

THE PHYSICAL OCEANOGRAPHY OF THE NORTHERN

BAFFIN BAY REGION

by


ROBIN DAVIE MUENCH

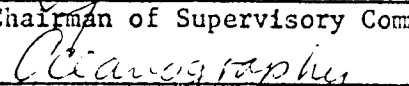
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We have carefully read the dissertation entitled The Physical Oceanography of
The Northern Baffin Bay Region

Mr. Robin Davie Muench submitted by
in partial fulfillment of
the requirements of the degree of Doctor of Philosophy
and recommend its acceptance. In support of this recommendation we present the following
joint statement of evaluation to be filed with the dissertation.


In northern Baffin Bay, waters from the north Atlantic Ocean interact with Arctic Ocean waters entering through major passages in the Canadian Arctic Archipelago. The region is of particular scientific interest because it is (a) the site of the North Water, a semi-permanent polynya which persists throughout the winter and offers unique opportunities for studies of air-sea interaction phenomena and (b) suspected to be the site of formation of Baffin Bay Bottom Water.

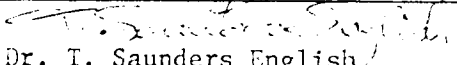
Mr. Muench has investigated the hydrography and circulation of the northern Baffin Bay region by analyses of pertinent oceanographic and atmospheric data. The hydrographic structure is examined in detail and new features of the temperature and salinity distributions revealed. Temperature-salinity analyses clarify the probable sources of the water masses, including the Bottom Water. Surface drift data and subsurface current measurements are used to clarify net flow patterns. Dynamical analyses corroborate the known cyclonic baroclinic flow and detect geographic and temporal variations in the baroclinic currents. Probable causes for the circulation are investigated for the first time.

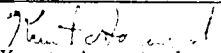
Knowledge concerning the North Water is summarized, past theories for its formation and maintenance discussed, and recent data utilized to test the validity of these theories. The impossibility of solving the North Water problem without acquisition of winter data is pointed out.

This dissertation consolidates past knowledge of the northern Baffin Bay region and enlarges on this knowledge by analyzing recent data. It represents a significant advance in oceanographic knowledge of the eastern Arctic.

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I. INTRODUCTION

1.1. Scientific and historic background

Northern Baffin Bay (Fig. 1) was first explored in 1616 by Baffin, a British pilot and navigator who had joined an exploratory mission in search of a northwest trade passage to the Orient. He entered Baffin Bay via Davis Strait and sailed northward as far as Smith Sound, being stopped there by south-flowing pack ice. The ice-filled nature of the channels exiting from Baffin Bay's northern end led him to conclude that there was no waterway leading northward or westward from that region (Markham 1881). In addition, he became the first European to record the existence of the ice-free area in northern Baffin Bay known as the North Water, although his observation was made during the summer when the phenomenon is not clearly defined. Baffin's report that northern Baffin Bay was landlocked so disillusioned believers in a northwest passage that it discouraged further exploration in the eastern Arctic for nearly two centuries.

Arrival of the whaling ships LARKIN and ELIZABETH at the North Water in 1817 marked the establishment of a northern Baffin Bay whale fishery. This, coupled with international competition to reach the North Pole and the assumption by Canada of sovereignty over the Arctic Islands, furnished the motivation for further exploration of the northern Baffin Bay region.

The British Arctic Expedition under Ross in the late 1840's explored much of Baffin Bay and penetrated through Lancaster Sound into the channels further west. Their vessels, the INVESTIGATOR and the ENTERPRISE, became icebound west of Lancaster Sound and drifted eastward to a point east of Bylot Island. This drift provided the first indication of an eastward surface current through Lancaster Sound (Ross 1850).

