in Baffin Bay is largely thin first-year ice with a modal thickness of 0.5–1.5 m (Wadhams *et al.* 1985). In the southern Greenland Sea the ice, although composed largely of partly melted multi-year ice, also has a modal thickness of about 1 m (Vinje 1989; Wadhams 1992), with the decay rate in thickness from Fram Strait giving a measure of the fresh water input to the Greenland Sea at different latitudes.

(*d*) *Evidence for changes*

In order to understand whether, and how, the thickness of sea ice in the Arctic is responding to climate change it is necessary to measure $g(h)$ repetitively on a synoptic scale, preferably using the same equipment. Data available at present offer only indications that climate-related change may be occurring.

McLaren (1989) compared data from two US submarine transects of the Arctic Ocean in August 1958 and August 1970, running from Bering Strait to Fram Strait via the North Pole. He found no major changes in the Eurasian Basin and North Pole area, but significantly milder conditions in the Canada Basin in 1970. The difference is possibly due to anomalous cyclonic activity as observed in the region in some summers (Serreze *et al.* 1989). Also, since August is the month of greatest ice retreat in the Beaufort Sea, the difference may be due to differences between the ice edge positions in the Chukchi and southern Beaufort Seas during the respective summers.

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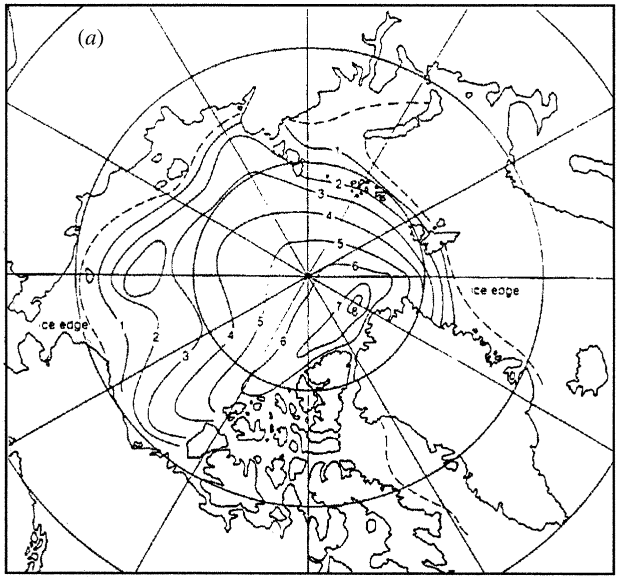

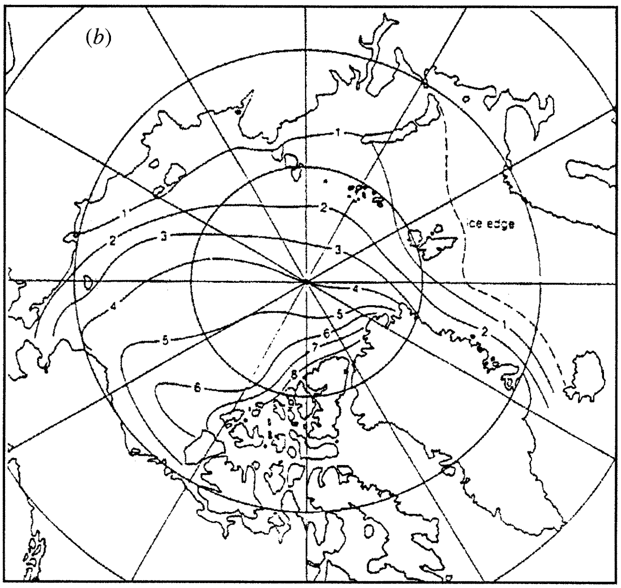

Figure 4. Contour maps of estimated climatology of mean ice thickness for (*a*) summer (*b*) winter in the Arctic Basin, neglecting open water. (After Bourke & Garrett 1987.)

An unusually open southern Beaufort Sea would lead to more open conditions within the pack itself.

