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THE ICE COVER OF THE GREENLAND SEA

AN EVALUATION OF OCEANOGRAPHIC AND METEOROLOGICAL CAUSES FOR YEAR-TO-YEAR VARIATIONS

BY

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WITH 21 FIGURES AND 6 TABLES IN THE TEXT

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Abstract

A criterion is defined to compare seasonal ice coverage in the Greenland Sea for the years 1900–57, and the areal coverage is graphed using the 1898–1913 average as a standard. The factors yielding possible influence on short-term variations of the ice cover are examined individually and their relative importance established.

The influence of ocean currents is evaluated by analysis of hydrosections across the East Greenland Current at 74°–76° N and across the North Atlantic Current in the Faeroe-Shetland channel. Data from the latter area are used for numerical analysis of heat imported to the Greenland Sea by the North Atlantic Current in the 1927–52 period. Details about the Irminger Current's behavior are derived from station data from Denmark Strait and from surface temperatures at Selvogsbanki south of Iceland. Year-to-year variations are found to exist in the flow volumes of all three currents, and correlations with seasonal ice coverage in the Greenland Sea are shown.

Above-average precipitation in conjunction with below-average storm activity is found to have negligible influence on the ice regime, and no significant correlation is found. The possible effect of evaporation is computed to be far below the threshold of detectability. Air temperatures in the Norwegian-Greenland Sea region display a trend of increase throughout the period studied, in harmony with a concurrent trend of decreasing ice cover; but no causal relationship is in evidence.

The effects of strengths and directions of predominant winds are examined, and good correlations are shown between ice cover fluctuations and easterly wind components at Norwegian coastal stations. At the points of major currents' entrances to and exits from the Greenland Sea the wind effects are complex and cannot be fully evaluated on the basis of existing data.

The fluctuations of ocean currents entering and leaving the Greenland Sea and of water movements within the Greenland Sea remain as the apparent determinant of year-to-year variations of the ice cover.

I. INTRODUCTION

Investigators of Arctic ice fluctuations have almost to a man concerned themselves with effects of rather than causes for variations in the severity of ice seasons. Consequently, the profound influence exerted by the ice cover upon European weather has been well documented by KNUDSEN (1905), HENRY (1929), PETTERSON (1929), MONTGOMERY (1940) and others, but almost no effort has been made to explore the factors that might bear on the phenomenon, including those in the realm of oceanography.

In a recent manuscript on the ocean's role in climatic change (WEYL, 1966) it was suggested that low storm activity in the period between ice melting and freezing in the Atlantic Arctic could significantly influence subsequent ice formation. The author argued that normal excess og precipitation over evaporation would tend to form a low-salinity surface water lens. With stable quiet weather preventing admixture of deeper, high-salinity water, sea ice in the following winter would freeze more rapidly and extend farther south than would be the case after a stormy summer. The possibility of the existence of any such simple direct relationship seemed utterly intriguing.

As it turns out, no such causal relationship is in evidence, but the path of inquiry led to the broader question of possible relationships between ice fluctuations and measurable variations in oceanographic and meteorological parameters. The factors which could influence the ice regime can be enumerated as follows:

1. The magnitudes, distribution into Greenland Sea, salinities and surface temperatures of –
   a) the East Greenland Current,
   b) the North Atlantic Current
   c) the Irminger Current;
2. Precipitation, as it dilutes the surface water and raises the freezing point;
3. Evaporation, as it cools the surface and raises the surface salinity;

4. Air temperatures, as they affect the surface water temperatures;
5. Winds, as they affect the oceans currents and the ice transports.

Since variations do exist in both ice conditions and the oceanographic and meteorological parameters, it is my thesis that close, causal relationships exist, some of which can be roughly verified on the basis of existing data, and all of which can be quantified and perhaps predicted, given enough observational material. In the study supporting this thesis it is endeavored to encompass all of the factors influencing the ice conditions, rather than concentrating on one or two.

II. ESTABLISMENT OF AN ICE INDEX

As winter settles, heavy pans and floes begin their influx into the Greenland Sea, Figure 1, reinforced by quantities of locally formed ice. The accumulation spreads along the coast of East Greenland, signalling at successively lower latitudes the beginning of the ice season. Throughout the winter and spring, large masses of ice continue to push southward and to spread away from the coast. Denmark Strait is normally more or less chocked in spring, while in a bad year the ice blankets the northern coast of Iceland at that season.

The configuration of the mean ice limit points up the influences of oceanographic and meteorological factors upon the ice regime. North of Iceland the fringe in spring displays a characteristic curve toward the east, as shown in Figure 2, probably under the control of the East Iceland Current. An equally impressive feature is a V-shaped retrenchment immediately west of Spitzbergen, an unmistakable effect of the lingering warmth in the eddies of the northward thrusting branch of the North Atlantic Current. A less distinct indent just north of Jan Mayen has been in evidence in most seasons, presumably caused by a smaller counter-clockwise eddy off the West Spitzbergen Current.[1])

Establishment of a practical criterion for severity of a given ice season, or definition of an "ice index," has been attempted in the past by MEINARDUS (1906) and others.[2]) The method devised by MEINARDUS takes into account the duration of ice occurrence on the Iceland coast as well as the severity of the observed condition. The effect of weighting severe months is to enhance a graphic display and facilitate identification of the cyclic nature of the variations. Despite this advantage, MEINARDUS's method will not be emulated here, as the assessment of severity is a subjective matter open to substancial error.

The criterion used in the following is one of areal extent of sea

[1]) A favorite seal hunting ground in times past, it was known to sealers as "Nord-Bukta" (North Bay).

[2]) SCHOTT (1904) considered the position of the ice limit for each month of the season being evaluated. Smith (1931) used a scale of 1–10 to indicate degree of coverage in the area. Hansen's method (1943) is mentioned in chapter III.

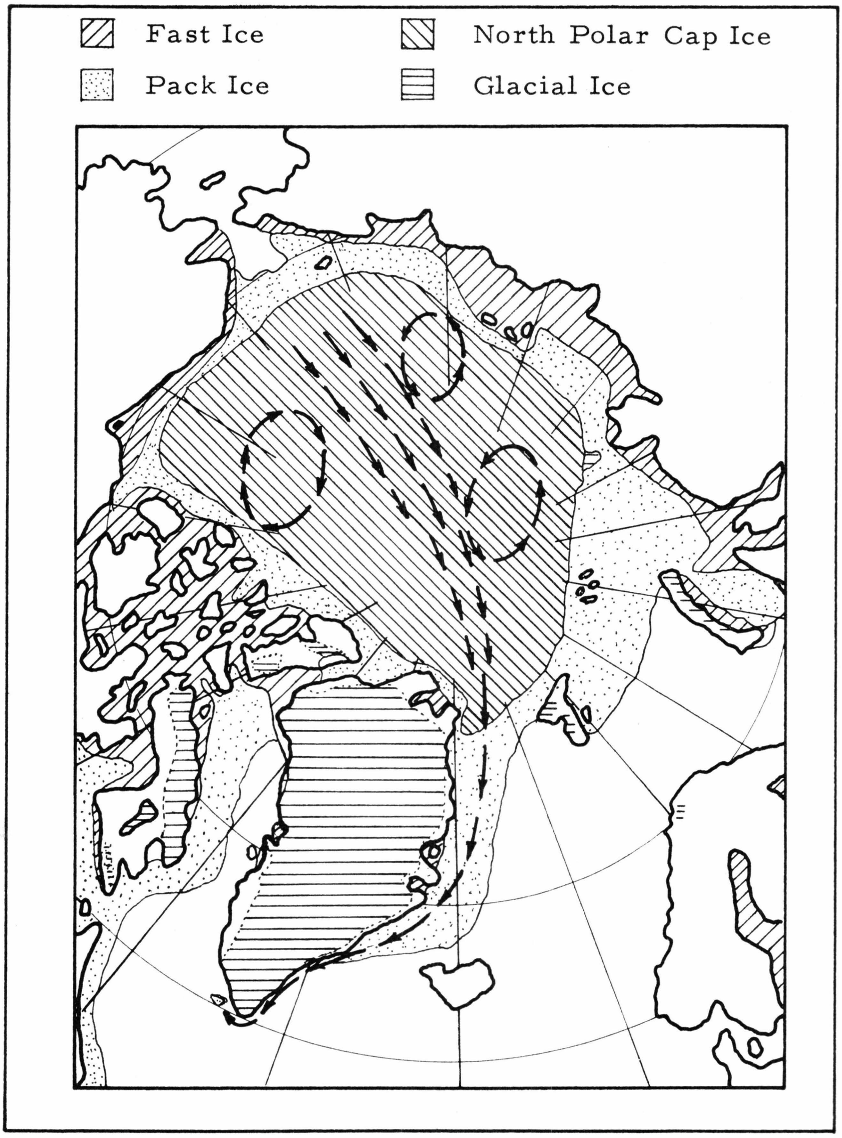


Figure 1. Distribution of Arctic ice types.

surface coverage between the meridian through Kap Farvel and 15° E.[3]) The Danish Meteorological Institute has compiled a detailed record of

[3]) There is apparently no correlation between ice conditions in the Greenland Sea and those in the Barents Sea (SCHELL, 1940). Hence separate treatment of Greenland Sea is justified.

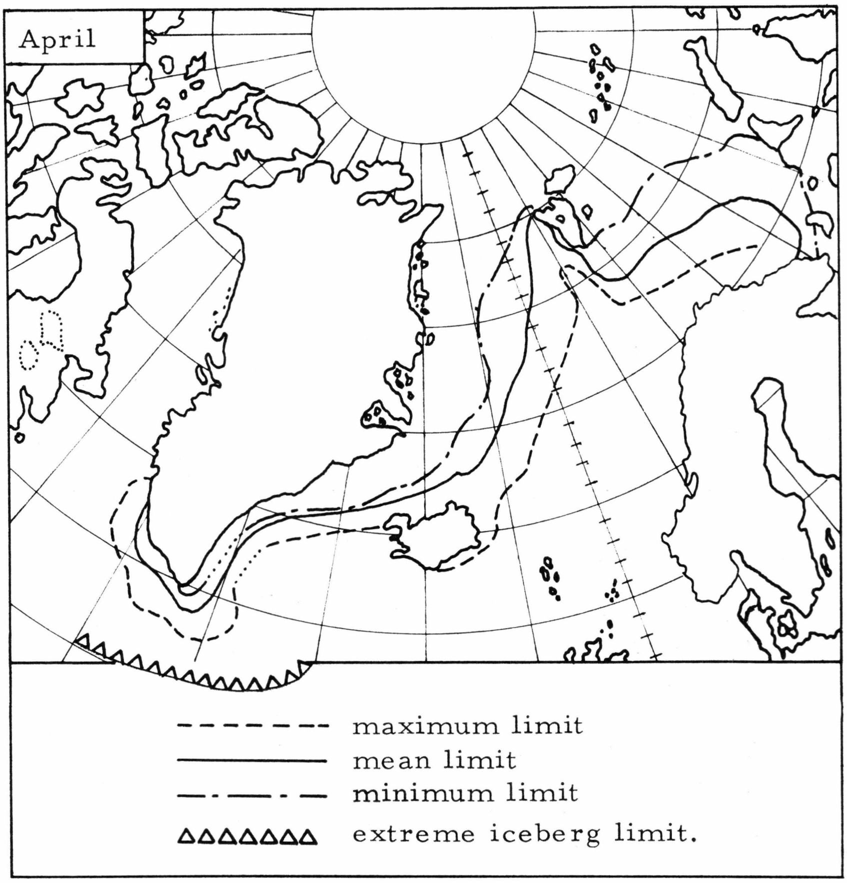


Figure 2.

ice conditions in the Arctic seas since 1896, and its annually published charts and reports have provided the basic data.[4])

Figure 2 shows the average and extreme ice limits in the month of April for the period 1898–1913. The April average for this period serves in the present study as a "standard" against which all subsequent April charts have been compared. Deviations from this standard have been measured for each of the years 1900–39 and 1946–57, counted positive when the ice cover exceeded the area enclosed by the standard limit, negative when in a given year the ice covered a lesser area of the sea surface.

[4]) The charts are identical in scale and layout from 1899 to 1957, hence provide a good comparison between years of this period. No data exist from the war years 1940–45. The charts' title "The State of the Ice in the Arctic Seas" was in 1957 changed to "The Ice Conditions in Greenland Waters".

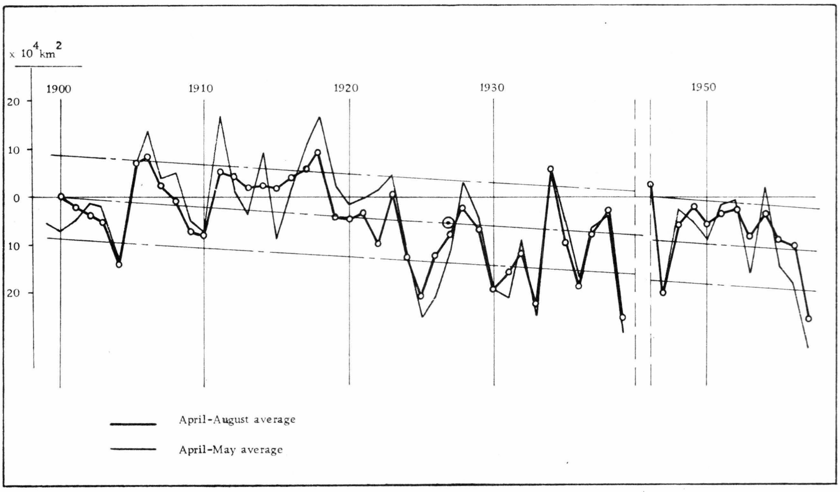


Figure 3. Extent of ice cover relative to 1898–1913 average (0-line).

