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Variability of convective conditions in the Greenland Sea

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Recent observations and data compilations show decadal and interannual variations in the depth of wintertime convection in the Greenland Sea. In a qualitative study the fluctuations are related to changes in wind and thermohaline forcing. Changes in both wind-stress curl and sea-ice cover concur with the results from hydrographic observations indicating that no renewal of deep water has taken place during the 1980s.

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Introduction

The process of convective renewal of waters in the Greenland Sea is a major influence on the thermohaline circulation of two systems (Aagaard *et al.*, 1985): on the one hand it is one of the sources for the formation of intermediate waters which leave the area as overflows across the Greenland–Scotland ridge and drive the global circulation system of the North Atlantic Deep Water. On the other hand it contributes significantly to the deep circulation internal to the Arctic Ocean/Nordic Seas system, which is confined to the area north of the Bering Strait and north of the Greenland–Scotland ridge. The division between the two systems can be defined by the sigma − 1 level of 32.785, which is found around 1500-m depth in the Nordic Seas. This is the maximum depth of winter convection that determines whether convective renewal is limited to intermediate water or affects both intermediate and deep levels.

With this background a study on the variability of convective depths in the Greenland Sea is important for the role of the oceans in climate. This contribution will illustrate the variability and discuss the mechanism causing it.

The convective system

A much improved database and conceptual modelling have led to a fairly consistent picture of convective water mass transformation in the Arctic Ocean/Nordic Seas system. The thermohaline part is dominated by the regionally varying interaction of three members in TS-space (Aagaard *et al.*, 1985; Rudels and Quadfasel, 1991), illustrated in Figure 2: cold and fresh water from runoff and melting, cold and saline water from brine released during sea-ice formation, and warm and saline water contributed from Atlantic inflows.

In the Arctic Ocean wide shelf areas the brine from seasonal sea-ice formation accumulates in troughs and trenches (Midttun, 1985) before spilling over the shelf edge. It sinks down the slopes as dense high salinity plumes at near freezing temperatures, entraining in particularly warm intermediate water of Atlantic origin, and results in Arctic Ocean Deep Water (AODW), the warmest and most saline deep water component in the Arctic Ocean/Nordic Seas system.

In the case of the deep Greenland Sea the saline plumes from sea-ice formation enrich the relatively fresh upper Arctic layer with salt, such that subsequent plumes penetrate the halocline into warm intermediate layers. To preserve continuity, the warm water rises to the surface, where it melts the ice and thereby stops the convection. After cooling and new ice formation a new haline convection cycle can start (Rudels, 1990). This results: in a stepwise increase of convective depths (Rudels *et al.*, 1989), in extremely weak mean stratification (Fig. 2), and in Greenland Sea Deep Water (GSDW), the coldest and freshest deep water component. The other deep water components, which are TS-wise between the characteristics of those from the Arctic Ocean and the Greenland Sea, are a product of mixing between the latter (Aagaard *et al.*, 1991).