Wadhams (1989) compared mean ice drafts for a region of the Eurasian Basin north of Fram Strait, from British cruises carried out in October 1976, April–May 1979 and June–July 1985, all using similar sonar equipment. A box extending from 83°30′ N to 84°30′ N and from 0° to 10° E had an especially high track density from the three cruises (400 km in 1976, 400 km in 1979 and 1800 km in 1985). It is distant from land boundaries, and is representive of the Trans Polar Drift Stream prior to

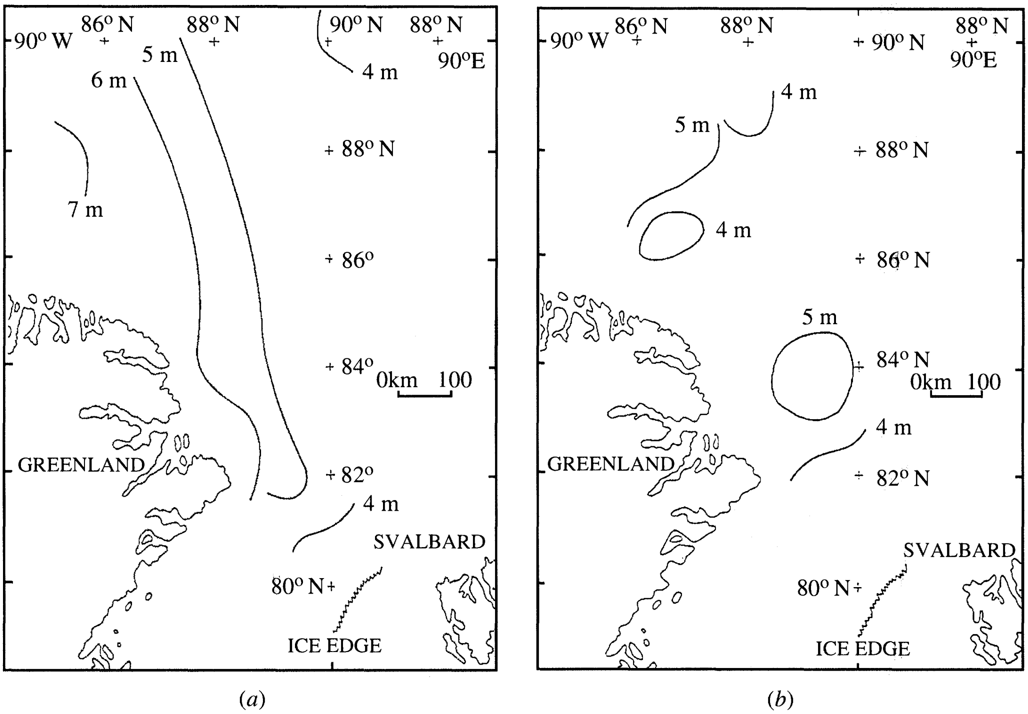

Figure 5. Contour maps of mean ice drafts from Eurasian Basin: (*a*) October 1976 and (*b*) May 1987. (After Wadhams 1990*a*.)

the acceleration and narrowing which occur in Fram Strait. The mean drafts from the three cruises were remarkably similar: 4.60 m in 1976; 4.75 m in 1979; and 4.85 m in 1985, despite being recorded in different seasons as well as different years.

Using newer data, Wadhams (1990*a*) compared data from a triangular region of 300 000 km^2 extending from north of Greenland to the North Pole, recorded in October 1976 and May 1987. Mean drafts were contoured to give the maps shown in figure 5. Over the whole area there was a decrease of 15% in mean draft from 5.34 m in 1976 to 4.55 m in 1987. The decrease was concentrated in the region south of 88° N and between 30° and 50° W. An analysis of the shape of the distributions showed that in 1987 there was more ice present in the form of young ice in refrozen leads (stretches of ice with draft less than 1 m) and as first-year ice (draft less than 2 m). There was less multi-year ice (interpreted as ice 2–5 m thick) and less ridging (ice more than 5 m thick) in 1987. The main contribution to the loss of volume was thus the replacement of multi-year and ridged ice by young and first-year ice.