The same procedure has been followed for the months of May through August, and the average of all five months provides the final index for a given year. The ice fluctuations are graphically shown in Figure 3, where the 0-line represents the 1898–1913 average, and the ice cover's deviation is shown by year in ten-thousands of square kilometers. It is apparent that a trend exists toward less severe ice winters. Assuming the regression to be linear, the estimated regression equation is

$$\bar{y}_x = a + b\,(x - \bar{x})$$

where $a = \frac{\Sigma y}{n} = -5.72$ and $b = \frac{SP}{SS} = -0.066$. The resulting regression line is indicated in the graph, together with the lines representing $\pm s$, where $s = \frac{(\Sigma(y - y_x)^2)^{1/2}}{n-2}$.

MEINARDUS (1906) found April and May to be the peak months of the ice season. Ice was sighted most frequently and prevailed longer in those months than in any other. The average of April and May has been shown separately in Figure 3.

The source material is subject to errors arising from differences in observers' experience, size (height) of reporting vessel, faulty positioning of reporting vessel[5]) and particularly from occasional paucity

[5]) Prior to the 1920s the observing vessels lacked wireless, a circumstance severely impairing longitude determination.

of observations[6]) necessitating extensive extrapolations in drawing the charts.

In the present study, however, extreme precision in the relative magnitude of an ice season is of less significance than a clear indication of a change from light to severe ice conditions, or vice versa, in consecutive years. Several such sequences are readily identifiable from the graph and are used in the following.

[6]) Decline of seal hunting early in the century was accompanied by a corresponding reduction in ice reports from some crucial areas north of Jan Mayen.

III. PREVIOUS WORK

Although the explorations of the 18th and 19th centuries were motivated largely by a quest for geographical discovery, the task of gathering scientific knowledge early became an important adjunct to the aims of the explorers. SCORESBY thus gave account in 1820 of ice drift on the East Greenland Current, and BEECHEY followed with a discussion of that current in 1843. Drifting ice off Iceland was mentioned by IRMINGER in 1891, and at about the same time we find the first reference, by LYELL,[7]) to the ice drifts' influence and periodicity. Subsequent work on the ice fluctuations can be classified into one of two categories labelled cause and effect, respectively. The latter is extensive, embracing work of numerous investigators,[8]) whereas the former is quite limited, though of primary concern here.

The original and classic work on ice fluctuations *per se* is a paper by THORODDSEN from 1884. After researching original Icelandic sources, THORODDSEN described years of severe ice conditions off the Icelandic coasts in some detail, covering the period from 1233 to 1883.

In 1906, MEINARDUS published his previously mentioned study of ice conditions in the 19th century, in which he showed peak occurrences to be repetitive with a period of $4^1/_2$ years. He believed sunspots to be a cause and endeavored to show a relationship with an 11.1-year sunspot cycle.

Of more recent date is an interesting study by HANSEN and SVEISTRUP (1943) of Arctic ice fluctuations in Julianehåb Bay during the years 1901–37. Julianehåb is located some 90 miles northwest of Kap Farvel, and in different years Arctic ice in varying quantities round the

[7]) "It is a well-known fact that every four or five years a large number of icebergs, floating down from Greenland, double Cape Langaness, and are stranded on the west coast of Iceland. The inhabitants are then aware that their crops of hay will fail, in consequence of fogs which are generated almost incessantly; and the dearth of food is not confined to the land, for the temperature of the water is so changed that the fish entirely desert the coast." (LYELL, 1866).

[8]) With no pretense of a complete bibliography, some of the major works were published by HANN (1904), BRENNECKE (1904), KNUDSEN (1905), HELLAND HANSEN (1920), WIESE (1924), BROOKS and QUENNELL (1928), NANSEN (1928), PETTERSON (1929), and SCHELL (1955).

cape and accumulate along the Greenland coast in the Labrador Sea. The two workers used this ice cover west of the cape as their index and looked into its correlations with surface water temperatures, air temperatures and winds. The surface temperature data derived from sundry merchant ships' reports from the area between the Hebrides and the Faeroes. Air temperatures were taken from Cape North in Norway, Svalbard, Jan Mayen and Bear Island, and the winds from Grimsey and Angmagssalik. HANSEN and SVEISTRUP found correlations of some significance and concluded that the fluctuations have no simple cause but are due to the interaction of a number of factors.

A detailed treatment of the East Greenland ice was published by KOCH in 1945, concluding a lifetime of field work in the region. KOCH discussed at length the manifestations and possible reasons for the long-term fluctuations, which caused the demise of the Norse culture in Greenland in the 15th or early 16th century. The short-term behavior of the ice during the period 1898–1938 was also described with great thoroughness and superbly documented, but no short term causes were sought. MEINARDUS is thus the last scientist to concern himself with the causes for the ice fluctuations. In the more than six decades since his work was published, research on the subject shows a void.

Much effort has been expended, however, on collection of basic data which in the following are brought to bear on the Greenland Sea's ice phenomena. It is therefore appropriate to discuss some of their sources. In the oceanographic field, the *Fram* voyage in 1893–96 marked the starting of modern systematic investigations in the Arctic, and since NANSEN's successful return numerous expeditions, both single-ship missions and more elaborate land-sea efforts, have brought back hydrographic data.[9])

Regular collection and issuance of weather data began with the publication of the Danish annual *Meteorologisk Aarbog* in 1873, which lists observations from several Greenland stations. In Norway, *Jahrbuch des norwegischen meteorologischen Instituts* commenced in 1874 to provide data from points in Norway, adding Spitzbergen in 1915, Jan Mayen in 1922 and Bear Island in 1928. Icelandic weather data have been published in *Veðráttan* since 1920, when this task was transferred from the Danish

[9]) A complete enumeration of these efforts is difficult, as many Danish investigations of the Greenland regions touch upon hydrography incidentally and peripherally, but the following post-*Fram* ventures must be included: *Nathorst* expedition (1898 and 1899); *Gjøa* (1901); *Michael Sars* (1901–04); *Belgica* (1905); *Danmark Ekspeditionen* (1906–08); *Alabama* (1909–12); *Conrad Holmboe* (1923); *Godthaab* (1924, 1930); *Øst* (1929); *Meteor* (1928, 1929, 1933, 1935); *Tre Aars Ekspeditionen* (Danish, 1931–34); *Polarbjørn* (1931–32); *Polaris* (1932); *Heimland I* (1933); *North Pole* (1937–39); *Atha* (1954); *Scotia* (1954); *Dana* (1955–59); *Ob* (1956); *Johan Hjort* (1958, 1961–62); *Conrad Holt* (1960); *Edisto* (1961, 1964, 1965); *Westwind* (1963).

to the Icelandic meteorological service. At the same time, the reporting was expanded to comprise many additional stations in Iceland proper and on offshore islands. From the Shetlands, weather data have been listed in the British *Weekly Weather Report* published by the Meteorological Office in London since 1892.

Centralized ice reporting was organized by the Danish Meteorological Institute in 1896, and the task was formally requested from the institute by the VIIth International Geographical Congress in Berlin in 1899. The reports were issued uniformly until 1956. Since 1959, air reconnaissance has furnished the ice cover data, essentially by visual observations. Supplementary reports on sea ice in Denmark Strait have been printed since 1953 in *Jökull*, sponsored by the National Research Council in Reykjavik. The Icelandic reports are compiled from sightings by coastal vessels and from aerial photographs. Air records have also been collected by the U. S. Naval Oceanographic Office since 1954.

IV. SOURCES AND PROCESSING OF DATA

Evaluation of the sundry factors influencing the ice cover has been made after gathering all such data as might contribute to the complete picture. The collection has been unevenly distributed, as more data can be found on some items than on others. Moreover, after preliminary investigation revealed precipitation and evaporation to be relatively insignificant, further efforts were directed primarily toward enriching the material pertaining to appraisal of the more dominant causes.

East Greenland Current data have been reconstructed from hydrographic material preserved in old expedition reports, notably from the *East Greenland*, *Belgica*, *Danmark*, *Godthaab*, *Tre Aars* and *Meteor* expeditions. These records in some cases contain original lists of stations and hydrocasts. As the original data were not compiled to serve the purpose of this study, only a modest part has been useable, but it sufficed to construct four hydrosections across the current at 74°–76° N.

Tracing of water movements around Iceland and in the Faeroe-Shetland channel was made possible largely through the cooperation of the staff at the Oceanographic Data Center in Washington, D.C., who supplied magnetic tape records of all hydrographic data in their files from Marsden Squares 217, 218, 220, 254 and 288. This yielded 61,542 temperature and salinity data from the upper 100 m at approximately 840 stations. These stations have not been individually identified. All of the station data were instead categorized by one-degree unit areas, each Marsden Square being subdivided into 100 such unit areas. Programming and data processing were accomplished in Oregon State University's computer center, which provided computation by unit areas, month and year of means, ranges and standard deviations. These parameters in turn permitted interpretation of the time dependence of flow intensities and location of high flow intensities.

Seasonal variations in Denmark Strait were computed from the Data Center's tapes, and additional information was obtained from the records of the *Dana* and *Øst* stations. Surface isotherms, isohalines and ice limits were drawn from the latter data group, and the positions at different times of the front separating Polar Water and Atlantic Water were determined.

A body of sea surface temperatures at Selvogsbanki on Iceland's south coast, 504 monthly values from the years 1895–1936, was obtained from a study by THOMSEN (1937), who in turn had compiled them from ships' logs in the Copenhagen files of the *International Association for the Exploration of the Sea.* The data have been used here to correlate means of the months February–May with the ice cover. This has been done by subtracting each annual February–May mean from the total February –May mean of the period 1900–36. The correlation coefficient between the ensuing anomaly and the ice cover was then computed.

In the Faeroe-Shetland channel data were obtained from a series of stations variously occupied by the research vessels *Dana*, *Explorer* and *Scotia.* The station values were used to compute manually by planimetry a time series of heat transport through those parts of the sections Nolsø-Flugga and Faeroe Bank-Butt of Lewis which straddle the meandering current. Anomalies by month and year were determined in 17 years of the 1927–52 period, and annual volumes of flow were inferred from the results. A correlation between flow and ice cover was then established.

The meteorological data derive from weather stations in or near the area of interest. Pressure distributions are determined from 5568 data summarized from 169,220 observations. They are used to find the trend of annual pressure averages at the stations where barometric pressure has been consistently recorded. The means of the first two decades of this century is compared with the means of the second two decades, and the differences are used to infer a displacement of the Icelandic low pressure cell.

Precipitation graphs are worked up from 6168 data, which are monthly values based on 187,832 observations of Danish, Icelandic and Norwegian origin. From the former data, annual anomalies are calculated for a series of stations in and around the Greenland Sea. The sum of these anomalies is deemed to be a value representative of the ice region, and its correlation with the ice cover exaimed, as is the effect of average wind force in the period 1914–37.

Air temperature work is based on 10,028 data, also monthly values derived from 220,920 observations. As in the case of precipitation, variation of annual temperature in the ice region is inferred from summation of anomalies calculated for selected stations. Spitzbergen's annual and January–March anomalies are calculated and graphed separately. Correlation coefficients between ice cover and temperature are determined by all three criteria.

The effect of wind on the formation and retention of the ice cover is sought by defining an index quantifying the influence in certain months each season when persistence of a particular component at a given station may affect ice formation. The basic material consists of 105,204 monthly data on wind direction covering 9,627,260 observations, and from 5928 monthly data on wind strength covering 1,808,040 observations. The index is established at each station by multiplying monthly wind force by the number of times the pertinent component was observed at that station. The correlations between the indices and the ice cover fluctuations are then calculated.

The search for supporting information has entailed review of a number of papers on the Arctic, which are listed in the bibliography. This has been done not as a process of judging contents on their intrinsic merits but rather for the sake of relating their contents and conclusions to the task at hand, a purpose for which none of them was intended.

V. CURRENT EFFECTS

The East Greenland Current

The surface waters of the Greenland Sea are formed by the confluence of water from the East Greenland Current through the Jan Mayen Current and the East Iceland Current with water from the North Atlantic Current, transported through the Irminger Current, the Norwegian Current and, northernmost, the West Spitzbergen Current, as shown in Figure 4. Most of the surface mixing takes place in two gyres, one located south of Jan Mayen and the other northeast of that island.

This current picture has been known in general outline since the first internationally coordinated survey was launched in 1902–04,[10]) but recent observational programs, particularly during the IGY, have refined and somewhat modified several of its details. The picture thus emerging reveals the current structure at any given time to be both complex and ephemeral in the area under discussion.

A dearth of information makes Western oceanographers speak with caution about the East Greenland Current, but TRESHNIKOV states the annual outflow between Greenland and Spitzbergen to be 162,000 km^3,[11]) or an average 5.1 Sv.[12]) In 1962, MOSBY summarized information from sundry sources in a general description of this current, but since then a profound and illuminating study has been published by AAGAARD (1966).

It appears that in the northern part of the East Greenland Current Polar Water is found from the surface to a depth somewhat greater than 150 m. The temperature is between 0°C and the freezing point, and the salinity varies linearly from about 30 ‰ at the surface to slightly more than 34 ‰ at the bottom of the layer. Underlying the PW is the Atlantic

[10]) This first attempt at a synoptic survey was planned at the International Oceanographic Congresses in Stockholm in 1899 and in Christiania in 1901.