The thermohaline part of the convective system

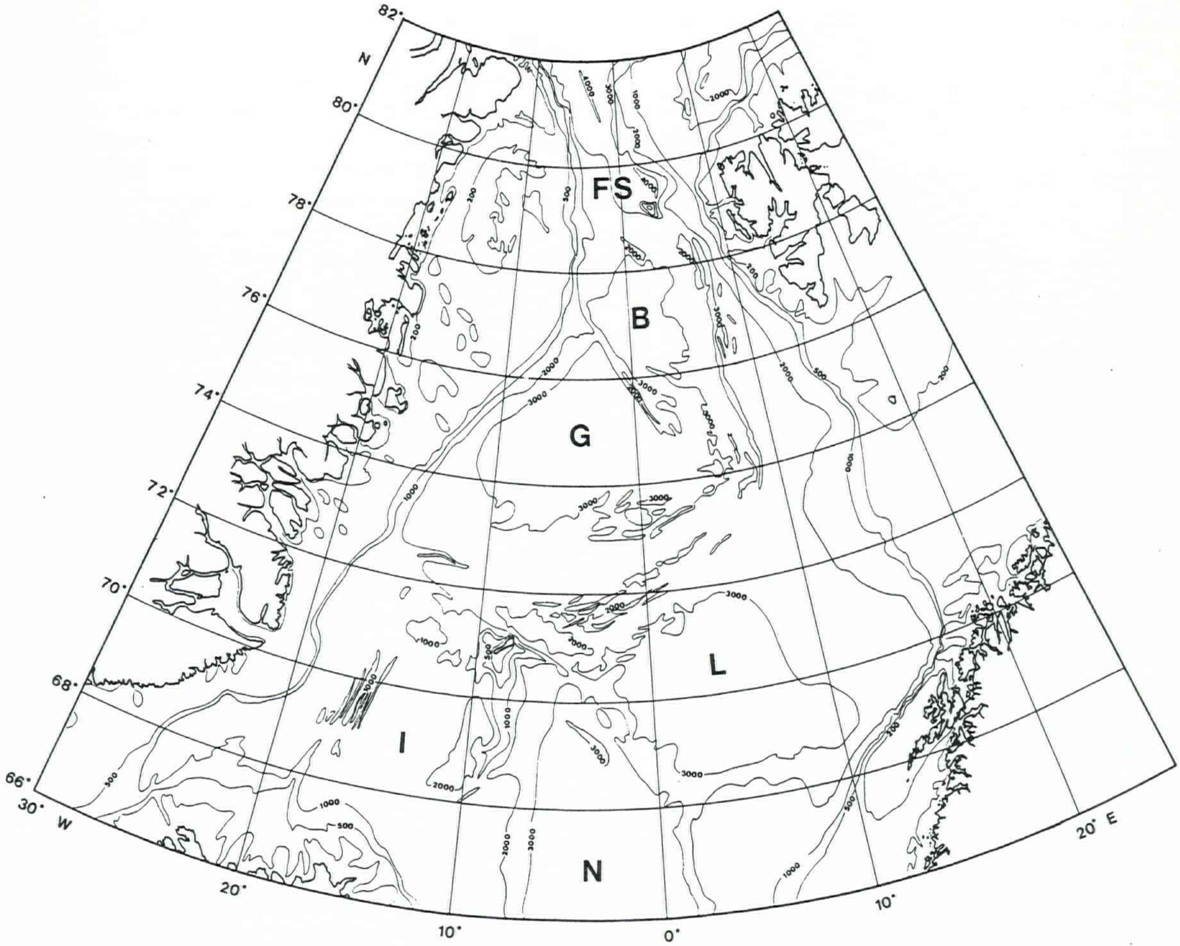

Figure 1. Topography of the Nordic Seas. FS = Fram Strait, B = Boreas Basin, G = Greenland Basin, B+G = Greenland Sea Basin, L = Lofoten Basin, N = Norwegian Basin, I = Iceland Sea Basin.

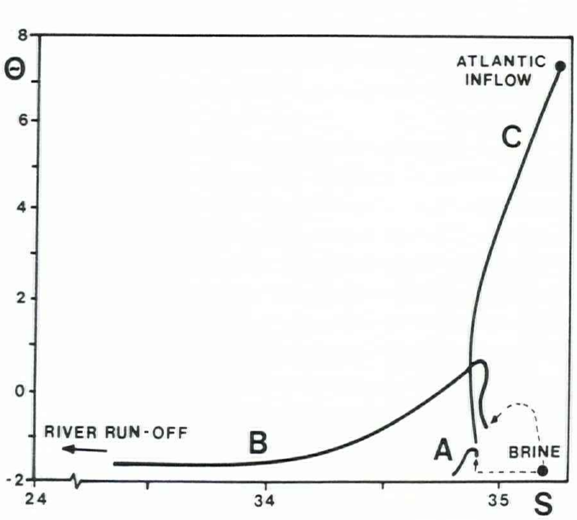

Figure 2. Schematic TS-relationships for the Arctic Ocean/ Nordic Seas system. A = Greenland Sea, B = Arctic Ocean, C = Norwegian Sea. The end members are named.