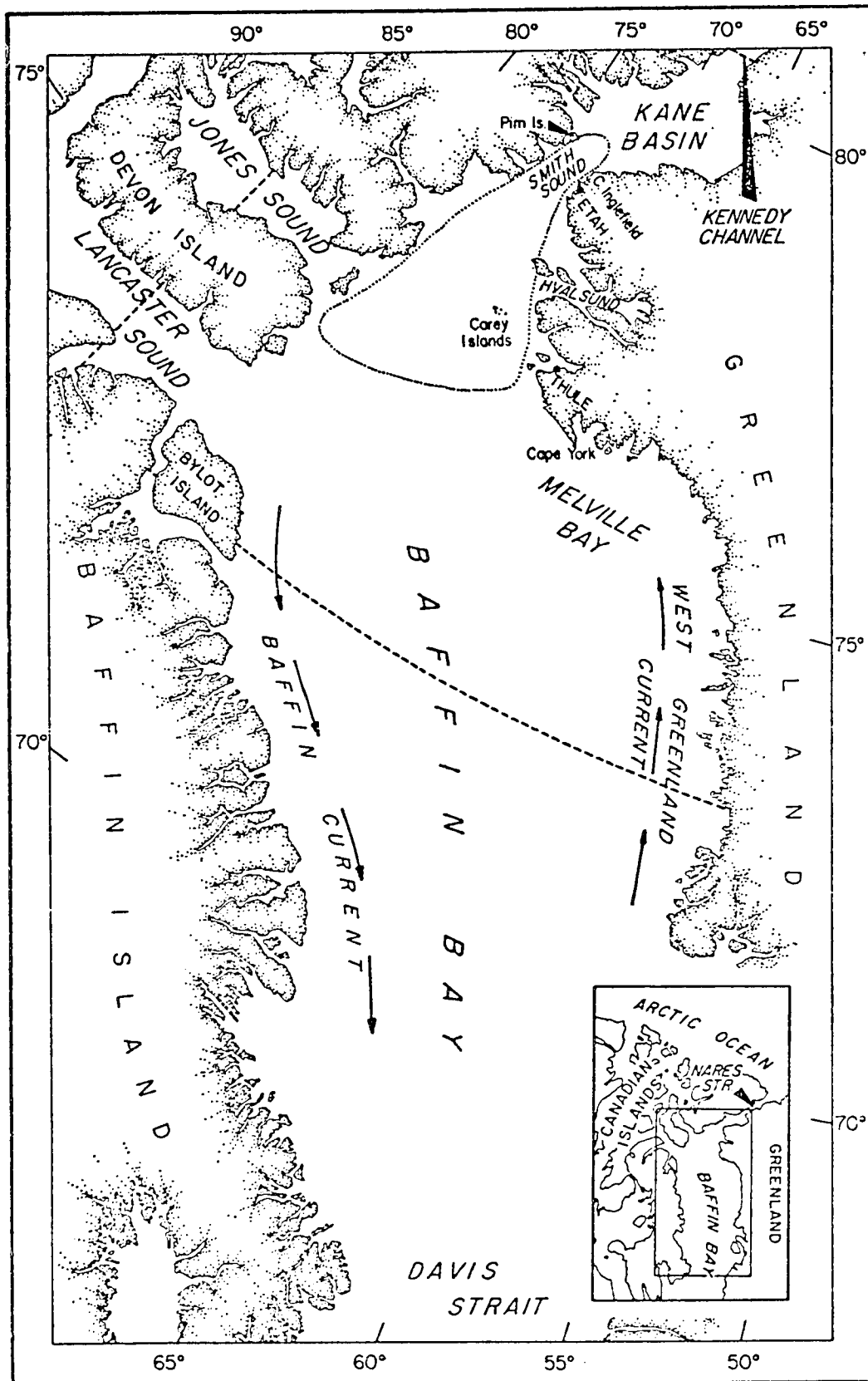


Figure 1. Geographical locations in the Baffin Bay and surrounding regions. The dashed line delineates the area encompassed by the present study, while the dotted line indicates the approximate winter boundary of the North Water.

The Ross party was closely followed by Parry, who sailed the HECLA and the GRIPER through Lancaster Sound and obtained the first subsurface temperature measurements and water samples from that region. Unfortunately, these were so imprecise as to be of little use today.

In 1845-47 Franklin journeyed from West Greenland through Lancaster Sound and thence southward where his ship was icebound and subsequently deserted. Caught in the Arctic winter, he and his party perished. Arctic exploration for the next decade was carried out primarily by expeditions searching for the lost Franklin party.

In 1850 the United States Grinnell Expedition in search of Franklin, led by Kane, penetrated Lancaster Sound in the ADVANCE and the RESCUE. The vessels became icebound west of Devon Island and drifted southward then eastward through Lancaster Sound, parallelling the previous drift of the INVESTIGATOR and the ENTERPRISE, then southward to Davis Strait. This corroborated the previous observation of an eastward surface flow through Lancaster Sound and, in addition, indicated a southward surface flow in western Baffin Bay (Kane 1857a).

The Inglefield party followed the Grinnell Expedition and, in 1852, entered Kane Basin in the RESOLUTE. They then journeyed westward into Lancaster Sound, where they became icebound. The RESOLUTE's ensuing drift indicated an eastward surface flow in Lancaster Sound and a southward flow in western Baffin Bay. The charts drawn by Inglefield are among the earliest to depict Smith Sound as a through channel extending northward and Jones and Lancaster sounds as through channels extending westward from northern Baffin Bay.

In 1853 Kane led the Second Grinnell Expedition in search of Franklin, and in the ADVANCE was able to penetrate to the southeast coast

of Kane Basin where the party became icebound. The expedition consequently spent two winters there, during which they obtained surface air temperature and wind velocity measurements. In addition, Kane observed icebergs moving southward through Smith Sound, attributed this movement to a southward surface current, and first noted frequent "water skies" and fogs in the winter which he attributed to the presence of the North Water (Kane 1857b). Exploratory parties reported occasional northward flow in Kennedy Channel, so that the surface current flow through the system seemed variable though exhibiting a net southward flow.