[11]) Although TRESHNIKOV speaks unequivocally, he does not disclose the exact nature or quantity of observations, making only the statement: "During the last eleven years Soviet high-latitude air expeditions and drifting stations have thoroughly explored the Arctic Basin, at a wide network of deep-water oceanographic stations" (TRESHNIKOV, 1959).

[12]) The Sverdrup unit is used here, 1 Sv = $10^6 m^3/sec$.

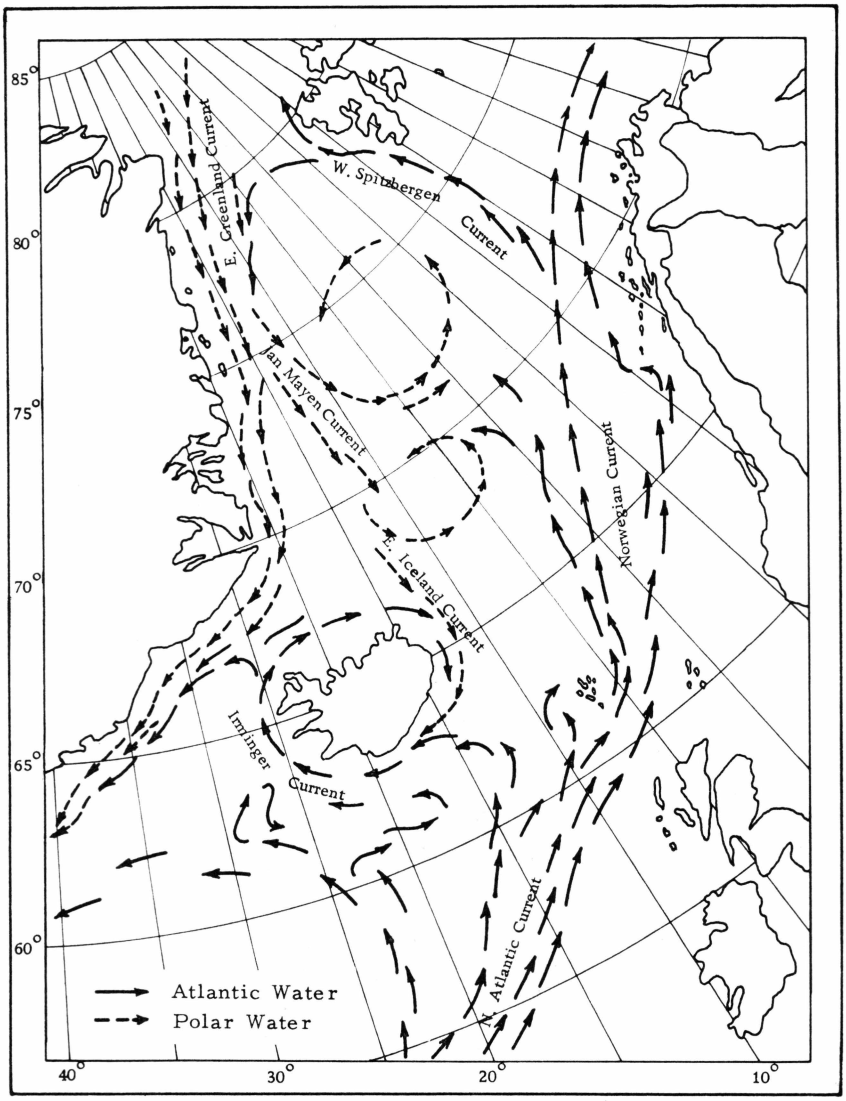


Figure 4. Surface currents in the Greenland Sea.

Intermediate Water, advected through the current paths mentioned above and extending to a mean depth of 800 m. The temperature of this water mass remains above 0°C with a maximum evident between 200 m and 400 m, while the salinity increases downward until it reaches a value between 34.88 $^0/_{00}$ and 35.00 $^0/_{00}$. Below 400 m the salinity remains virtually constant at this value. Below 800 m Deep Water is found, but for present purposes its characteristics are of no consequence.

Calculation of current speeds and volume transports has been made by AAGAARD (1966) who found typical speeds of 10–15 cm/sec. He further calculated the PW transport at 7.7 Sv, dominated by a barotropic mode, the AIW transport at 21.3 Sv and a deep water transport of 2.5 Sv. These figures add up to a rather startling 31.5 Sv total, which is an order of magnitude greater than previously believed,[13]) but they are well-supported.

Hydrographic data from the passage at Nordostrundingen are too scarce to give any indication as to fluctuations in the PW outflow. On the basis of heat budget considerations AAGAARD concluded that significant seasonal variations are likely.[14])

LEE (1962) has estimated West Spitzbergen Current transports at 74° N, using 11 years of hydrographic measurements, and found strong seasonal fluctuations in the upper 400 m with a maximum transport in February and a minimum in April. TIMOFEYEV (1962) has reported on 13 years of data across the West Spitzbergen Current at 78° N, finding a maximum in December–January and a minimum in May–June, the former being about twice that of the latter. These fluctuations may well contain a clue to East Greenland Current fluctuations, for several workers have claimed finding a coupling between the two streams; but unfortunately there is disagreement on whether or not the currents are in phase.[15])

The available hydrodata from the area between Nordostrundingen and Spitzbergen do not allow an evaluation to be made of year-to-year variations, but a fairly sizeable body of material exists from the records of past expeditions in the area from 74° to 76° N, and further south. The station lines are shown in Figure 5, and the current in these latitudes crosses obliquely the *Arlis II* station line which was oriented 210° T; it probably closely parallels the 100 m isobath, giving a current direction of ca. 235° T. In Figure 6 hydrodata gleaned from the records of the *Belgica* and *Danmark* expeditions have been used to project hydrographic sections onto a plane normal to the current off Shannon Island.

Measuring the areas enclosed by the $-1.0°$ isotherm shows the cross

[13]) It should be noted that AAGAARD's results exceed as stated such Western investigators as VOWINCKEL and ORVIG (1962), who have estimated a total transport of 3.4–3.7 Sv. The PW far exceeds MOSBY's estimate (1962) of about 2 Sv, but is fairly close to the above mentioned Russian figure of 5.1 Sv (TRESHNIKOV, 1959).

[14] CHAPLYGIN (1960) has stated that the East Greenland Current is of highest intensity in winter due to prevailing strong northerly winds, but he offers no evidence. LAKTINOV, SHAMONTEV and YANES (1960) are of the same view but also fail to provide substantiation.

[15]) LAKTINOV *et al.* (1960) believe the currents to be out of phase, counteracting each other, whereas TIMOFEYEV (1962) has concluded from ice drift observations that they are in phase.

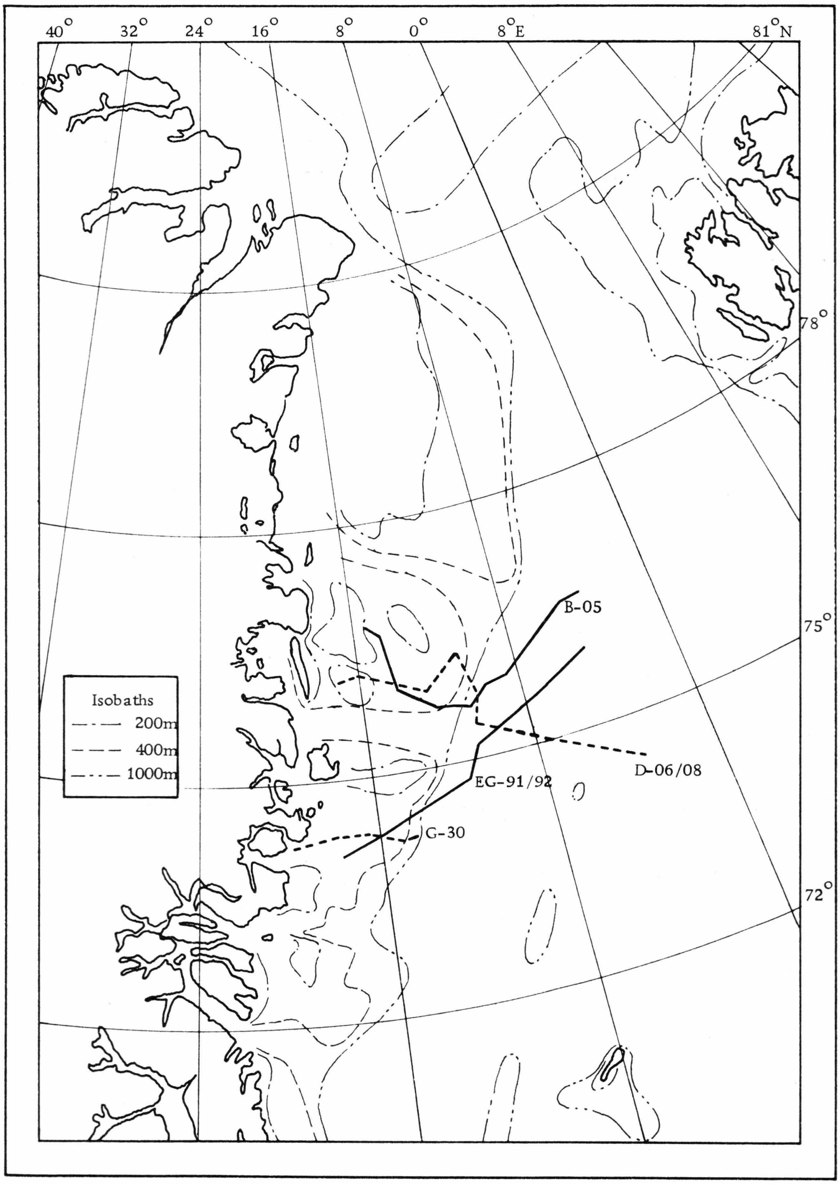


Figure 5. Location of station lines B-05, *Belgica* expedition 1905; D-06/08, *Danmark* expedition 1906–08; G-30, *Godthaab* expedition 1930; and EG-91/92, E. *Greenland* expedition 1891–92.

section at the time of Belgica's work in 1905 to be 53 $^0/_0$ larger than was the case in 1906–08, when the *Danmark* expedition made its hydrocasts. The former was in a period of increasing ice cover, the latter in a sequence of years of reversing trend, see Figure 3.

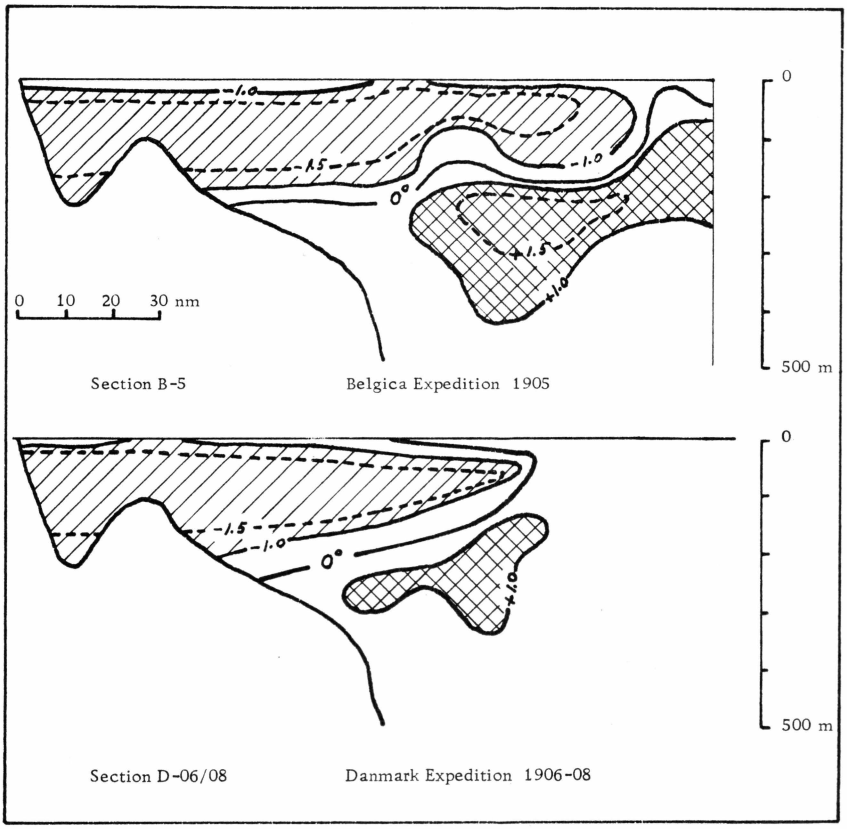


Figure 6. Positions of isotherms in July 1905 and August 1906–08. Sections are in same location, normal to East Greenland Current at 75°–76° N, projected from *Belgica* and *Danmark* expedition data.

Another telling comparison can be made between the *Godthaab* section of July 1930 and the *East Greenland* expedition section of August 1891, Figure 7. These sections are in nearly the same location, in fact intersect each other at about 16° W. The difference in size of the PW flows is striking, as is the difference in ice severity: 1930 was one of the lightest ice years on record, whereas MEINARDUS (1906) reported that 1891 was one of only four years in the 19th century when the ice arrived at Iceland before Christmas.

It is also noticeable that the quantity of water of temperatures $>1.0^{\circ}$C is greater when the PW flow is strong. This seems to indicate a general strengthening of the current with more entrained AIW eddying off the West Spitzbergen Current. The significance of this will be discussed later.