closely interacts with the wind-driven part. Next in importance to the large-scale wind forcing, which governs the flux of Atlantic waters and Polar waters throughout the system, the basin-scale wind forcing is particularly important. Legutke (1990) has shown that it is the matching between the along-isobath circulation and the contour integral of the wind stress along f/H-contours around the basis that determines the effectiveness of wind forcing. This has two consequences: with varying scales of wind forcing, different-sized cyclonic gyres are excited in accordance with the different sizes of topographic basins (see Fig. 1). For the Greenland Sea this ranges from a one-gyre situation covering both the Greenland and Boreas Basin to a two-gyre situation with separate cyclonic circulations for each of the two basins. The second consequence is related to the varying intensity of wind forcing. It results in a varying deformation of the interface between the deep and intermediate waters, i.e. a varying intensity of the "doming" of the deep water in the gyres.

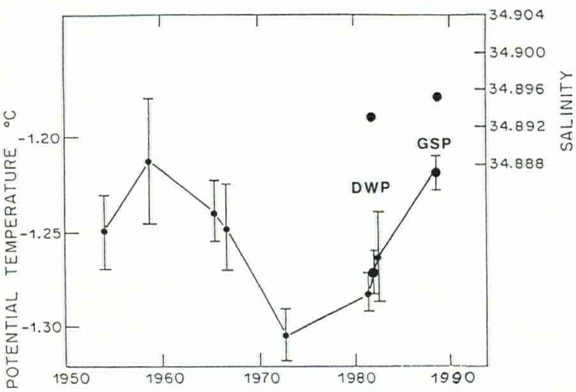

Figure 3. Time series of average potential temperature below 2000 m in the Greenland Basins. Data from stations with bottom depth exceeding 3000 m. The two single dots mark corresponding salinities obtained during the ICES Deep Water Project (DWP) (Clarke *et al.*, 1990) and during the Greenland Sea Project (GSP) (GSP Group, 1991). Θ- and S-scales were chosen such that in the event of constant deep-water density Θ- and S-variations would show as parallel curves.

Observed variations in convective activity

The reasons for the scarcity of time-series information on the convective renewal of water masses in the Arctic Ocean/Nordic Seas system are obvious: extreme difficulty of access to the Arctic Ocean and signal amplitudes close to the measurement resolution for the traditional oceanographic parameters are two major reasons. It is only since the mid-1970s that salinities have been routinely measured to 0.002 psu and that anthropogenic tracers have become available as adequate parameters. Consequently, a time series on convective renewal over a decadal time scale can only be inferred from temperature measurements in the Greenland Basin (Fig. 3). They show two warm (late 1950s and late 1980s) and one cold (early 1970s) period for the deep water, suggesting a time scale in the order of 30 years. Since the water masses of the gyre of the Greenland Basin are topographically trapped and have horizontal temperature gradients much smaller than the observed variations in time, the observed cooling can be interpreted as being dominated by renewal of GSDW, whereas the warming periods are dominated by mixing with waters around the gyre.

There are two salinity values of sufficient quality added to Figure 3. The salinity scale was chosen such that in the case of constant density the temperature and salinity curves would be parallel. This is not the case, indicating that non-isopycnic mixing occurs (Aagaard *et al.*, 1991).

The interpretation of the warming period for the 1980s in the deep water can be substantiated from increased observational activity, resulting in more detailed estimates of the annual depth of convective overturning. The Deep Water Project (Clarke *et al.*, 1990) surveyed the Greenland Sea in winter and early summer 1982 and found the convection depth at 500 m. For the period 1986–1989 (GSP-Group, 1990) convective depths were determined to be 200, 1300 and 1600 m for the Greenland Sea gyre. In combination with analyses of anthropogenic tracers covering the period 1972 to 1989, which show a cessation of deep water formation from 1980 onwards (Schlosser *et al.*, 1991), it can be concluded that the 1980s have to be characterized by no detectable renewal of waters in the Greenland Sea below 1600 m. Observations by Nagurny and Popov (1985) suggest convection down to the bottom in winter 1984. This is regarded as questionable, since salinity time series presented for the GSDW by Popov (pers. comm.) show unreasonable fluctuations.