To determine the mechanism, tracks of drifting buoys from the Arctic Ocean Buoy Program (Colony *et al.* 1991) were examined. Four buoys were in the region during the months prior to the 1987 cruise. Three of the buoys, located in the Beaufort Gyre, remained almost stationary during January – May 1987, while the fourth, in the Trans Polar Drift Stream, moved towards Fram Strait at a mean speed of 2 km d^{-1}. The result of this anomalous halting of the motion of part of the Beaufort Gyre should be a divergence within the experimental region, leading to the opening up of the pack and the creation of young and first-year ice. Thus an ice motion anomaly rather than, or as well as, an ice growth anomaly could cause the observed

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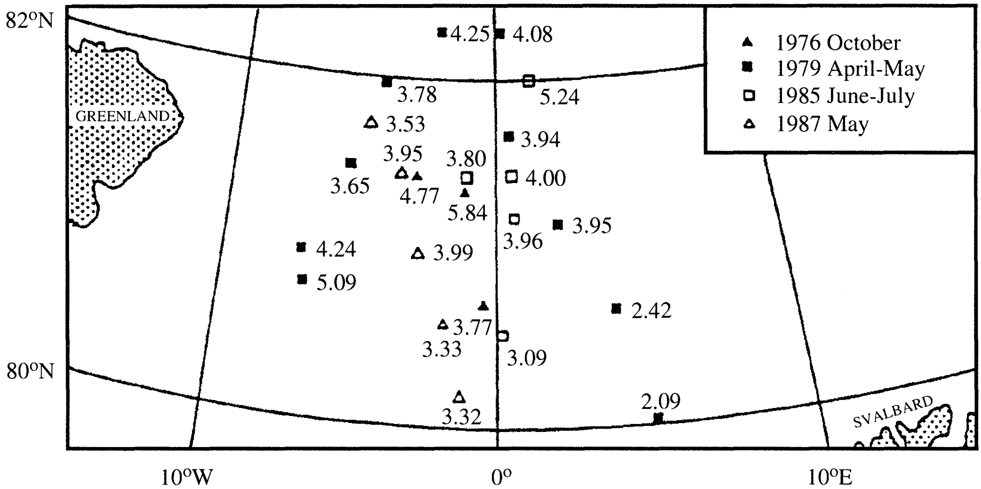

Figure 6. Comparisons of mean ice drafts measured in the region north of Fram Strait (after Wadhams 1992). Each position is the centroid of a 50 km section of submarine profile.

decrease in mean ice draft. This supposition is supported by modelling results from Chapman & Walsh (1993), who ran the Hibler ice model using daily wind forcing and monthly thermodynamic forcing, both varying interannually. They found that the simulated thicknesses showed negative anomalies in this region in May 1987 and positive anomalies in October 1976. A build-up of mean ice draft towards Greenland appears in the model predictions of Hibler (1980) and in the tentative seasonal climatology of Bourke & Garrett (1987) and was previously thought to be a stable aspect of ice climatology. Clearly the ice cover, like the ocean, possesses a weather as well as a climate.

The 1987 dataset allowed a further regional comparison immediately north of Fram Strait, between 82° N and 80° N, where data from 1976 1979 1985 and 1987 are now available. This is a mixing zone between streams of old, deformed ice moving S into Fram Strait from the North Pole region and SE from the region north of Greenland; and a stream of younger, less deformed ice moving SW from the seas north of Russia. Figure 6 shows all available mean drafts from 50 km sections (from Wadhams 1981, 1983, 1989 and current analyses). There is good consistency among the data, regardless of year or season; fluctuations appear to be random in character, and where results from different experiments lie close to one another, the mean drafts are usually similar. Only the 1976 data points appear anomalously thick.

McLaren *et al.* (1992) analysed 50 km and 100 km sections of ice profile centred on the North Pole from 6 cruises from 1977 to 1990. They found that the mean ice draft from 50 km sections in the late 1970s (1977, 1979) was 4.1 m (4.2, 4.0 m respectively), while the mean draft for the late 1980s was 3.45 m (2.8, 4.1, 3.3, 3.6 m for 1986, 1987, 1988, 1990). They showed using a t-test that the difference of 0.65 m (15%) between the means is significant only at the 20% level, i.e. non-significant. They also claimed that this showed that the 15% decline in mean drafts found by Wadhams (1990*a*) between 1976 and 1987 is similarly non-significant. The claim is invalid because the McLaren *et al.* (1992) comparison is between two datasets of total length 100 km and 200 km, while the Wadhams comparison is between two datasets of length 3900 km

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(1976) and 2200 km (1987), giving a much greater statistical stability to the mean values.