An item of interest is presented by a comparison drawn by JAKHELLIN (1936) between East Greenland Current volumes in 1905 and in

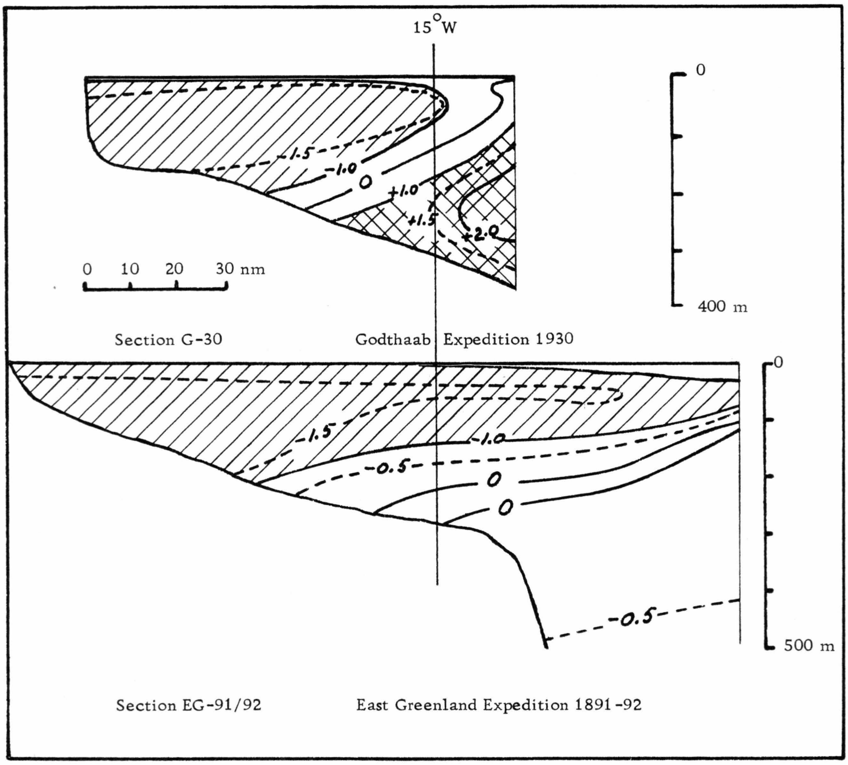


Figure 7. Positions of isotherms in August 1891 and July 1930. Sections are located about 74° N, based on *Godthaab* and *East Greenland* expedition data.

1931–32. JAKHELLIN calculated the former to be 1.6 Sv and the 1931–32 average to be 1.32 Sv. It will be seen from Figure 3 that the ice cover in those years deviated from the 1898–1913 mean by +7.2 and −14 ($\times 10^4$ km^2), respectively, showing agreement with the change in volume.

The *Arlis II* current measurements show that, generally, no depth of negligible horizontal motion is found in the East Greenland Current. Although the hydro sections therefore are not translatable into meaningful velocities, it is clear from the foregoing that year-to-year variations in the East Greenland Current do exist.[16]) That they are directly linked to the ice occurrence is also empirically indicated.

[16]) It is worth noting that the *Hansa* survivors in 1870 drifted from 75° N to Kap Farvel at an average rate of four miles per day, whereas IRMINGER reported that the same route was covered by shipwrecked whalers in 1777 at a progress of 11–12 miles per day. Finally, in 1937–38 the ice station NP-1 traveled from 82.8° N, 7.0° W to 70.7° N, 19.3° W, at a net progress of 3.3 miles daily. It is open to speculation whether the differences in drift speeds are due to the different tracks traveled, or to current velocity differences in different years; but these case histories merit recall, as they represent direct measurements.

The North Atlantic Current

After leaving the vicinity of the American coast, the North Atlantic Current delineates the Polar Front, variable in time and space except at 52° N, 30° W, where it seems permanently fixed over the breakthroughs in the Mid-Atlantic Ridge. Subtropical water, coursing through the current's fingers, maintain secondary fronts in the Norwegian and Greenland Seas and in Denmark Strait. This water of about 8° C and 35.3 ‰ is a dominant factor in the ice picture, and it is of interest to assess its flows.

The North Atlantic Current's potential influence may be placed in proper perspective by comparing the magnitude of a 20 % variation in the heat flowing through the Faeroe-Shetland channel in a given month with the heat needed to melt a season's ice in the Greenland Sea. The former can be approximated by using a value of 2.3 Sv[17]) at 10° C, giving a value of 4.6×10^{12} cal/sec of such a fluctuation, or 1.2×10^{19} cal/month. The latter can be roughly estimated as the heat needed to melt 500,000 km^2 of ice one meter thick, or 4×10^{19} cal. If the current and with it the heat import in fact do fluctuate from year to year, it is obvious that an influence on the melting of the ice cover must be detectable.

In the critical Iceland-Faroe-Shetlands area, where it gains entrance to the Norwegian and Greenland Seas, the current has been the object of numerous explorations, most recently during the IGY and in the subsequent ICES investigation. Sufficient data have thus been collected on the current to gain several synoptic glimpses of its vagaries. As the core from examination of 68 hydrosections appears centered at about 100 m depth, this level will be analyzed.

It will be assumed that a cross-current temperature gradient exists wherever a major ocean current is located, so that a horizontal gradient will be a both necessary and suficient reason to infer a current's existence. FUGLISTER (1954) proposed this reasoning as a basis for current analysis, and it has been applied in the following. Between 60° and 65° N in the Atlantic, seasonal temperature variations by one-degree latitude belts have been computed by planimetry from Hydrographic Office charts of isotherms at the 100 m depth. The variations average 0.69° C in the five one-degree belts from 60° to 65° N, showing a temperature minimum in February and a maximum in August.

For purposes of comparison with this North Atlantic average, the seasonal variation in four-degree units straddling the Faeroe-Shetland channel has been computed. The area comprising the four units heavily outlined in Figure 9, and the area's mean monthly temperatures have

[17]) This is a value estimated by MOSBY (1962).

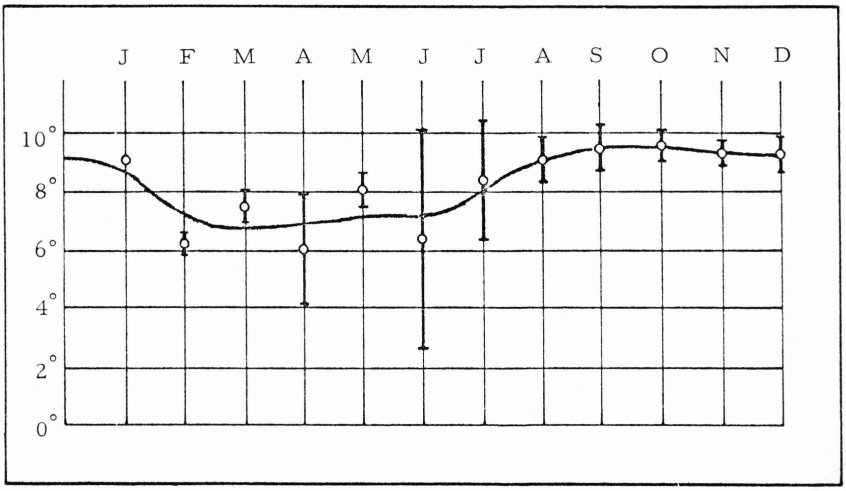


Figure 8. Monthly temperature averages plus/minus one standard deviation for the area 60°–61° N, 3°–5° W and 61°–62° N, 5°–6° W (framed in red in Figure 9).

been plotted in Figure 8 together with their standard deviations. A curve representing the variations has been drawn, and it can be seen that the seasonal variation in this area is almost ±2°C vs. the 0.69°C North Atlantic average for the same latitude. If temperature is equated with current strength, the explanation can be sought in the different current intensities prevalent in different seasons: fall-winter flows exceed spring-summer flows.

Petterson descriptively applied the word "Anschwellung" to the June–September intensities when analyzing data from the turn of the century,[18]) and the phenomen is reliable, as shown by the small standard deviations in the autumn months.

Certain longer term variations have been found in the surface layer by Brown and Rodewald (1953), who determined decreases of 0.3°–0.7°C in this area from 1910 to 1950. Rodewald (1966) also discovered cyclic surface temperature variations of the order of several years, but in the area in question they are well below 0.5°C. A sizeable temperature range in a small area must therefore be due to the presence of a current-induced horizontal gradient in that area, including the effect of current meander.

The material used is from 4115 hydrographic station casts, everything the Oceanographic Data Center in Washington has on file from Marsden Squares 217 and 218. Using temperatures at the 100 m depth,

[18]) "Zwischen Ende Mai und Mitte September hatte eine mächtige Anschwellung des atlantischen Wassers stattgefunden, begleitet von einer Zunahme des Salzgehalts und der Temperatur" (Petterson, 1929).

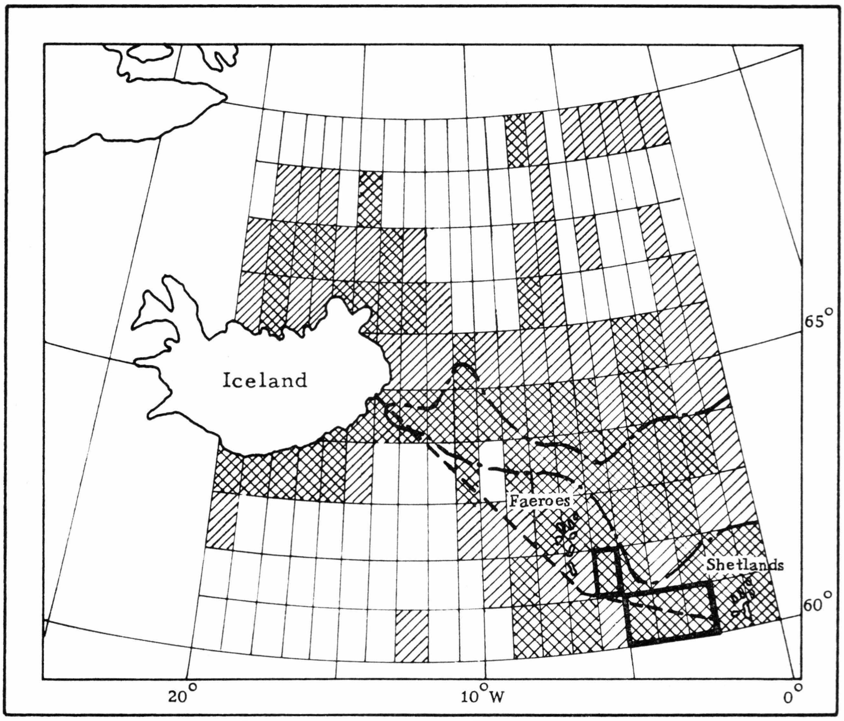


Figure 9. Temperature ranges in Marsden Squares #217 and #218. Double-hatched area greater than 5°, single-hatched area > 2°–5° range. Area of max currents framed heavily. Area enclosed by (— — —) is Fuglister's 5° range. Hydrosection line is shown as (— — —).

the ranges of observed temperatures have been computed by one-degree unit areas such as 60°00′–60°59′ N, 0°00′–0°59′ W. Each range has been found by subtracting the minimum observed temperature from the maximum observed, regardless of date of observation.

In Figure 9 the range 2°–5° C has been single-hatched, and the range > 5° C is double-hatched. FUGLISTER's outline of the same > 5° C range is also shown, and it will be noticed that it is somewhat smaller. The reason must be primarily that he looked at the 200 m depth, whereas the core of the flow as previously mentioned appears centered at the 100 m level.

In order to locate more precisely the currents traversing the section Iceland-Faroes-Shetlands shown by a heavy stipled line in Figure 9, the mean temperature, the temperature range and the temperature's standard deviation at the 100 m level have been computed for each of the one-degree unit areas the section traverses from Iceland to the Shetlands.

Plotted in Figure 10, the average temperature curve shows that the maximum gradients lie between unit areas centered on 14°30′, 13°30′

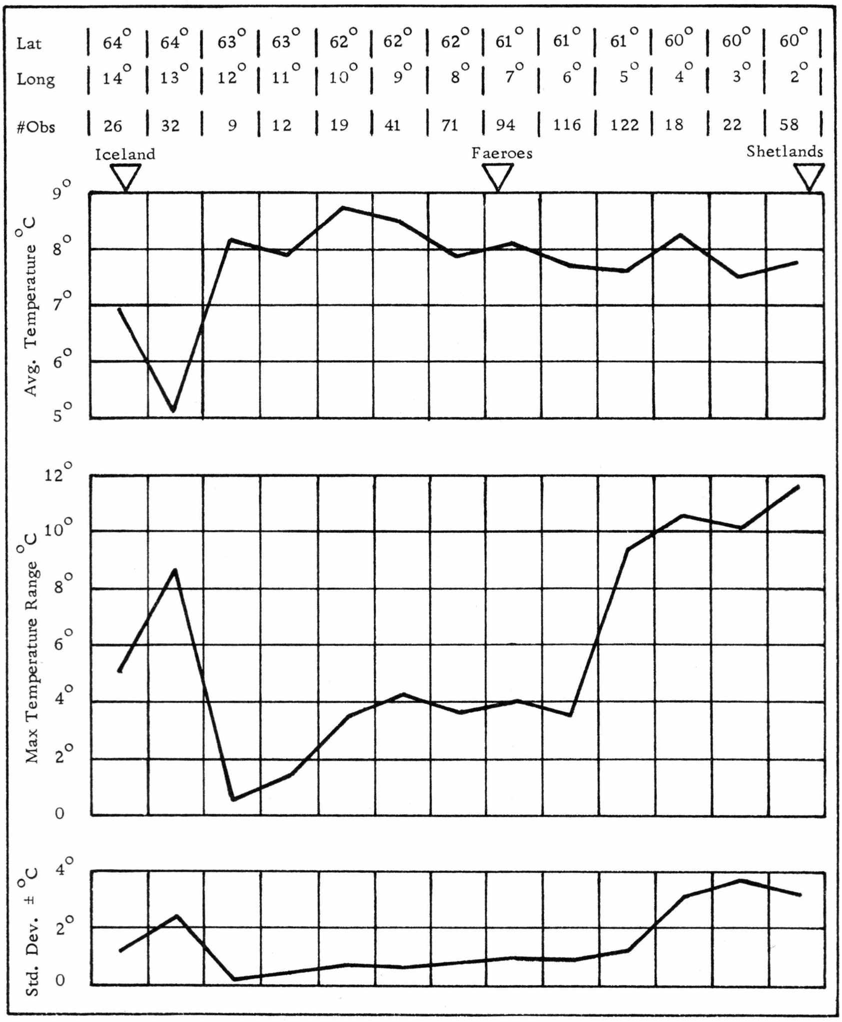


Figure 10. Average temperature, maximum range of temperature, and standard deviation by one-degree unit areas covering Iceland-Faeroe-Shetland section, as shown in Figure 8.

and 12°30′W; 11°30′ and 10°30′W; and 5°30′ and 4°30′W. Standard deviations are highest in the unit areas centered on 13°W and 3°W, indicating major currents to be located at these points. Perusal of the charts in Figures 4 and 9 reveals the current at 13°W to be the Irminger Current, whereas the North Atlantic Current is located between 5° and 2°W. The range curve shows high values narrowly peaked at 13°W, meaning that the Irminger Current is fixed in position here. One the other hand, high range values throughout the area 5° to 2°W show the North Atlantic Current to assert itself over this entire width, a conclusion which is further supported by the matching high standard deviations.