There are two pieces of observational evidence for the cooling period in the early 1970s. Malmberg (1983) reports having observed convective conditions of the central Greenland Basin down to 3500 m in winter 1971. The other evidence comes from tritium determinations (Schlosser *et al.*, 1991), which suggest intensive deep water formation having occurred between the mid-1960s and the early 1970s.

Causes of convective variability

The description of the convective variability of the Greenland Sea has revealed decadal and interannual time scales. The literature offers only a few discussions on possible causes. Aagaard (1968) correlated long-term changes and interannual changes of intermediate and deep water temperatures with a winter cooling index derived from air temperature observations. The results support Nansen's (1906) concept of the overturning of cooled surface waters, neglecting any haline component in the system.

In discussing the wind causing variability in the convective intensity we take up Jónsson's (1991a) suggestion: an analysis of wind-stress curl over the Nordic Seas by means of empirical orthogonal functions. Jónsson (1991b) shows significant decadal and interannual variations of intensity and spatial scales. The lower modes, which compare in scale to the Greenland Sea, show interannual variability with high energy in the 1960s and early 1970s and low energy in the late 1970s and 1980s. Figure 4 is a relevant presentation of this scale. The higher modes compare in scale with the individual basins (Fig. 1), have little energy in the 1960s and 1970s, and reach high levels in the 1980s.

With respect to wind-driven circulation and convection a one-gyre situation is favoured for the Greenland Sea because of the dominance of the lower modes in the