An alternative way of interpreting the results of McLaren *et al.* (1992) suggests that they are more significant than their authors state. When mean ice draft is computed from a dataset of finite length it is subject to a statistical variability which is independent of variations in ice climatology and which is merely due to the fact that the profile is sampling only a finite number of ice floes, leads and pressure ridges. This variability may be termed the sampling error. There is no *a priori* way of estimating what the sampling error should be. The only valid method is an experimental one: to examine the statistical stability of sections taken from long ice profiles obtained within a homogeneous ice regime at a single instant. Wadhams *et al.* (1992) discussed two such experimental datasets and found them compatible. The first comprised 23 50 km sections obtained from a restricted area of the Beaufort Sea (Wadhams & Horne 1980) while the second comprised 18 100 km sections from a homogeneous part of the central Arctic Basin (Wadhams 1981). If we assume that the ice regime was homogeneous, then the standard error in the means of the 50- or 100 km sections is the sampling error which we seek. The results from the two experimental datasets were that the 50 km sections gave (3.67 ± 0.19) m as the overall mean draft (a 5.2% sampling error) while the 100 km sections gave (4.51 ± 0.18) m (a 3.9% sampling error). The difference in mean is due to the different ice regimes, but the percentage sampling errors are almost identical, given that the sampling length is double in the second case (if they were completely identical, a 5.2% error in 50 km would become a 3.7% error in 100 km). Assuming that a 3.9% sampling error per 100 km is indeed characteristic of Arctic sea ice, and applying it to the McLaren *et al.* (1992) data listed above, we see that on the hypothesis that all 1970s data and all 1980s data were drawn from only two populations the means and sampling errors would be (4.1 ± 0.16) m (1970s) and (3.45 ± 0.10) m (1980s). The appropriate test for the significance of the difference between the means is to compute the d-statistic of the Fisher–Behrens distribution (Campbell 1974), which gives $d = 3.444$, significant at better than the 1% level. The data presented therefore do show evidence of a difference between 1970s and 1980s data at the North Pole.

Further data supportive of the above are some recently analysed, hitherto unpublished, profiles from the region within 1° of the North Pole, obtained by the author during an earlier phase of the 1987 experiment. An analysis of the 10 50 km sections obtained gives a mean draft of 3.845 m with a standard error of 0.155 m, implying a 4.0% sampling error in 50 km of track. This is lower than the error observed in the previous two experiments, and when applied to the McLaren *et al.* data gives an even more significant difference between the mean thicknesses for the two decades.

(*e*) *Summary*

From limited data comparisons made to date, the following results were obtained on thickness variability:

(i) Ice in parts of the Trans-Polar Drift Stream which are not heavily influenced by a downstream land boundary show interseasonal and interannual consistency in mean thickness, especially at latitudes from 84°30′ N to 80° N and longitudes near 0°.

(ii) Ice near the land boundary of Greenland can show large variability in mean draft, notably a significant decline between 1976 and 1987, but this can be ascribed to a variable balance between pressure ridge formation through convergence and open water formation through divergence. Deeper knowledge of ice dynamics and more adequate data are needed to understand these changes fully.

(iii) Data obtained from the North Pole region in 1977–90 show evidence of a decline in mean ice thickness in the late 1980s relative to the late 1970s.

(iv) Ice in the Canada Basin in summer also shows variability in mean ice draft, but here there is a free boundary with ice-free marginal seas, permitting relaxation of the ice cover into a less concentrated state under certain wind conditions.

There is no conclusive evidence of progressive thinning of the sea ice cover, as would be caused by the impact of the greenhouse effect, but the data are suggestive enough to make more systematic data collection essential.