Knowing the location of the North Atlantic Current, one can evaluate its relative strengths in the years from which data exist, and also the relative magnitude of the heat carried by it into the Norwegian-Greenland Sea region. The current is an upper water phenomenon, defined in depth by the 550 m Wyville Thomson Ridge and breaking surface at times over a wide area. For present purposes estimates have been made of the heat transported in a cross section 100 m deep, 182 km wide and one centimeter thick. The upper 100 m layer may reasonably be expected to influence ice conditions. The 182 km width represents the distance from 2°30′ to 5°30′W along the section line in Figure 9, which the high range values in Figure 10 showed to be the location of maximum flow.

Analyses have been made of 50 sections deriving from *Explorer*, *Scotia* and *Dana* data from the period 1927–52. The reasoning has been applied that a strong current through the Faeroe-Shetland channel will manifest itself in a positive temperature anomaly, *i.e.* the temperature at any point in the stream will be higher than the North Atlantic mean for that latitude. For latitude belts 60°–61°N and 61°–62°N the average temperatures as computed from Hydrographic Office charts of 100 m isotherms are 6.65° and 6.54°C, respectively. Adding about 2°C to allow for seasonal variations as determined in Figure 8, a value between 8° and 9°C results. It can accordingly be stated that the 9° isotherm in the section under study must enclose an area of substantial flow.

In each of the 50 sections, 100 m deep, 182 km wide and one centimeter thick, the areas enclosed by the 9° and higher isotherms, where present, have been measured planimetrically and their heat contents calculated and summed. The results are listed in Table 1, zero values denoting cases where all temperatures were <9°C. The figures in parentheses below the heat values are the anomalies by month. Thus for example the mean of all July values has been subtracted from each July value, showing that in 1934 the section's heat content was 42 units below the average for the month. When contemplating Table 1, it should be borne in mind, that a high heat value indicates both the presence in the section of an above average heat content available for transport *and* that a stronger-than-average current may be inferred. The effect on heat import of a high heat value in the section is exponential.

Without reading more into the figures than is reasonably warranted, it may be considered significant that the two largest monthly anomalies are in May–June of 1929, a year when the curve in Figure 3 was plunging toward one of the lightest ice years on record. If the years are classified

Table 1. Heat content in cals $\times 10^{10}$ enclosed by the 9° isotherm in the section, 182 km wide, 100 m deep and 1 cm thick traversing the peak current intensy in the Faeroe-Shetland channel. Monthly anomalies are listed in parentheses.

	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
1927....							211 (−17)				
1929....				224 (+159)	240 (+64)						
1931....				68 (+3)							
1933....					185 (+9)						
1934....				59 (−6)		139 (−42)					
1935....					192 (+16)						
1936....					161 147 (−22)						
1937....				48 100 (+9)							
1938....				45 (−20)	165 (−11)	210 (+29)					
1939....					213 140 (0)	220 (+39)					
1946....							280 (+52)				
1947....					263 153 (+32)	188 (+7)	235				
1948....				0 116 (−7)				220 (+1)			
1949....		0 (−54)		0 (−65)				227 (+8)		121 (−48)	
1950....		161 (+107)			162 (−14)	200 (+19)				161 (−8)	
1951....				0 (−65)	130 (−46)	104 (−77)	185 (−43)	211 (−8)	311		195
1952....	0	0 (−54)		0 100 82 (+26)	138 (−38)	193 198 (+15)				153 240 (+37)	

into four categories of rising, peak, declining and low ice coverage, summation of the monthly anomalies in these categories appears as listed in Table 2. Since the peaks and lows are more pronounced in the years 1927–39, these have been summed separately, but in either case negative totals accrue in the first two categories, positive in the last two. It is interesting that the declining category is more strongly positive than is the low one. It may be explained by the delay of the warm water's arrival at the ice edge. At an average progress of 10 cm/sec, the trip from the channel to the gyre north of Jan Mayen would take just six months, so that the effect would be evident largely in the succeeding season.

Turning for a moment to the salinity of the current, a paper by SMED (1943) deserves attention. From an analysis of the International Council's records from 1900 to 1940, SMED found a definite increase in surface salinity in several areas of the North Atlantic. For the crucial area between Iceland and Scotland, his results are shown in Figure 11. From 1902/17 to 1919/39, the increase amounted to 0.06 ‰. Except as a possible indicator of intensity of the flow of Atlantic water, this salinity change is of little direct consequence in the ice picture; in this case, it lowers the freezing point by a mere 0.003° C. It is nevertheless worth noting that a minimum winter salinity occurred in 1918, the severest ice season of all the years under study here, whereas winter salinity had progressed to a maximum in 1930, an exceptionally light ice year. SMED felt that the onset of the salinity increase began about 1920,[19]) and if the graph in Figures 3 and 13 are viewed with this in mind, the coincident decrease in ice cover is striking. Averaging over SMED's periods, the recession of the ice can be computed as $1.3 \times 10^4 \text{km}^2$ from 1920/17 to 1910/39.

Table 2. Summation of monthly anomalies in years with ice severity classifiable as rising, peak, declining or low.

	1927–1939	1927–1952
Rising	−14	−20
Peaks.	−50	−258
Declining	+239	+215
Lows	+26	+180

[19]) SMED stated: "The investigations . . . all point towards the probability of the increase in salinity having commenced about 1920."

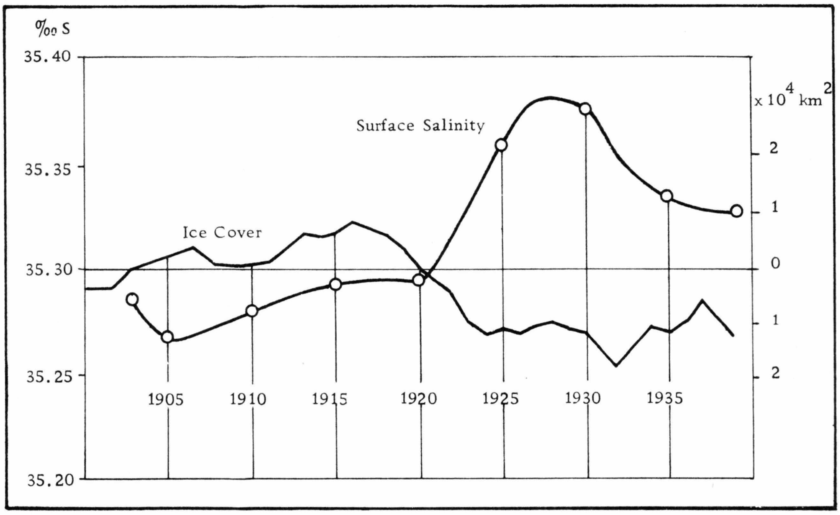


Figure 11. Average annual surface salinity of the area 58°–63° N, 10°–15° W.

The Irminger Current

Although the Irminger Current was discovered and described before any other in the Greenland Sea, data from which its behavior may be reconstructed are scanty. During the IGY investigations, Denmark Strait and Irminger Sea charts of geopotential surface topography were produced, revealing rather intricate flow patterns including a large cyclonic whorl southwest of Iceland and a bifurcation to the northwest. From the latter, one branch curves north around Iceland, while another probes westward to join the East Greenland Current. Together the two currents wind their way west and south, traceable by a band of crowded isotherms as shown in Figure 12, which is constructed from *Anton Dohrn* data from the summer of 1955. The 500 m and 1000 m isobaths are also shown, as they mark the slope of a steep-sided trough occupying the central part of the strait, with a 650 m sill depth west of Iceland's Cape Horn. The mean ice limits for the months of April, July and August have been superimposed.

In the years 1929, 1931 and 1932 detailed ice information was collected in Denmark Strait by *Øst* and *Dana*, and a comparison can therefore be made of the ice borders in the months of July of those years. In Figure 13 the three borders have been portrayed, and it is evident that certain features are of a semi-permanent character. It is apparent that the ice border coincides with the position of the Polar Front and that hydrographic conditions creating the front in turn depend on local

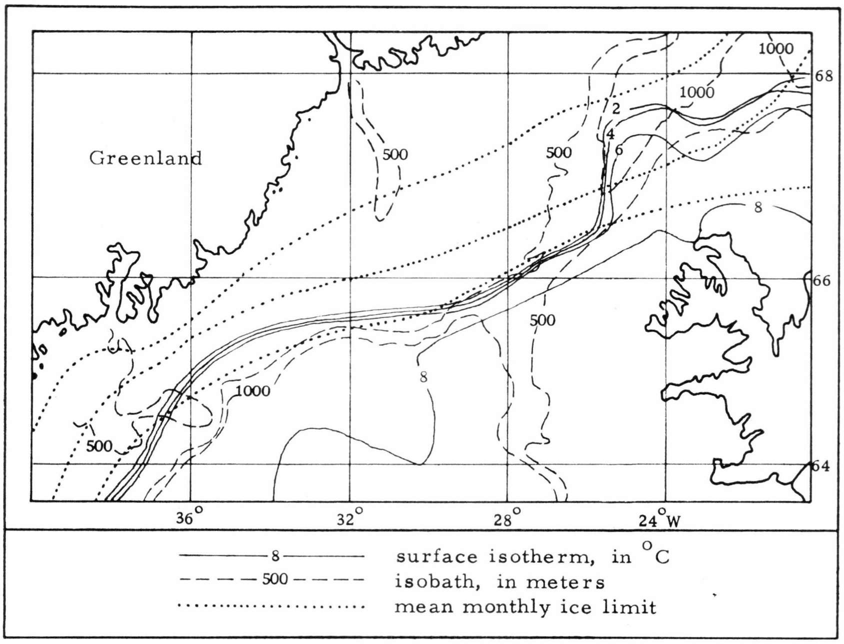


Figure 12. Surface isotherms (blue) delineating boundary between Atlantic Water of Irminger Current and Polar Water of East Greenland Current in Denmark Strait, based on *Anton Dohrn* data from June 1955.

bathymmetry. The precise bathymmetric charts produced in the 1950s confirm this.[20])

The current-ice relationship in the strait can accordingly be described as follows. The Irminger Current's bifurcation point is marked by a northward bulge of the 33 ‰ and 34 ‰ isohalines and the 1°–6° isotherms, which carve a matching bay into the edge of the ice. The point is displaced in different years (see Figures 12 and 13), along a line crossing the sill in a NE–SW direction, and meanders are spawned from it as the western branch of the Irminger Current proceeds together with the East Greenland Current toward the southwest. The meanders pendulate about an axis lying near the edge of the continental shelf and move down-stream at about 1/10th the current velocity. In the process they mold the ice border like the teeth of a giant saw.

It is interesting that the ice recedes in successive summer months to lines which remain parallel to the shelf edge. This takes place under the combined influence of diminishing supply of ice from the north and the July–August swelling of the Atlantic Water flow. Existence of this annual variation has been verified in Figure 14 by using monthly mean

[20]) THOMSEN (1938) drew similar conclusions based on successive chartings of the ice border from June through August in 1929.