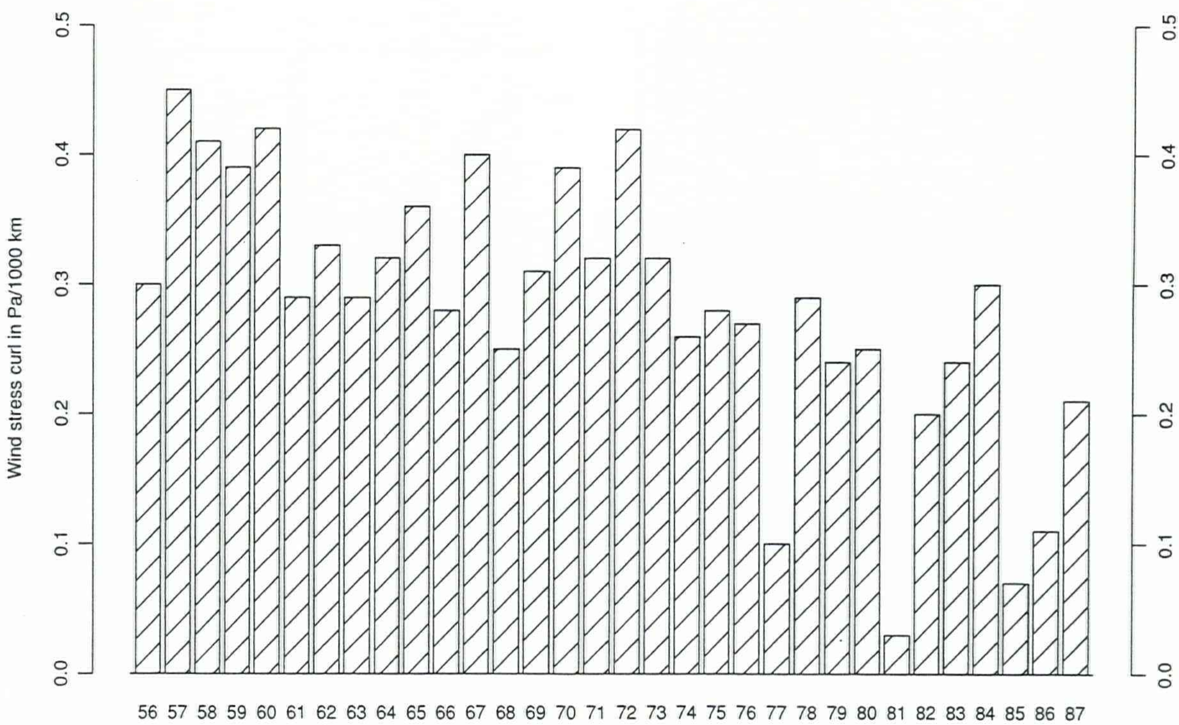

Figure 4. Interannual variations of wind stress curl over the Greenland Sea. The values are averages over an area of 200 000 km^2 in the central part of the Greenland Sea and over a one-year period from summer of year -1 to the following summer. Note the decrease in the late 1970s and 1980s.

1960s and early 1970s, with strong doming, upper layer divergence due to Ekman pumping and well developed Arctic and Polar fronts separating the convective regime from the Polar and Atlantic waters. This situation presents favourable conditions for deep reaching convection. During the 1980s, the lower mode forcing was less intense and the potential for the generation of separate gyres in the Greenland and Boreas Basin was much larger, thus repeatedly interrupting the prevailing, but weakened, one-gyre situation for certain periods. Quadfasel and Meincke (1987) indeed observed a two-gyre situation in 1986. This causes advective "leakages" of the frontal zones in the area of the gyre separation, bringing fresh Polar waters into the surface layer and warm Atlantic water into the intermediate layers of the central Greenland Sea. Both effects are adverse to convection. Aagaard and Carmack (1989) have shown this for excess fresh water input into the convective gyre and Figure 5 shows this for an intrusion of warm Atlantic water. It occurred between August 1986 and March 1987, restricting the convective depth in the Greenland Basin to 200 m because too much heat had to be removed by the convective cycles during one winter.

This makes decadal and interannual changes in the wind field one of the likely causes of corresponding changes in convective depths in the Greenland Sea – quantitative proof has still to be evaluated.

Turning to thermohaline causes of convective variability, the convective processes in the Arctic Ocean and Greenland Sea, as described in section 2 above, imply net fluxes of heat and salt as follows: in the Arctic Ocean the downslope penetration of high salinity shelf water at freezing temperatures and the entrainment of warm Atlantic waters produce a net downward flux of heat and salt for the Arctic Basins. In the Greenland Sea the entrainment of fresh water in haline plumes at freezing temperatures and the penetrative character of the plumes in the intermediate warm levels result in an upward flux of heat and salt. To compensate for gains and losses in the interior of the Arctic Ocean/Nordic Seas system the two reservoirs have to couple by exchanges through the Fram Strait. Figure 6 is a schematic diagram of this concept. Since the size of the reservoirs is vastly different, the forcing of the fluxes differs and since non-linearities of the TS-relationships are involved (see Fig. 3) the coupled system will react with oscillations. No time-series data are available, but the description of the "Great Salinity Anomaly" by Dickson *et al.* (1988) hints that such coupling exists: the authors have described an anomalous high freshwater accumulation in the area north of Iceland in the late 1960s. It was believed to have originated from an increased outflow of fresh water from the Arctic Ocean. This outflow, which Aagaard and Carmack (1989) esti-