4. Ice variations and ocean convection

One region in which ice retreat and thinning may have already triggered an ocean response is the centre of the Greenland Sea gyre. South of the main gyre centre an ice tongue usually develops during winter, growing eastward from the main East Greenland ice edge in 72–74° N latitude and often curving round to the northeast until it reaches east of the prime meridian. It is called Odden, its curvature embracing a bay of open water known as Nordbukta. Odden forms mainly by local ice production on the cold surface water of the Jan Mayen Polar Current (the southern part of the Greenland Sea Gyre), largely in the form of frazil and pancake ice, since the intense wave field inhibits ice sheet formation. Frazil and pancake production implies high growth rates and high salt fluxes, possibly on a cyclic basis related to cold air outbreaks from Greenland. In recent models of winter convection in the central Greenland Sea (Rudels 1990; J. Backhaus, personal communincation), this periodic salt flux is responsible for triggering narrow convective plumes.

Field operations during the European Subpolar Ocean Programme (ESOP) in 1993–94 involved the direct study of convective plumes and water structure in the central gyre region, and simultaneous investigations of ice characteristics (Wadhams 1994*b*). In 1993 Odden developed as a tongue, later an island, of dense pancake and frazil ice, whose ability to grow and melt rapidly gave the Odden a rapidly changing shape on SSM/I images. The ice physics programme (Wadhams & Viehoff 1994; Wadhams *et al.* 1994) involved recovery of pancakes and frazil, with measurements of thickness and salinity. By correlating ice properties to assumed age, an estimate was made of ice volume and salinity changes involved in the winter fluctuations of Odden; the resulting salt fluxes (Wadhams *et al.* 1995) could then be compared to fluxes required to trigger convection in model studies.

If under global warming ice were to cease forming in Odden it may cause deep convection to cease. Already tracer studies (Schlosser *et al.* 1991) show a severe reduction in the renewal of the deep waters of the Greenland Sea by convection during the last decade. Convection reached only 1000 m in 1993 and 400 m in 1994 when Odden did not form at all (D. Quadfasel, personal communication). If deep convection were to cease it would have a positive feedback effect on global warming, since the ability of the world ocean to sequester CO_2 through convection would be reduced (Rudels, this volume). A general weakening of the thermohaline circulation in the northern North Atlantic is predicted by Manabe & Stouffer (1994) based on the freshening of surface water due to melt of sea ice and ice sheets.

5. Climatic feedbacks and model predictions

(*a*) *Extent*

As surface air temperatures increase, it is legitimate to question how ice extent will vary. One test is to examine correlations between ice extent and past temperature changes. Chapman & Walsh (1993) showed that longitudinal patterns of sea ice variability in the Arctic during the periods 1961–75 and 1976–90 had a significant negative correlation with annual air temperature changes in the latitude zone 55–75° N, i.e. the latitude range corresponding approximately to the seasonal ice edge. Most transient general circulation models (GCMs) predict a retreat of sea ice: Manabe *et al.* (1990), for instance, predicted a significant northward retreat of the ice edge under a CO_2 doubling, with thinning throughout the ice cover but with the effect being greatest in the region 60–70° N, i.e. the sub-Arctic. Other recent models predict that eventually the ice cover will wholly disappear in summer, and will become essentially seasonal such as the Antarctic today.

(*b*) *Fast ice thickness*

The simplest effect of warming will be on the thickness of fast ice, which grows in fjords, bays and inlets in the Arctic, along the open coast in shallow water, and in channels of restricted dimensions. Here oceanic heat flux is negligible, and the ice thickness is determined by air temperature history modified by the thickness of the snow cover. Empirical relationships have been successfully developed relating thickness achieved to the degree-days of freezing since the beginning of winter (e.g. Bilello 1961). If the average daily air temperature increases by a known amount, the ultimate ice thickness will diminish by an amount which is easily extracted from these relationships, and the ice-free season will lengthen. Using this technique, Wadhams (1990*b*) predicted that in the Northwest Passage and Northern Sea Route an air temperature rise of 8 °C (equivalent to about a century of warming) will lead to the winter fast ice thickness declining from 1.8–2.5 m (depending on snow thickness) to 1.4–1.8 m and the ice-free season increasing from 41 to 100 days.