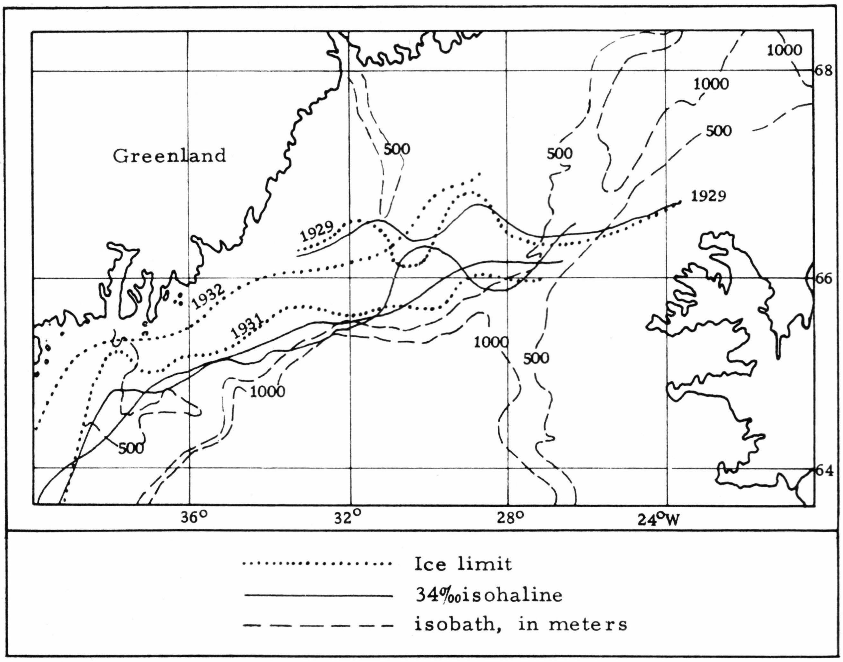


Figure 13. July ice limits and mean positions of 34 ‰ isohalines in the years 1929, 1931 and 1932.

temperatures from selected unit areas. Their differences from the annual mean provide the points of the curve, and the autumn temperature rise indicates the same characteristic swelling of the flow as was seen to exist in the Faeroe-Shetland section.

Year-to-year variations in the position of the ice border in Denmark Strait can then be attributed at least in some measure to the Atlantic Water flow's attaining different strengths in different years. Proof of

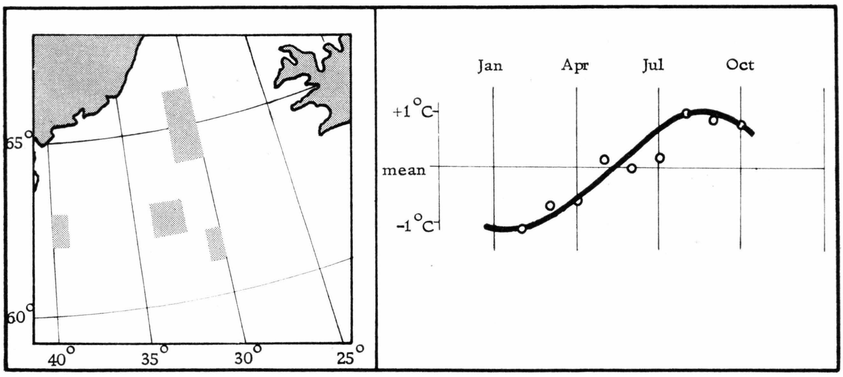


Figure 14. Seasonal variation of temperature at 100 m depth from annual mean in the eight degree unit areas shown.

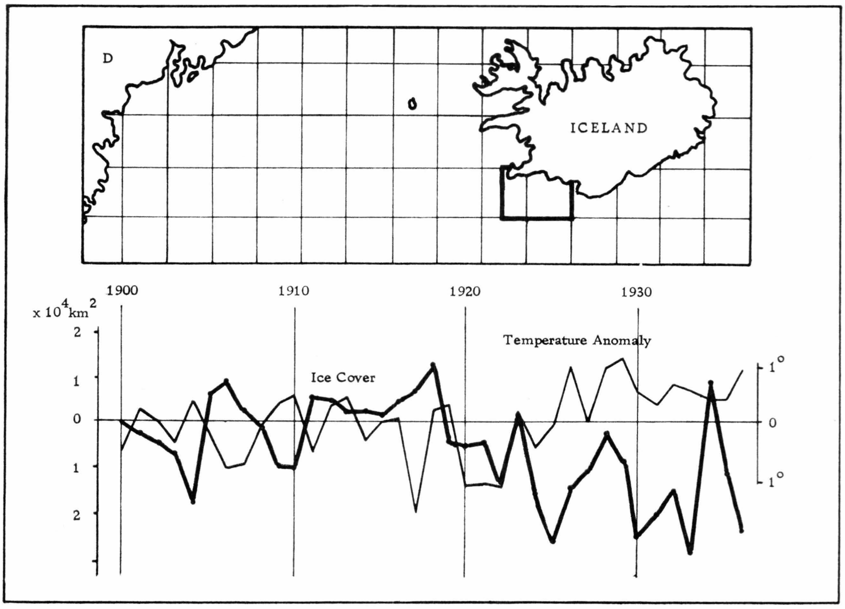


Figure 15. February–May surface temperature anomaly at Selvogsbanki and ice cover curve (from Figure 3).

this is almost totally lacking, as the current in the Irminger Sea generally is very broad and sluggish, not at all amenable to the kind of analysis that was possible on the Faeroe-Shetland channel flow. An attempt at a possible correlation can nevertheless be made by looking at the surface temperatures, the one type of data which is on hand in sufficient quantity.

In a study of long-period trends, THOMSEN (1937) compiled surface temperatures at Selvogsbanki in the area framed in Figure 15 and obtained monthly means for the years 1895–1936. Drawing upon this data series, the mean of the February-May temperature anomalies is graphed in Figure 15 together with the ice cover curve for the same period. The correlation coefficient has been found to be -0.30, not an impressive value but with the proper sign. It is here assumed that Selvogsbanki surface water will reach the ice edge through the Irminger Current with a time lag of one month.

It is probably more illustrative to analyze the years in which the ice cover differed significantly from its mean value. This has been done in Table 3, where the surface temperature anomalies have been summed for those years in which ice cover deviated from its regression line by an amount greater than s. In this manner temperature anomalies totaling $-2.68°$C accrue in six years of high (positive) ice cover values, whereas temperature anomalies totaling $+2.55°$C accrue in six years distinguished by low ice cover values. An inverse relationship thus seems to exist between intensity of the Irminger Current as indicated by Selvogsbanki surface temperatures on the one hand, and extent of the ice cover on the other.

Table 3. Tabulation of years in which ice cover (x) deviated from its mean by a value larger than s (Figure 3). The February-May surface tempersture anomaly is listed under each year.

Years of $x < +s$	1906	1907	1917	1918	1934		Total
Temperature Anomalies, °C	−.83	−.78	−1.60	+.15	+.38		−2.68
Years of $x < -s$	**1904**	**1925**	**1930**	**1931**	**1933**	**1936**	**Total**
Temperature Anomalies, °C	+.35	−.10	+.50	+.30	+.53	+.97	+2.55

VI. CLIMATIC EFFECTS

Relevant Weather Stations

Testing the influence of climatic factors is aided by a body of weather information which is rather sizeable—at least if compared with Arctic oceanographic observations. In Fig. 16 the stations are shown which have provided the basic data used in the following. Their operational records vary a great deal, Thorshavn, Lerwick and the Norwegian mainland stations being the only ones with practically unbroken records in this century. Observations from points in Iceland are fragmentary, and gaps also exist in those from Angmagssalik.

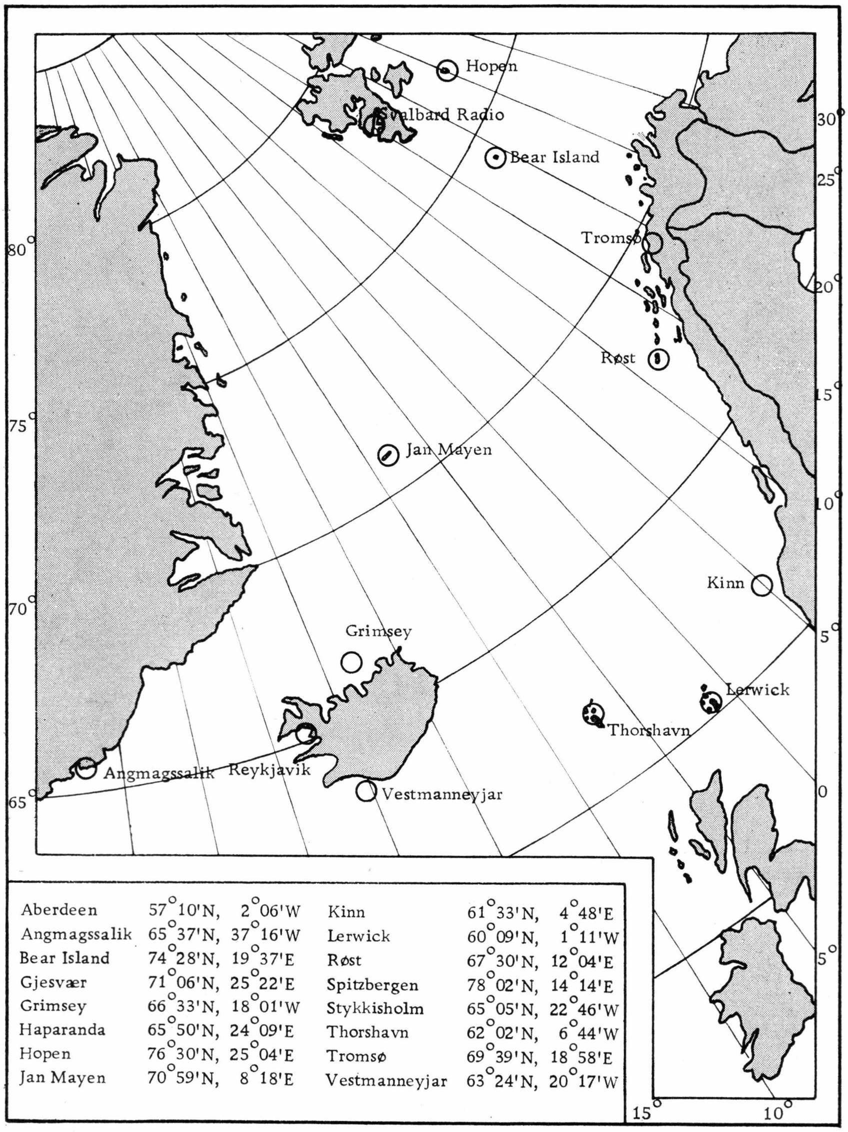
Spitzbergen was equipped with a complete weather station in 1914,[21]) Jan Mayen in 1922, Bear Island in 1928 and Hopen not until 1946. Their records have been reliably continuous except for the WW-2 period.

Some stations have been moved slightly in the period under review. The Tromsø station was in 1927 moved to the town's *Geofysiske Institut*, changing altitude from 38 meters to 102 meters. Røst data prior to 1907 derive from Skomvaer, some eight miles SE of the present location. The Spitzbergen station was in 1930 moved to Svalbard Radio from Green Harbour, which is 23 miles NE of the present spot. None of the moves should have any measurable effect on the comparative value of the observations as they are used here. Preference is given, where possible, to island stations well off the nearest mainland, so that data sources on the fringe of the area of interest represent oceanic rather than terrestrial climates.

Weyl's Precipitation Hypothesis

At this point, it is pertinent to revert momentarily to the hypothesis proposed by Weyl (1966), to wit, that severe ice conditions must be expected after quiet, rainy summer seasons. Angmagssalik, Grimsey, Jan Mayen, Spitzbergen, Bear Island and Tromsø should together give a good indication of rainfall in the Greenland Sea. In Figure 17 the precipitation anomalies from these points have been summed by year for

[21]) A number of observational series mostly of 1- or 2-year duration were obtained from various localities on the island, the earliest dating back to 1894.



Aberdeen	57°10'N, 2°06'W	Kinn	61°33'N, 4°48'E
Angmagssalik	65°37'N, 37°16'W	Lerwick	60°09'N, 1°11'W
Bear Island	74°28'N, 19°37'E	Røst	67°30'N, 12°04'E
Gjesvær	71°06'N, 25°22'E	Spitzbergen	78°02'N, 14°14'E
Grimsey	66°33'N, 18°01'W	Stykkisholm	65°05'N, 22°46'W
Haparanda	65°50'N, 24°09'E	Thorshavn	62°02'N, 6°44'W
Hopen	76°30'N, 25°04'E	Tromsø	69°39'N, 18°58'E
Jan Mayen	70°59'N, 8°18'E	Vestmanneyjar	63°24'N, 20°17'W

Figure 16. Locations of weather stations from which data were used in this study.

the period 1900–39. The curve is assumed to be composed of four station values; when fewer station values are available in a particular year, the sum has been factored up; when more, down.

It is apparent that the curve correlates poorly with the ice cover curve shown above it. If the correlation coefficient

$$r = \frac{SP}{(SS_x SS_y)^{1/2}}$$

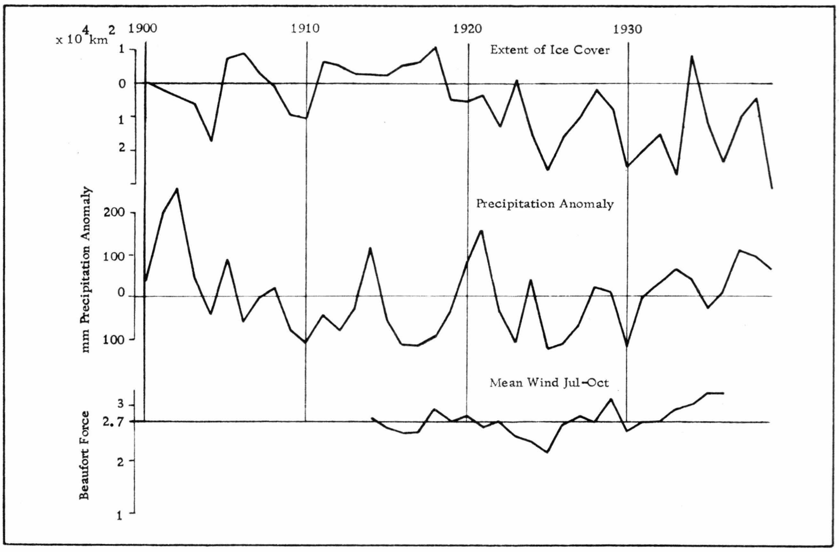


Figure 17. Precipitation anomalies of Angmagssalik, Grimsey, Jan Mayen, Spitzbergen, Bear Island and Tromsø summed by year and juxtaposed with ice cover graph (from Figure 3). Below is graph of mean wind for the period July–October from the same stations.

is computed with x as the ice cover's deviation from the 1898–1913 mean in km² and y as the cumulative precipitation anomaly in mm as graphed, it is found that $r = 0.08$. Hence precipitation *per se* is of no significiant consequence.