In 1857 the FOX became icebound in Melville Bay. Her ensuing drift suggested a southward surface flow in the central portion of Baffin Bay from Cape York to Davis Strait (McClintock 1859). The FOX expedition was the last of the parties sent in search of Franklin.

In 1860 Hayes, aboard the UNITED STATES, took soundings and made ice observations in Baffin Bay and Smith Sound, and did so again in 1869 from the PANTHER. During the first of these expeditions he recorded the presence of warm Atlantic Water at subsurface levels in northern Baffin Bay (Hayes 1867). Open water in the northern Baffin Bay region was observed as late as February 1861, again corroborating the existence of the North Water.

Oceanographical, geographical and meteorological knowledge of the entire Baffin Bay region, due primarily to the explorers Baffin, Ross, Inglefield, Kane and Hayes, was summarized by Petermann (1867). He described a warm northward current in eastern Baffin Bay, a colder southward current in western Baffin Bay and the cold water flowing inward through Smith, Jones and Lancaster sounds. The northward current, thought

by him to be an arm of the Gulf Stream, was held responsible for the formation of the North Water.

In 1871 the United States Arctic Expedition under Hall, with Bessels as chief scientist, penetrated north of Kane Basin into Nares Strait in a U.S.-sponsored attempt to reach the North Pole. On their return trip southward they obtained the first subsurface temperature measurements in the southern Kane Basin-Smith Sound region. These measurements indicated a uniformly cold south-flowing water column, rather than the warm subsurface layer which Hayes' earlier work had led them to expect. Bessels (1876) said that the North Water could therefore only be due to a combination of swift offshore winds and tidal currents. The expedition's vessel POLARIS became icebound in Smith Sound; the party abandoned the ship and drifted southward through Davis Strait on an ice floe (Tyson 1874), providing additional evidence of a southward surface flow through Baffin Bay.

A first attempt to explain the presence of warm water in northern Baffin Bay was made by Carpenter (1876) who, using data from Davis Strait, assumed the northward flow of warm water to be a compensating flow to "fill the void" left by a southerly flow of colder, denser water.

In 1875 Nares sailed the ALERT, with Moss as scientist, northward through Nares Strait as far as the Arctic Ocean in a British-sponsored attempt to reach the North Pole. Scientific observations included temperature and chloride content measurements down to 115 fm (210 m). These measurements, made at locations ranging from the Arctic Ocean to southern Baffin Bay, allowed the identification of water types with different temperature and salinity characteristics. Nares Strait was observed to contain an upper layer of cold, low salinity Arctic Water underlain by

a warmer, more saline layer of Atlantic Water. Moss (1878) suspected that the warmer water was flowing northward from Baffin Bay.

Following the Nares expedition, interest rose in the fisheries potential of some relatively shallow banks off the western Greenland coast. Exploratory parties were sent to Baffin Bay to study the hydrography on these banks. The first was the Swedish Nordenskjöld expedition, working from the SOFIA with Hamberg as scientist, which cruised up the western Greenland coast nearly to Smith Sound. They measured temperatures and chlorinities down to several hundred meters depth, leading to the first description of the waters that form the West Greenland Current and the three-layered water structure of Baffin Bay. This structure was described as a cold, low salinity surface layer, a warmer and more saline mid-depth layer and a colder deep layer slightly less saline than the mid-depth layer (Hamberg 1884).

In 1898-99 an expedition aboard the WINDWARD obtained additional information on ice cover in northern Baffin Bay and surface currents as deduced from ice drift. During 1898-1902 the Norwegian Arctic Expedition aboard the FRAM, under O. Sverdrup, measured surface water temperatures in Baffin Bay and Jones and Lancaster sounds.

Petterson (1900) analyzed the available data and referred to the warm mid-depth layer extending northward into Baffin Bay as Atlantic Water. He also depicted two currents flowing southward in Baffin Bay: one along the western shore from Smith Sound to Davis Strait; and one originating off Cape York and joining the other in the Davis Strait region. Mecking (1906) recognized these two southward currents and considered them in explaining the North Water. He thought it due to upwelling of the warm Atlantic Water, in the region of divergence between the two southward

currents in northern Baffin Bay, coupled with decreasing depth and resulting constriction of flow of the north-flowing warm water upon reaching Smith Sound. The concept of a southward current originating off Cape York appears to have arisen primarily due to the difficulty generally experienced by mariners navigating through Melville Bay to the North Water area. Additional factors were the presence of cold Arctic surface water and a southward drift pattern southwest of Cape York.