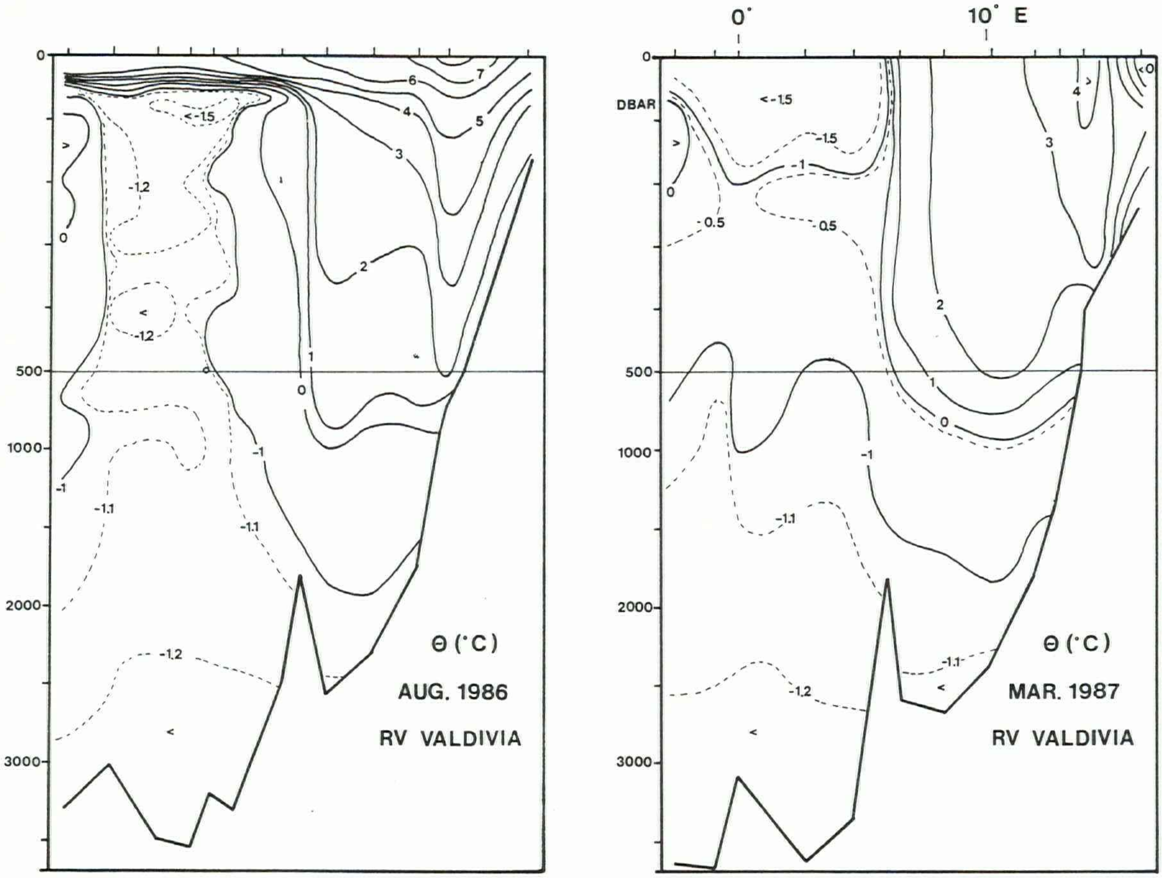

Figure 5. Temperature distribution for summer 1986 (left) and winter 1987 (right) along 74°45′N starting off at Bear Island and running due west across the central Greenland Basin. Note the warm intrusion in the interval 200 to 500 m observed in winter.

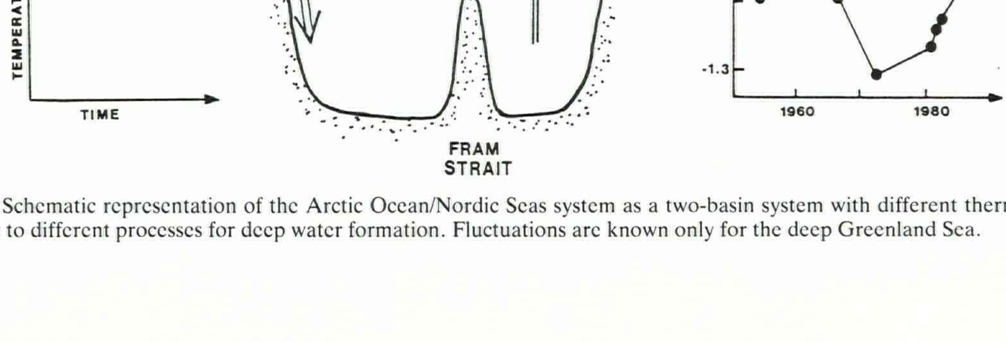

Figure 6. Schematic representation of the Arctic Ocean/Nordic Seas system as a two-basin system with different thermohaline fluxes due to different processes for deep water formation. Fluctuations are known only for the deep Greenland Sea.