Even in this simple case, however, there is a feedback with snow thickness. If Arctic warming produces increased open water area and thus increased atmospheric water vapour content, it will lead to increased precipitation. Thicker snow cover decreases the growth rate of fast ice, as has been directly observed (Brown & Cote 1992), except if the thickness increases to the point where the snow does not all melt in summer, in which case the protection that it offers the ice surface from summer melt leads to a large increase in equilibrium ice thickness (Maykut & Untersteiner 1971).

(*c*) *Moving pack ice thickness*

In moving pack ice, thermodynamic growth and decay rates no longer determine the area-averaged mean thickness. Pressure ridge building causes a redistribution of ice from thinner to thicker categories, with the accompanying creation of open water areas. It also makes the ice cover as a whole more resistant to convergent than to divergent stresses, and this causes its motion field under wind stress to differ from that of the surface water. Thus the exchanges of heat, salt and momentum all differ from those that would occur in a fast ice cover.

The variable thickness has thermodynamic effects. It has been found that the overall area-averaged growth rate of ice is dominated (especially in autumn and early winter when much lead and ridge creation take place) by the small fraction of the sea surface occupied by ice less than 1 m thick (Hibler 1980). In fast ice, climatic warming increases sea-air heat transfer by reducing ice growth rates. However, over open leads a warming would decrease the sea-air heat transfer, so the area-averaged change in this quantity over moving ice (and hence its feedback effect on climatic change itself) depends on the change in the rate of creation of new lead area, which is itself a function of a change in the ice dynamics, either driving forces (wind field) or response (ice rheology).

The overall pattern of ice motion has other effects. In the Eurasian Basin the average surface ice drift pattern is a current (the Trans-Polar Drift Stream) which transports ice across the Basin, out through Fram Strait, and south via the East Greenland Current into the Greenland and Iceland Seas where it melts. A typical parcel of ice forms by freezing in the Basin, the latent heat being transferred to the atmosphere; is then transported southward (equivalent to a northward heat transport); and then when it melts in the Greenland or Iceland Sea it absorbs the latent heat required from the ocean. The net result is a heat transfer from the upper ocean in sub-Arctic seas into the atmosphere above the Arctic Basin. A change in area-averaged freezing rate in the Basin would thus cause a change of similar sign to the magnitude of this long-range heat transport. An identical argument applies to salt flux, which is positive into the upper ocean in ice growth areas and negative in melt areas. Thus salt is also transported northward via the southward ice drift. A relative increase in area-averaged melt would cause increased stabilization of the upper layer of polar surface water, and hence a reduction in heat flux by mixing across the pycnocline, while a relative increase in freezing would cause destabilization and possible overturning and convection.

Finally, Hibler (1989) has drawn attention to the role of ice deformation in reducing the sensitivity of ice thickness to global warming in areas of net convergence. The largest mean ice thicknesses in the Arctic – 7 m or more – occur off the Canadian Arctic Archipelago (Hibler 1979) where ice is driven towards a downstream land boundary. Here the mean ice thickness is determined by mechanical factors, largely the strength of the ice, which sets a limit to the amount of deformation by crushing that can occur. In this area the thickness is likely to be insensitive to atmospheric temperature changes. The main sensitivity would be to a change in the overall wind pattern over the Arctic.

Given the complexity of these interactions and feedbacks, it is not clear at present what the quantitative effect of an air temperature increase on the Arctic ice cover and upper ocean would be. Further sensitivity studies using coupled ocean-ice-atmosphere models are required.

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Discussion

M. Wallis (*University of Wales at Cardiff, Cardiff, UK*). Given that the rate of ice formation depends on areas of open water and of thin ice, and is balanced on average by transport through the Fram Strait, would the system not be stable to global warming and constrain possible changes in North Atlantic currents?

P. Wadhams. There certainly could be a negative thermodynamic feedback along the lines suggested by Dr Wallis, but ice–ocean models show that both the extent and thickness of Arctic sea ice are far more dependent on dynamics than on thermodynamics. Thus it is the effect of global change on Arctic winds that is likely to be the determining factor.

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