Low (1906) explained the North Water as due to a transport of ice away from the area by wind and water currents; ice was prevented from entering by the ice dam across northern Smith Sound.

The results of physical oceanographic observations made in 1908-09 on the TJALFE expedition off coastal western Greenland by Nielsen led to some conclusions pertinent to oceanographic processes in northern Baffin Bay:

- a. Baffin Bay Deep Water (the lowermost layer in the three-layered structure) cannot be formed within the Bay itself, but must be formed from south-flowing Arctic Ocean Water due to the requirement that its salinity be high enough for the resultant mixture to sink; and
- b. A warm current flowing from Davis Strait to northern Baffin Bay does not exist insofar as effects on surface conditions in northern Baffin Bay are concerned; the warm water is separated from the surface by a colder, less saline layer (Nielsen 1928).

The first conclusion required a net Arctic Ocean Water inflow from the north, in agreement with the surface currents deduced from ship and ice drifts. The second ruled out the possibility, considering only summer

observations, that the North Water was maintained by a subsurface flow of warm Atlantic Water.

Further oceanographic knowledge of northern Baffin Bay was acquired when Wulff, traveling with Rasmussen in 1916, occupied 27 hydrographic stations from southern Baffin Bay nearly to Smith Sound. Wulff perished before analyzing his data, which were subsequently published by Knudsen (1923). The results allowed tracing the warm intermediate water across Melville Bay to a point off Cape York.

In April 1912 the TITANIC sank after colliding with an iceberg off the Grand Banks. As a result the International Ice Patrol was formed in 1914 and assigned to the U.S. Coast Guard. The Ice Patrol's initial duties were to report the positions of all icebergs menacing the northeast Atlantic shipping lanes. It became obvious, however, that some means were needed for predicting severity of an iceberg season. Knowledge of Baffin Bay was insufficient, at that time, for such predictions.

In order to acquire further oceanographic information from Baffin Bay the GODTHAAB carried out a cruise in 1928. The unknown factors prior to this expedition were enumerated by Riis-Carstensen of the GODTHAAB:

- a. The course and extent of the currents in Baffin Bay, including the extent and influence of the Atlantic Water; and
- b. Chemical, biological and hydrographical conditions north of Davis Strait and in its western part, particularly as applicable to the western Greenland fishery (Riis-Carstensen 1931).

The GODTHAAB expedition, in attempting to illuminate these problems, carried out the most complete oceanographic survey of Baffin Bay to that

time. Oceanographic stations were occupied from south of Davis Strait to Smith Sound.

A preliminary investigation of the GODTHAAB hydrographic data prompted Smith (1931) to suggest that the North Water might be maintained by southward currents sweeping the ice away as it is formed rather than by upwelling of warm water advected from the south. The fast ice in Kane Basin would be prevented from flowing southward by the arch-shaped ice front which had frequently been observed in northern Smith Sound approximately between Pim Island and Cape Inglefield. From the same data Kiilerich (1933) showed that the warm Atlantic Water was never found shallower than 400-500 m, and that the overlying water column was too stable to allow upwelling to maintain the North Water. He hypothesized that in the winter a southward flow of cold, more saline Arctic Ocean Water might lower the stability of the water column enough so that upwelling could occur, but emphasized that winter measurements would be necessary to prove this. Kiilerich thought that wind alone could not maintain the North Water, since this would also have caused open water in Melville Bay which had not been observed.

Kiilerich (1939) carried out a dynamical analysis of the GODTHAAB hydrographic observations. Major northward and southward baroclinic currents were found along the eastern and western shores of Baffin Bay; a cyclonic gyre was found in its northern portion. An intensification of the northwestward current southwest of Cape York, of unknown cause, was found. Heavy pack ice off the Baffin Island coast prevented completion of the sections there, so the south-flowing Baffin Current was only partially surveyed. New complexities in the water and current structures were revealed, however, which were deserving of further exploration.

Riis-Carstensen (1948) utilized the GODTHAAB data in explaining the North Water as due to a combination of winds, upwelling and a gyral motion which would cause newly formed ice to move from the center outwards, contributing toward keeping the area ice-free.

Several expeditions penetrated to northern Baffin Bay in the 1930's, but did little oceanographic research. In 1937 the ISBJORN carried out meteorological studies near Thule and charted the Carey Islands. In 1937-38 the British Arctic Expedition wintered in the Thule region and noted that Smith Sound remained ice-free throughout the winter, the fast ice to the north extending as a west-east arch from Pim Island to the north of Etah. The EFFIE M. MORRISSEY made several cruises to northern Baffin Bay and obtained trawl samples, but no physical or chemical data.