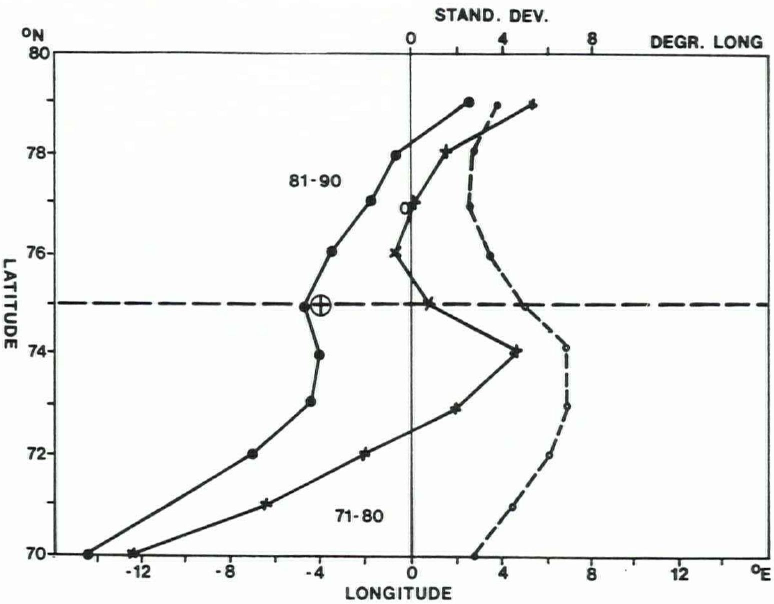

Figure 7. Extent of sea ice in the Greenland Sea. Solid lines: Wintertime (Jan to Mar) location of ice edge for the periods 1971–1980 and 1981–1990. Data sources: 1971–1980 Vinje (1984), 1981–1990 Joint Ice Center. Dashed line: Standard deviation as obtained for the period 1981–1990. No compatible estimates were possible for the 1970s. The centre for convection in the Greenland Sea is marked at 75°N, 04°W.

mated to represent a 30% increase in the mean freshwater flux from the Arctic Ocean, obviously did not immediately affect the favourable convective conditions in the Greenland Sea while passing en route to the Iceland Sea. However, while propagating from the Iceland Sea around the subpolar gyre the freshwater excess reduced convection in the Labrador Sea in 1972 and, after returning to the Greenland Sea in 1981, convective conditions became less favourable. A similar event occurred at the beginning of this century. We believe that this is some indication of the validity of the model on the generations of variability in deep water properties by thermohaline coupling of Arctic Ocean/Nordic Seas reservoirs.

The third reason for the variability is more straightforward and relates to local thermohaline forcing by sea-ice formation. As described in section 2, cycles of ice formation and melting are the surface expression of convective activity in the Greenland Sea. According to Rudels (1990) these cycles occur on time scales of the order of several days and on spatial scales of the order of kilometres. There are no routine data allowing direct observations of such cycles. Indirect indication can be obtained from the routine ice charts, which have a time resolution of the same order as can be expected for the ice cycles and a spatial resolution which is much larger. The most consistent parameter in these charts over the period 1970 to 1990 is the ice edge location. In Figures 7 and 8 the ice edge and its variance are plotted for decadal and interannual time scales.