The wind strength will now be taken into account as suggested by Weyl by summing from the same stations the mean winds each year for the period July–October, i.e. the four months preceding onset of ice formation. This curve is also graphed in Figure 17, but only from 1914, when the Spitzbergen station started operating, until 1936, when the Norwegian meteorological service ceased reporting wind force and switched to a different reporting criterion.

It is noticeable first of all that the winds are remarkably even in force from year to year, so that one would hesitate to classify any year as decidedly quiet. Secondly, the combination of precipitation and comparative calm is almost non-existent: wind and rain go together. If nevertheless all of the years showing below-average wind force are accepted as calm, it can be seen that two of those years experienced above-average precipitation, namely 1921 and 1924. The effects should have been felt in 1922 and 1925, respectively, but the ice cover was below average in 1922 and very low in 1925.

It is perhaps conceivable that even below-average annual rainfall could suffice for rapid ice formation, if only the fresh water layer were left undisturbed on the surface. In other words, storm activity, or the absence thereof, might be the only governing factor in regard to the precipitation effect. With this in mind the correlation coefficient between wind force and ice cover as portrayed in Figure 17 has been calculated. It turns out that for the 1914–36 period $r = 0.24$, an insignificantly small coefficient, and anyway positive, which is contrary to the requirement of Weyl's hypothesis.

All hope of forecasting ice seasons by the handy means of rain and wind reports should thus regrettably be abandoned. Although the effect of course must exist, as pointed out by WEYL, it is too slight to influence the ice regime in a detectable manner.

Evaporation

Despite scarcity and unreliability of surface evaporation figures from high latitudes, it is possible to establish that variations in heat withdrawal from the ocean surface due to varying evaporation in successive years can exert but negligible influence on the ice cover. The area from which evaporative cooling could be brought to bear is at most 800,000 km^2. Assuming 60 cm of annual evaporation, a total of 4×10^{19} cal would be a maximum heat withdrawal.

If a 10 % variation in this evaporative heat loss were to occur in a given year, the amount would be two orders of magnitude below the annual heat import through the Faeroe-Shetland channel as computed earlier.

Change in surface salinity from a 10 % variation in evaporation would amount to 0.2 ‰, changing the freezing point by 0.01 °C, if the surface water is uniformly mixed to a depth of 10 m. The change in heat import needed to melt 10^5 km^2 of ice one meter thick would be only 10^{15} cal, an amount six orders of magnitude below the annual heat import through the Faeroe-Shetland channel. It would seem safe to disregard evaporation as a factor for present purposes.

Temperature Variations

The gradual climatic warming in northern Europe and the Arctic in the 19th and 20th centuries has received much attention (ÅNGSTRÖM, 1939; SCHERHAG, 1939; HESSELBERG, 1940; ERIKSSON, 1943; HOVMØLLER, 1947), and a strengthening of the atmospheric circulation is generally surmised to be the cause. The trend is particularly pronounced in the Greenland Sea. HESSELBERG (1940) has called attention to the temperature increase at Spitzbergen, where the annual mean has gained 2.5 °C from 1899/1908 to 1929/38. Breaking down the increase, it is found that

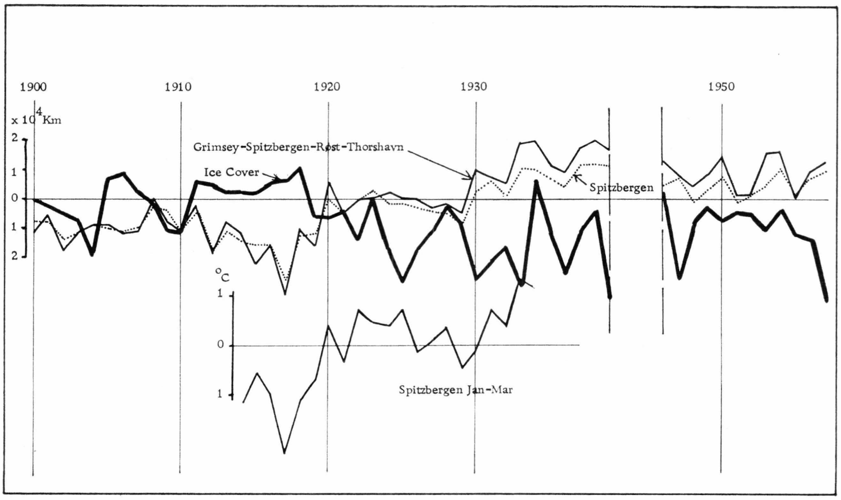


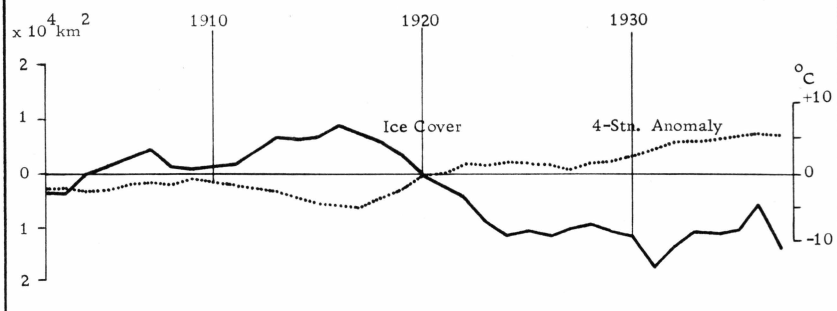
Figure 18. Air temperature anomalies of Spitzbergen and of Grimsey-Spitzbergen-Røst-Thorshavn combined. Spitzbergen's January–March air temperature anomaly is shown separately below.

the temperature has gone up by 5.0° C in the winter and only by 0.2° C in the summer, a circumstance HESSELBERG ascribes to heat consumption by melting of ice in summer, which keeps the air temperature near the freezing point.

In regard to the shorter term fluctuations, it is reasonable to expect air temperatures in the Greenland-Norwegian Sea areas to be strongly dependent on the ice regime at any time. In Figure 18a curve has been drawn of the combined temperature anomaly of four stations in the area, so that the air temperature's relation to the ice cover may be observed. The stations are Grimsey, Spitzbergen, Røst and Thorshavn, of which the first two are in the fringe of the ice, whereas Røst and Thorshavn are separated from it by 500–600 miles of open water.

The correlation coefficient between the combined anomaly curve and the ice is -0.512. The Grimsey-Spitzbergen anomaly has been drawn separately, and as one would expect its correlation is somewhat better at -0.682.

The question presents itself whether the correlation varies seasonally, and a 20-year series of monthly temperature means from Spitzbergen has been used to provide a partial answer. It appears that the January–March anomaly relates closely to the ice cover, as the coefficient jumps significantly to -0.818. On the other hand, calculation of Spitzbergen's May–August means and anomalies turns up a coefficient of only -0.292 when related to the ice cover. Both curves are shown in Figure 18. This casts some doubt upon HESSELBERG's argument referred



Temperature anomaly vs. previous year's ice	r = -0.816
Temperature anomaly vs. same year's ice	r = -0.775
Temperature anomaly vs. subsequent year's ice	r = -0.739

Figure 19. 5-year moving averages of ice cover curve (Figure 3) and annual combined temperature anomaly of stations Grimsey, Spitzbergen, Røst and Thorshavn.

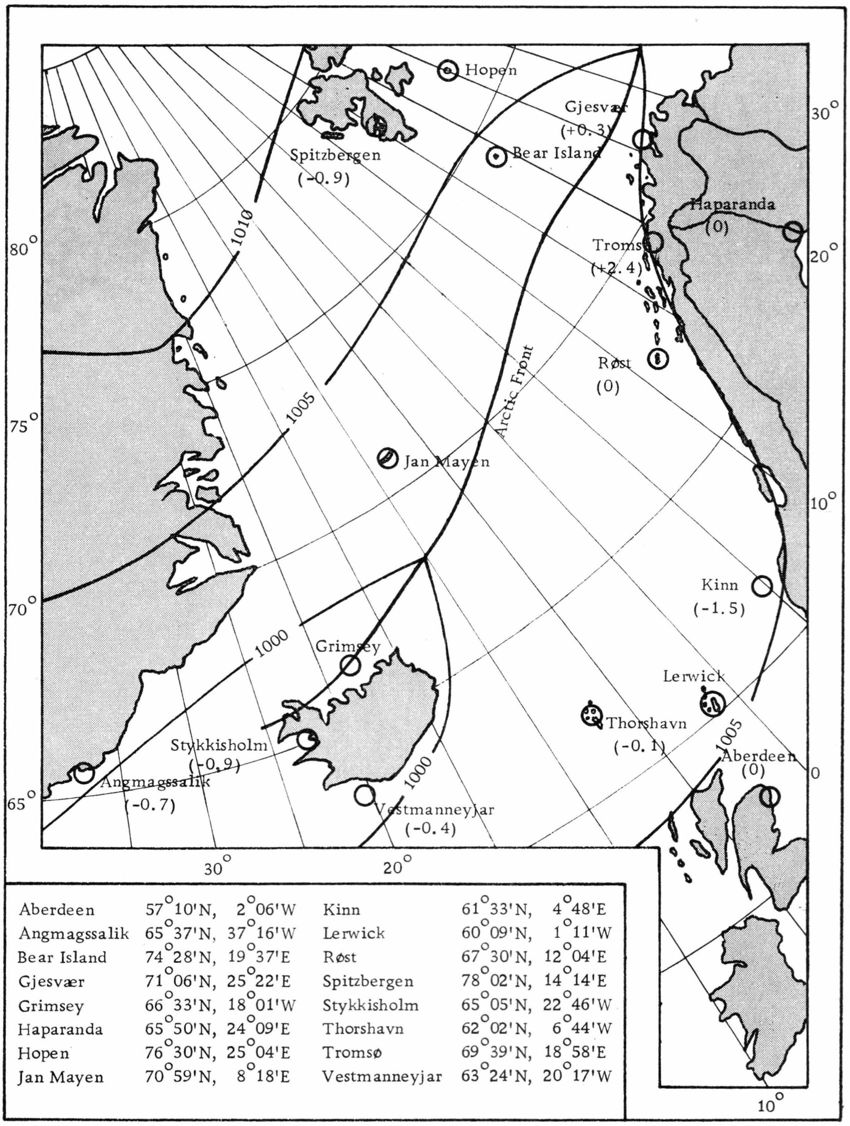
to above. If ice melting so severely depresses the summer temperatures, one would expect a rather closer correlation between summer means and ice cover, the more so since ice cover here is defined as that of May–August, the ice melting period.

In Figure 19 both the ice cover graph and the 4-station anomaly have been smoothed by using 5-year moving averages. In an attempt to obtain an indication of the causality, the temperature anomaly has been compared also with previous and subsequent year's ice cover. The results are listed below the graphs, and although the latter correlation is slightly better, it can be said that the temperature anomaly neither manifests itself as a lingering effect of the ice regime nor does it presage ice variations to any marked degree.

Wind-Current Relationships

Weather over the Greenland Sea depends on the momentary location of the atmospheric Arctic front and its generative low pressure cell, the extensive cyclone known as the Icelandic low. Rodewald (1966) has charted the mean pressure distribution in the area, and his findings have been used to show the front's location in Figure 20.

It is of interest to see whether this position has remained statistically fixed during the period in which the ice cover has displayed a decreasing trend, and to this end the pressure changes registered at 11 stations in the area have been calculated and listed in Table 4. Each change has been found as the net difference between the mean pressure in the 1900–19



Aberdeen	57°10'N, 2°06'W	Kinn	61°33'N, 4°48'E
Angmagssalik	65°37'N, 37°16'W	Lerwick	60°09'N, 1°11'W
Bear Island	74°28'N, 19°37'E	Røst	67°30'N, 12°04'E
Gjesvær	71°06'N, 25°22'E	Spitzbergen	78°02'N, 14°14'E
Grimsey	66°33'N, 18°01'W	Stykkisholm	65°05'N, 22°46'W
Haparanda	65°50'N, 24°09'E	Thorshavn	62°02'N, 6°44'W
Hopen	76°30'N, 25°04'E	Tromsø	69°39'N, 18°58'E
Jan Mayen	70°59'N, 8°18'E	Vestmanneyjar	63°24'N, 20°17'W

Figure 20. Mean pressure distribution in the European Arctic (after RODEWALD). Pressure change from 1900–19 period to 1920–39 period is shown below each station.

period and that in the 1920–39 period. The pressure changes have been noted in Figure 20 and they indicate that in the first four decades of this century the Icelandic low has shifted northward, particularly between Spitzbergen and Norway, and that it has become somewhat stronger.

To verify this strengthening of the cyclone, a measure of the steepness of the pressure gradient in the key area of the Faeroes can be devel-

Table 4. Change in mean atmospheric pressure at selected stations from the period 1900–19 to 1920–39.