In 1940 the CGC NORTHLAND was able, due to unusually light ice conditions, to carry out a hydrographic survey in the region of the Baffin Current. The results indicated the current (at that time) to be accompanied by two countercurrents. The computed dynamic topography correlated well with that from the 1928 GODTHAAB data for the rest of Baffin Bay, suggesting that the countercurrents observed from the NORTHLAND were not anomalous (Barnes 1941).

Hare and Montgomery (1949a; 1949b) suggested, using winter air temperature data from land stations in the northern Baffin Bay region, that the North Water might actually be of much larger extent than was previously believed. They listed the causes of polynyas, such as the North Water, as turbulent mixing of surface with warmer subsurface water, upwelling and excessive tidal ranges.

Dunbar (1951) reviewed, in general fashion, knowledge of eastern Arctic waters at that time. He suggested that phenomena similar to those

which cause the North Water might lead to observed open water areas off Jones and Lancaster sounds. In particular, he cited Kiillerich's (1933) theory.

Vibe (1950) summarized ice conditions along the northwest coast of Greenland, noting the North Water and several lesser open water areas in Kane Basin and near the mouth of Hvalsund.

Moir Dunbar (1954) published the first map based on both air and ground observations to depict winter ice conditions in northern Baffin Bay, and indicated the direction and pattern of ice withdrawal during the onset of the summer melt. Both winds and oceanographic conditions were held responsible for the North Water.

From 1954 to 1957 HMCS LABRADOR carried out summer oceanographic investigations in the eastern Arctic. Bailey (1957) considered the resulting data and explained the North Water as partially due to a pool of water warmed in summer which, he claimed, persisted in that area throughout the winter and prevented ice formation. He also (1956) reaffirmed the theory that Baffin Bay Deep Water originated in the Arctic Ocean and flowed southward via Nares Strait.

In the 1940's and 1950's, acquisition of meteorological data from the eastern Arctic was proceeding concurrently with acquisition of oceanographic data. Establishment of the joint U.S.-Canadian weather stations in the Canadian Islands and the U.S. air base at Thule provided year-round observations. Inasmuch as oceanographic and atmospheric conditions are closely interrelated, acquisition of meteorological data was a major step forward in the study of northern Baffin Bay and eastern Arctic waters in general.

Several qualitative studies of the North Water problem in the late 1950's and early 1960's considered mechanisms of ice removal by currents and prevention of ice formation by mixing or upswelling (Schule and Wittman 1958; Simpson 1958; Kupetskii 1962). Collin and Dunbar (1964) pointed out, however, that summer data showed no vertical mixing or upwelling in the North Water area.

The 1960's saw extensive oceanographic exploration in the northern Baffin Bay region. HMCS LABRADOR carried out oceanographic programs, primarily in Baffin Bay, during the summers of 1960 to 1965. The CGC EVERGREEN obtained detailed oceanographic data from northern Baffin Bay and Smith Sound in 1963, taking advantage of unusually light ice conditions to penetrate to northern Kane Basin. The general oceanographic conditions observed in 1963 were presented by Franceschetti, McGill, Corwin and Uchupi (1964).

The water exchange through Nares Strait was analyzed, using the dynamic method and water mass analyses, by Collin (1965) and Muench (1966). The occurrence of an irregular, net southward flow was deduced.

Nutt (1966) chronicled the southward drift in 1963 of ice island WH-5 from the Arctic Ocean through Nares Strait and Baffin Bay. This study revealed a net southward, highly irregular surface flow through the system, with at least one instance of flow reversal in Nares Strait. Walmsley (1966) attempted to compute the heat budget for Baffin Bay, but was unable to obtain conclusive results.

In 1966 the CGC EDISTO acquired oceanographic data from northern Baffin Bay and Smith, Jones and Lancaster sounds. Dynamic and water mass analyses were carried out, using the 1966 data, by Palfrey and Day (1968).

Godin (1966), in a theoretical study, indicated the probable nature of diurnal and semidiurnal tides in Baffin Bay.

At the onset of the present study in 1967, oceanographic features in northern Baffin Bay which remained poorly defined were:

- a. Details, such as gyres, countercurrents and possible regions of upwelling, of the summer circulation;
- b. Mass and heat budgets throughout the year;
- c. Occurrence and causes of periodic and non-periodic temporal variations in water masses and currents;
- d. Driving forces for the circulation;
- e. Details of the mixing between different water masses; and
- f. All aspects of the winter hydrographic structure and circulation.

Information concerning the above is requisite to solving the problems of the North Water and Baffin Bay Deep Water formation. To acquire this information, the Arctic Institute of North America formed the Baffin Bay-North Water Project in 1967. Under this program, cooperative among the Arctic Institute, the University of Washington and the United States Coast Guard, several field programs were conducted. In June 1967 oceanographic data were acquired from the Lincoln Sea (Seibert 1968) to better ascertain the nature of Arctic Ocean Water available for southward flow through Nares Strait. In September 1968 the CGC WESTWIND conducted a detailed oceanographic survey in the Smith Sound-Cape York region (Muench in press), and in September 1969 the area south and southwest of Cape York was surveyed

by the same vessel (Muench and Tidmarsh 1969). In May 1969 a series of oceanographic measurements were made through the pack ice of western Kane Basin (Tidmarsh et al. 1969).