By assuming the centre of convective activity to be at 75°N, 04°W (GSP-Group, 1990), by further assuming that a significant portion of the ice observed in the central Greenland Sea during winter has been locally formed and by implying the observed variance in the central Greenland Sea to be significantly related to ice formation/ice melting cycles, the figures can be discussed as follows: on the decadal time scale (Fig. 7) there was more ice formed during the 1970s than during the 1980s. Taking the variance of the 1980s to hold also for the 1970s, then the centre of convection was occasionally ice-free during the winter periods of the 1970s, but rather frequently during the 1980s. This result is consistent with the observed convective activity described in section 3, if more ice and more occasional open water mean more salt brine release into the system.

On the interannual time scale (Fig. 8) we find the central winter months in 1988 and 1989 just as has been discussed above for the 1980s. In 1987, however, there was so much ice that the centre of convection was permanently under an ice cover. This corresponds well with Figure 5, where the heat content of a warm intrusion of Atlantic water at intermediate depth could not be released to the atmosphere and consequently convective depths were limited to 200 m.

Conclusions

The foregoing discussion has indicated relationships between decadal and interannual variations in the depth

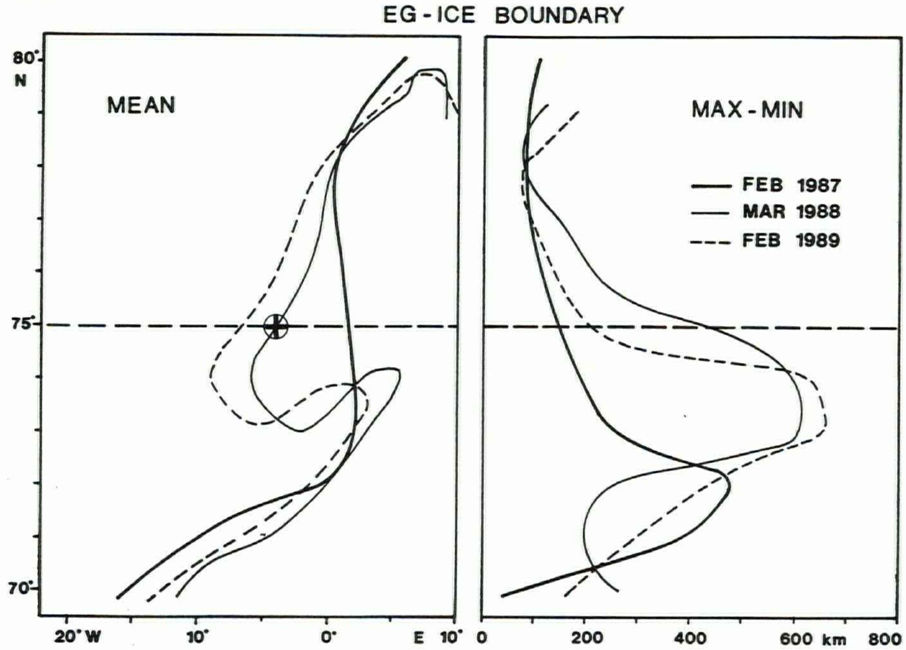

Figure 8. Monthly mean ice edge and its longitudinal variation (maximum extent minus minimum extent) for February 1987, March 1988, and February 1989. Data source: Joint Ice Center. The centre for convection in the Greenland Sea is marked at 75°N, 04°W.

of the wintertime convection and changes in forcing by wind and thermohaline processes. One cannot at present get much further along this route: the limitations on data availability and data consistency are one reason. Whereas limitations on data availability are obvious for the ocean observations, the data consistency problem concerns the longer term wind and ice data for the spatial scale of the circulation in the Greenland Sea. There have been significant changes in observation techniques for sea ice and in model-supported interpolation schemes for gridded meteorological data within the last two decades. A second reason is the fact that wind and thermohaline forcing are not independent of each other. Consequently, progress in further understanding variability in Greenland Sea convection is expected from mesoscale circulation modelling on the one hand and from continuation of relevant observations in the Greenland Sea , and hopefully in the Fram Strait and the Arctic Ocean, on the other.

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