Station	P 1900–19	P 1920–39	Change	Remarks
Aberdeen	1011.1	1011.1	± 0	
Angmagssalik.	1007.6	1006.9	− 0.7	1910–11, 1924 missing
Gjesvaer	1007.0	1007.3	+ 0.3	1927–39 values from Sletnes
Haparanda	1009.5	1009.5	± 0	
Kinn	1010.0	1008.5	− 1.5	1921–22, 1932–39 missing
Røst.	1006.8	1006.8	± 0	
Spitzbergen	1009.2	1008.3	− 0.9	1910–11, 1930, 1938–39 missing
Stykkisholm	1006.2	1005.4	− 0.8	
Thorshavn	1008.4	1008.3	− 0.1	
Tromsø	1003.6	1006.0	+ 2.4	
Vestmanneyjar	1005.3	1004.9	− 0.4	

oped by computing the difference between annual means of atmospheric pressure at Thorshavn and Stykkisholm. This difference is shown in Figure 21, and a slight gradual steepening of the gradient is noticeable from 1900 to 1939, as the difference has increased by 0.195 mb per decade.

Northward displacement of the Arctic front will cause depressions to pass north of Spitzbergen, strengthening the southerly wind component at Bear Island and Nordostrundingen. Southward displacement will cause depressions to follow a more southerly track from the Faeroes toward the Baltic, strengthening the southerly component over the Faeroe-Shetland Channel. The wind regime is dominated by these movements of the front and by the increase in atmospheric circulation resulting from the slight strengthening of the low pressure cell. This should be borne in mind in examining the influence on the ice formation, which the variations in prevailing winds can exert by water mass transport.

The water mass being created in the gyre northeast of Jan Mayen is the product of admixture of diluted Atlantic Water from the West Spitzbergen Current with Polar Water of the East Greenland Current. South of Jan Mayen, Polar Water of the Jan Mayen Current mixes with Atlantic Water from the Norwegian Current and possibly with some of the Irminger Current. As we are dealing with surface currents, the wind regime in a given season will influence both directions and intensities of flows and hence temperature and salinity of the water in the area of ice formation. The importance of the proportion of Atlantic Water advected is obvious.

If the winds at Jan Mayen are between south and west, while those of Tromsø and Røst (see Figure 20) are between south and southwest,

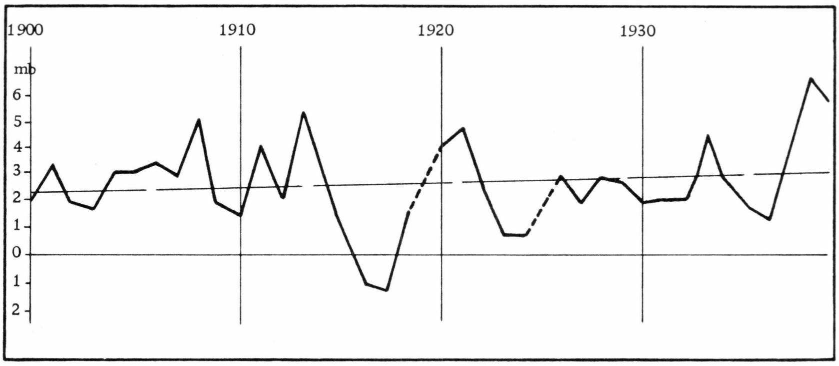


Figure 21. Difference in annual mean pressure between Thorshavn and Stykkisholm. The regression line, calculated by the method of least squares, indicates a steepening of the gradient amounting to 0.195 mb per decade.

the Norwegian Current is accelerated toward the Barents Sea, and mass transport into the Spitzbergen branch of the system is not aided. If on the other hand winds between northeast and southeast predominate, mass transport is diverted toward Spitzbergen, increasing admixture of warmer water in the gyre northeast of Jan Mayen and possibly opposing the East Greenland Current at its exit from the Arctic Ocean.

By analogous reasoning, winds between northwest and northeast at Spitzbergen will increase the admixture, whereas winds between east and south will speed the current on into the Arctic Ocean.

If the cumulative effect of all the winds in a season is expressed quantitatively, its correlation with the ice fluctuations will yield information about the actual importance of the wind regime. Determination of such cumulative effect will be attempted. In order first to quantify the influence of wind at a certain station, it is practical to define a wind effect index for winds of a given direction as WEI = duration × strength, where duration is measured as number of observations[22]) and strength is expressed as Beaufort force. If the Beaufort scale expresses wind velocity proportional to its force numbers, [23]) and if the wind's frictional effect on the surface current is assumed proportional to wind velocity, we have a crude but meaningful measure to work with.

The WEI at a given station will comprise the wind component which tends to aid transport of Atlantic Water to the ice border. The effect of high WEI value should accordingly diminish the ice cover,

[22]) Most stations record three daily observations. Where a greater or smaller number is recorded, the result has been factored down or up to become comparable.

[23]) This has been achieved in the following by counting wind strengths of Beaufort 3 and less as being of force 3, whereby the scale becomes roughly linear in terms of wind velocity.

Table 5. Correlation coefficients between Greenland Sea ice fluctuations and Wind Effect Indices (WEI) at selected stations in and near the Greenland Sea.

Station	WEI Component	Years used	Full Year Correlation	Period	Correlation
Bear Island . . .	E	1928–39, 46, 48–57	+0.048	Jan–Apr	−0.196
Jan Mayen. . . .	E	1922–39, 46–57	+0.099	Jan–Jun	−0.076
Røst.	E	1900–39, 46–57	−0.202	Oct–Mar	−0.460
Spitzbergen . . .	N	1912–39, 46–57	+0.043	Oct–Mar	+0.089
Tromsø	E	1900–39, 46–57	+0.481	Oct–Mar	−0.610
Vestmanneyjar	E	1900–19, 24–39, 46–52	+0.238	Dec–May	−0.030

whereas a low WEI value could be expected to enlarge it. In Table 5 are listed the wind components of which the WEI is composed at six stations. Correlations have been calculated between the WEI and the ice cover by full years and by periods which depend on the estimated travel time of a water parcel from the point of wind influence to the ice border, and which also depend on the season in which the advection of Atlantic Water can be expected to have a maximum effect on the ice. For example, at Tromsø the E component will aid mass transport toward the ice; the advection will influence ice formation from January through June; and the water's travel time is estimated at three months. Hence at Tromsø the E-component is summed over the period October–March.

The results from the six stations are listed in Table 5, and it will be noticed that the WEI from the periods of estimated maximum effect from Norwegian stations correlate well with the ice cover. At Spitzbergen there is no significant correlation, which can probably be explained by the northerly wind component's dual effects of impeding the outflow of water from the West Spitzbergen Current, shifting it instead into a path contiguous to the East Greenland Current, and of aiding the inflow of Polar Water and ice. This would account for the simultaneous presence of substantial amounts of both PW and diluted AW in the *Belgica* section in Figure 6. The Jan Mayen correlation is also poor, but the number of years from which data are available from this station is too short to accord them importance comparable to that of the Norwegian series.

The period WEI anomalies are in Table 6 summed over the years in which the ice cover deviated from its mean by an amount greater than *s*. The Norwegian stations again show significant results, confirming the expected relationship between wind effect and ice cover.

Table 6. Anomalies of Wind Effect Indices in years the ice cover x deviated from its mean by an amount greater than s.

Bear Island	$x > +s$	1905	1906	1917	1918	1934		Total
		..	..	..	..	− 3.1		− 3.1
	$x < -s$	1904	1925	1930	1931	1933	1936	Total
		..	..	− 0.8	+ 0.1	0	+ 0.4	− 0.3
Jan Mayen	$x > +s$	1905	1906	1917	1918	1934		Total
		..	..	..	..	− 0.6		− 0.6
	$x < -s$	1904	1925	1930	1931	1933	1936	
		..	+ 1.5	+ 2.7	− 2.7	+ 1.2	− 0.8	+ 1.9
Røst	$x > +s$	1905	1906	1917	1918	1934		Total
		− 0.9	− 0.8	− 0.5	− 0.5	− 0.6		− 3.3
	$x < -s$	1904	1925	1930	1931	1933	1936	
		− 0.7	+ 0.8	+ 1.0	+ 0.3	− 0.4	+ 1.4	+ 2.4
Spitzbergen	$x > +s$	1905	1906	1917	1918	1934		Total
		..	..	− 0.3	− 0.5	..		− 0.8
	$x < -s$	1904	1925	1930	1931	1933	1936	
		..	− 0.5	− 0.4	− 0.4	− 0.5	− 0.3	− 2.1
Tromsø	$x > +s$	1905	1906	1917	1918	1934		Total
		− 0.1	− 0.3	− 0.9	− 0.7	0		− 2.0
	$x < -s$	1904	1925	1930	1931	1933	1936	
		+ 1.2	+ 0.1	− 0.2	+ 0.4	− 0.2	+ 0.9	+ 2.2
Vestmanneyjar	$x > +s$	1905	1906	1917	1918	1934		Total
		..	..	..	..	− 2.8		− 2.8
	$x < -s$	1904	1925	1930	1931	1933	1936	
		..	− 4.0	+ 0.2	− 1.7	+ 2.0	..	− 3.5

VI. DISCUSSION

The task of evaluating from the foregoing findings the relative importance of the factors influencing the ice regime is complicated by the factor's simultaneous presence, by their variability, and by their having little or no apparent interdependence. It is practical to treat them in reverse order of their estimated importance, thus deleting some at the start.

Variations in surface salinity of the Atlantic water may be induced by disturbances in the evaporation-precipitation balance equatorward in the Gulf Stream system, or their cause may lie in the varying degrees to which vertical and lateral mixing takes place in the surface layers. It is evident that they exist as a seasonal occurrence, as a year-to-year phenomenon and as decadal trends. Their direct effects in lowering the freezing point are negligible, however. The 0.06 $^0/_{00}$ salinity increase from 1902–17 to 1919–39 would only register an effect four orders of magnitude below that of a 20 $^0/_0$ variation in one month's heat import through the Faeroe-Shetland channel. The salinity fluctuations can therefore be ignored.

Fluctuations in surface temperature as computed by Rodewald might be of greater importance than those of salinity by two orders of magnitude but are still insignificant. Besides, the data give no indication that the fluctuations extend to deeper parts of the North Atlantic Current. Their influence on the ice can therefore also be considered below the threshold of detectability.

Precipitation falls in the same category as the above two factors. If an anomalous 100 mm of rain water were added at the ideal time to the sea surface in the area of ice formation, and if it were mixed to a depth of only 3 m, the effect would still be an order of magnitude below the effects of known current variations. Hence precipitation can be ignored. As shown in Chapter VI, evaporation is also insignificant.

It bears emphasizing that the above factors certainly all affect the ice regime. The laws of physics compel us to recognize that such effects exist in any case. But in the composite picture of numerous natural forces interacting, the noise level is too high to permit their detection.

Air temperature variations present an enigma. There is no questioning that a pronounced decrease in ice cover over the first four decades of this century has been accompanied by a strong increase in air temperatures of the ice region, and it would indeed have been astounding if such had not been the case. The relationship nevertheless is obscure when year-to-year variations are considered. The best that can be said at this juncture is that available evidence fails to indicate air temperature fluctuations as a cause of the ice changes.

Wind influence is the most difficult factor to assess quantitatively. Despite extensive records from the perimeter and some shorter series from within the ice region, the effects of wind defy easy analysis. This is not merely due to the multifarious nature of wind's influence through evaporation, precipitation, heat exchange, mass transport, depth of the mixed layer and spreading or compaction of the ice, but it is particularly due to an opposition of effects on the ice by wind components in certain key locations. Some examples will serve to illustrate. A strong easterly component south of Iceland will oppose mass transport of Atlantic Water into the Norwegian Current; but it will simultaneously promote mass transport via the Irminger Current through Denmark Strait toward the ice. A strong southerly component between Iceland and the Faeroes will aid mass transport of Atlantic Water across the sills between the Shetlands and the Faeroes and between the Faroes and Iceland; but it will at the same time oppose outflow of Polar Water in the Irminger Current encircling Iceland. A strong northerly component between Nordostrundingen and Spitzbergen will transport both ice, Polar Water and Atlantic Water from the West Spitzbergen Current into Greenland Sea. Conversely, absence of compound influence may well explain the rather good correlation between the ice cover and the easterly wind component at Norwegian coastal stations. From there, an east wind can but transport water northwestward, toward the ice.

Detailed synoptic wind data and much additional information about the precise current structures are needed from the areas of oppositional influence before a clear picture can be developed of the winds' effects. At present, the effects can only be termed verifiable, but not dominant.

It is realized, of course, that this discussion has been confined to the wind-induced mass transports within the Greenland-Norwegian Sea area. It has been practical to deal with fluctuations in the volume of water entering from the North Atlantic Current under the heading of current effects, although the unknown, distant causes for this phenomenon probably lie in winds of the lower latitudes.

Year-to-year fluctuations in the flows of Polar Water and Atlantic Water into the Greenland Sea remain as the primary cause for year-to-year ice variations. This is concluded first on available evidence: 1) as calculated from hydrosections, the currents transport such quantities of heat (the southern flows) and ice (the northern flow) that their variations have the physical capacity to influence the ice cover; 2) their intensities in fact do vary, both seasonally and from one year to another; and 3) a correlation has been shown between high intensity and ice cover, negative for the southern flows and positive in the case of the northern. Second, the conclusion is supported on deductive grounds: the currents are the only factor analyzed which has the physical capacity to influence the ice cover.

In summary, the causes for the short-term ice fluctuations are to be sought in the relatively easily monitored flows of water and air into and out of the Greenland Sea. The much more complex air-sea energy exchanges and evaporation-precipitation relationships, difficult or impossible to measure synoptically, are unimportant by comparison. One can therefore anticipate the day when a suitable system of recording buoys together with shore-based installations will provide the information for reliable forecasts of ice conditions and their concomitant weather effects.

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