The present study is limited to northern Baffin Bay (Fig. 1). Reasons for limiting the study, rather than including the entire bay, are:

- a. The North Water, of primary scientific interest, is located in northern Baffin Bay;
- b. The answer to the problem of Baffin Bay Deep Water formation is suspected to lie in this region;
- c. The decay of the warm Atlantic Water layer appears to occur in this region;
- d. Data coverage is denser in northern than in southern Baffin Bay; further south, coverage is so poor as to preclude a detailed study; and
- e. Inclusion of the southern Baffin Bay and Davis Strait regions would add to the complexity of the study without appreciably illuminating the North Water and Deep Water problems.

1.2 Some bathymetric features

Baffin Bay is characterized by a deep (nearly 2400 m), level basin in the central western portion (Fig. 2). This basin is flanked on the east by a continuous series of banks shallower than 400 m which extend about 400 km westward from the western Greenland coast. The basin's western

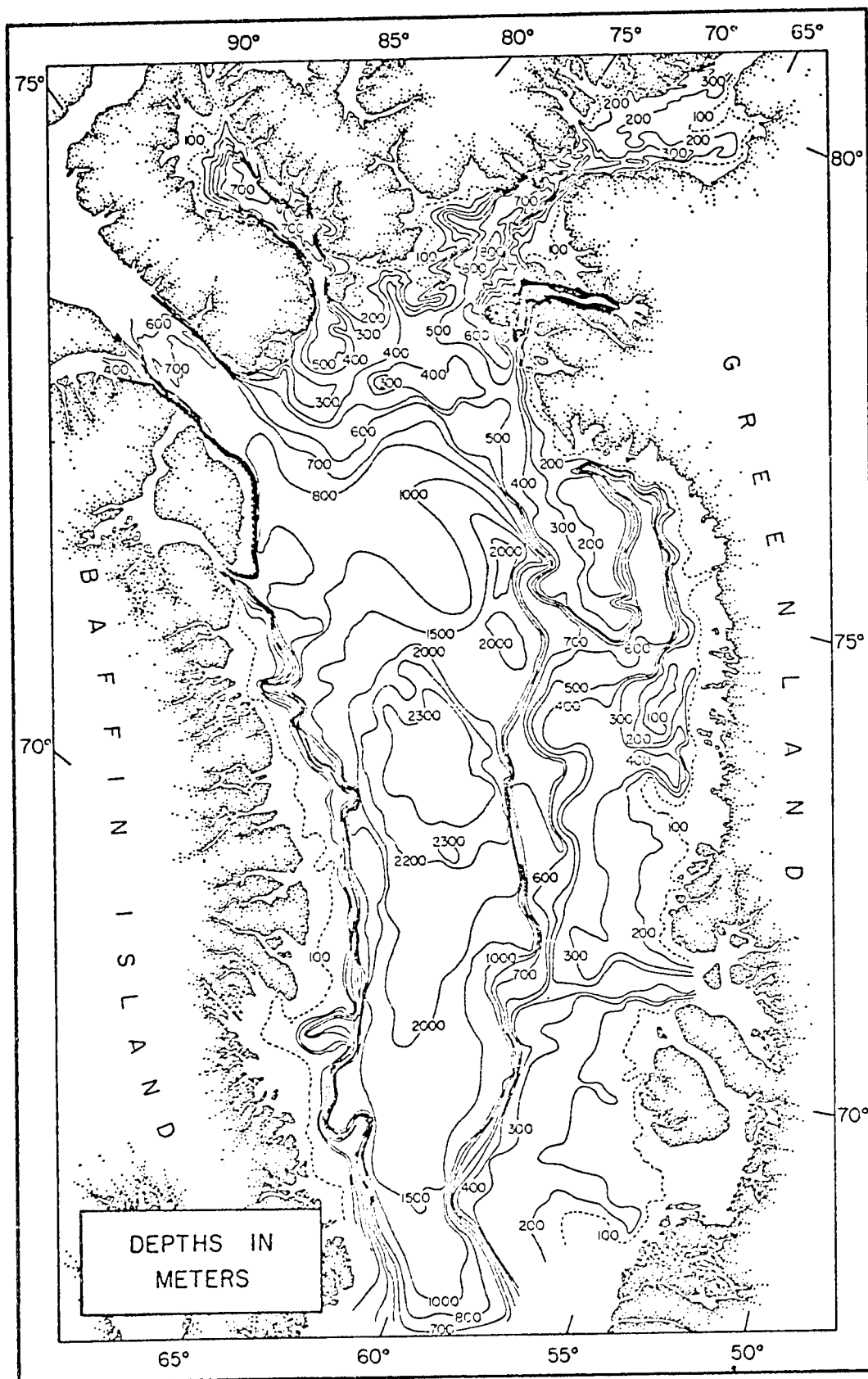


Figure 2. Bathymetry of the Baffin Bay region (after Pelletier 1964). Details were omitted for clarity of presentation.

boundary is more abrupt and abuts a relatively narrow, shallow (less than 200 m) shelf along the Baffin Island coast. In Davis Strait the bottom shoals to a maximum depth of about 675 m.

The deep central basin shoals to the northwest into Lancaster Sound, which is 700–800 m deep at its eastern end. The channels in the Canadian Islands to the west exhibit limiting sill depths of about 180 m.

The bathymetry of Melville Bay is dominated by a central bank, exhibiting depths of less than 200 m, which is skirted on the south, east and north by a relatively deep (more than 600 m), narrow channel. West of this bank, the bottom shoals rapidly northward from 2000 m at about $74^{\circ}30'N$ to 400–500 m in the area between Devon Island and Cape York.

Jones Sound has maximum depths of more than 700 m in its central portion; sills 175 m and 400 m deep occur at its western and eastern ends.

The bottom deepens to 700–800 m in Smith Sound, but shoals in Kane Basin to a sill having a maximum depth of about 250 m. The main channel in Kane Basin is located in its western portion.

The bottom in northern Baffin Bay exhibits extremely rugged and irregular topography, particularly in the Smith Sound region. The sill depths in Smith, Jones and Lancaster sounds and in Davis Strait represent depths below which access of water to Baffin Bay is impossible.

1.3 The oceanographic data

Oceanographic data were obtained from field work carried out from 1916 to 1969 (Appendix A). Three features of the data distribution in space and time are apparent. First, the majority of the northern Baffin Bay data were obtained between July and October. Exceptions are the

relatively shallow stations occupied by Wulff in 1916, as late as November; the stations occupied in May 1969 from KB-2; and the June 1967 Lincoln Sea stations. No oceanographic data have been acquired between December and April. The May, June and November data were obtained during only one year, precluding year-to-year comparisons and making it impossible to determine whether the observed conditions were representative. The May, June and November data are, moreover, areally restricted.

Second, both density and geographical distribution of stations vary from year-to-year. Relatively complete geographical data coverage was obtained only in 1928, 1960 to 1964 and 1966. In 1916 sampling was restricted to the banks off western Greenland. Only one station was occupied in the study area in 1940. Data acquired during 1954, 1956, 1957, 1959 and 1965 were obtained from restricted areas (e.g. Jones and Lancaster sounds) or were from isolated, widely scattered stations. In 1967, data were obtained only from the Lincoln Sea. The 1968 and 1969 WESTWIND data were restricted to the Smith Sound and Melville Bay-Cape York regions, and the 1969 KB-2 data were obtained from a time series of stations occupied at a fixed location in Kane Basin.

Third, stations occupied during different years were not necessarily occupied during the same month(s) of the year, so seasonal variations might be present.

The recorded current data are from the Smith Sound-Cape York region and are relatively short in duration. The shortest record covers a 3-day period and the longest 16 days.

II. THE WATER MASSES

The waters of the northern Baffin Bay region may be divided on the basis of their vertical temperature distribution into three masses: a cold ($<0^{\circ}$ C) upper Arctic Water layer; a warmer ($>0^{\circ}$ C) intermediate depth Atlantic Water layer; and a cold ($<0^{\circ}$ C) Deep Water layer. The 0° C isotherm was selected to divide the water masses because it delineates them approximately according to their origin.

2.1 The Arctic Water

2.1.1. Temperature and salinity distributions in the Arctic Water from vertical cross-sections and profiles

Vertical temperature and salinity cross-sections and profiles have been chosen to represent general subsurface conditions throughout the northern Baffin Bay region (Figs. 3 to 6). Details in the upper 30 m of the cross-sections have been omitted, since the common presence above 30 m of strong vertical temperature and salinity gradients precludes a clear presentation on the scale used. Near-surface features will be discussed where appropriate in conjunction with physical processes.

The vertical temperature structure of the Arctic Water layer is illustrated in Figure 4a. Arctic Water occurs between the surface and 150-600 m. It is characterized, except in the near-surface layers where summer warming may occur, by temperatures below 0° C. A cold core 50-200 m deep exhibits minimum temperatures in the water column and is generally deeper in northwestern than in northeastern or north central Baffin Bay. In northwestern Baffin Bay the cold core is frequently accompanied by a shallower, warmer and less well defined temperature

minimum. The cold core is overlain by a surface layer which exhibits large temperature ranges relative to those within the core.

The salinities (Fig. 4b) are lowest (<33 ‰) at the surface, increase

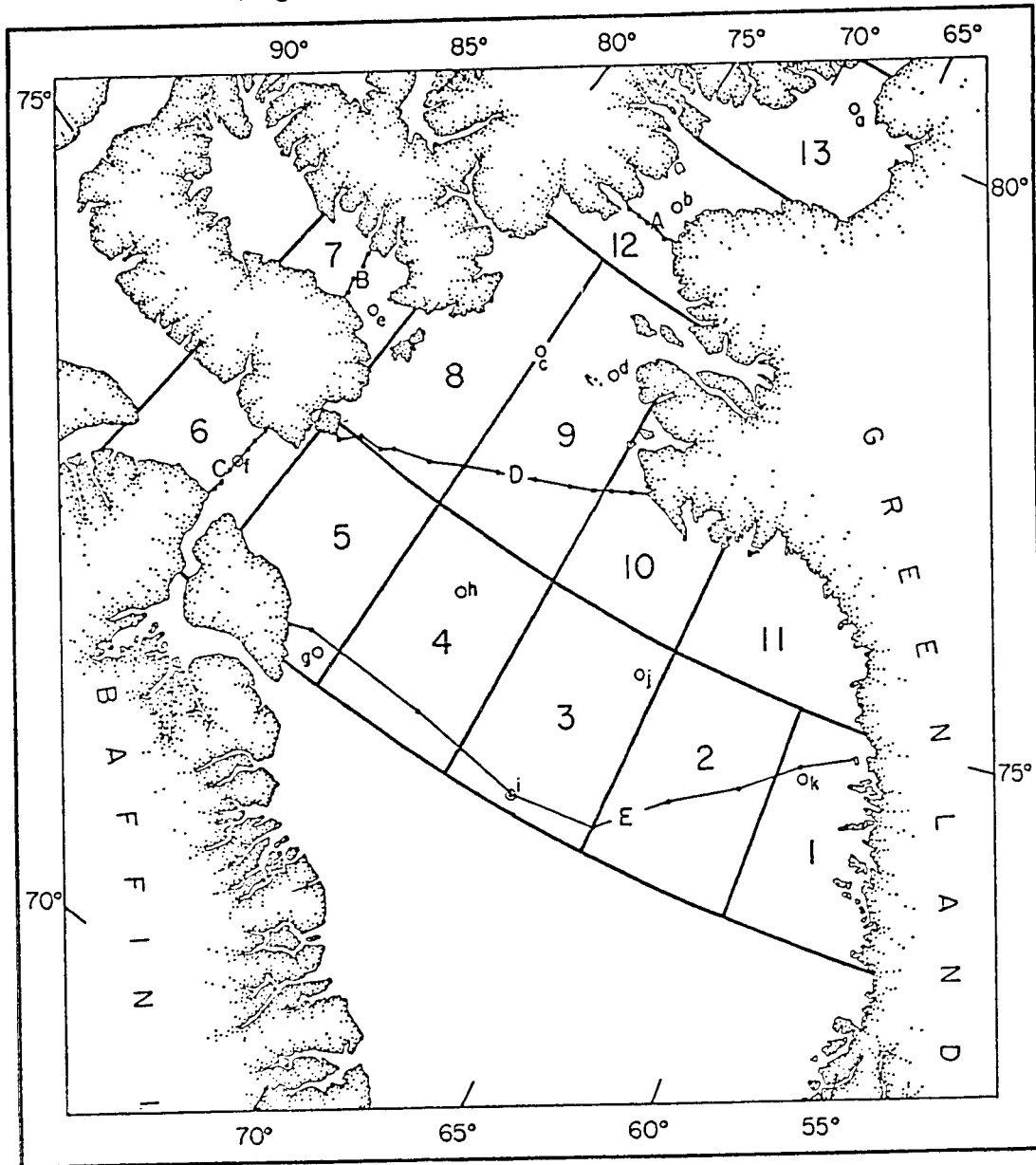


Figure 3. Geographical locations of the profiles shown in Figure 4 (open circles with small letter designations), the cross-sections shown in Figures 5 and 6 (solid light lines with dots indicating station locations and capital letter designations) and the 13 sub-areas used in the statistics.

rapidly with depth to about 34.2 ‰ at 300 m and then less rapidly to 500 m. Salinities are higher at 300 m in Kane Basin (profile (a), Fig. 4b)

than elsewhere in the northern Baffin Bay region. Below 500 m the salinity varies by only about 0.04 ‰ from 34.5 ‰. The Arctic Water layer is therefore the site of a strong halocline and, since salinity exerts the

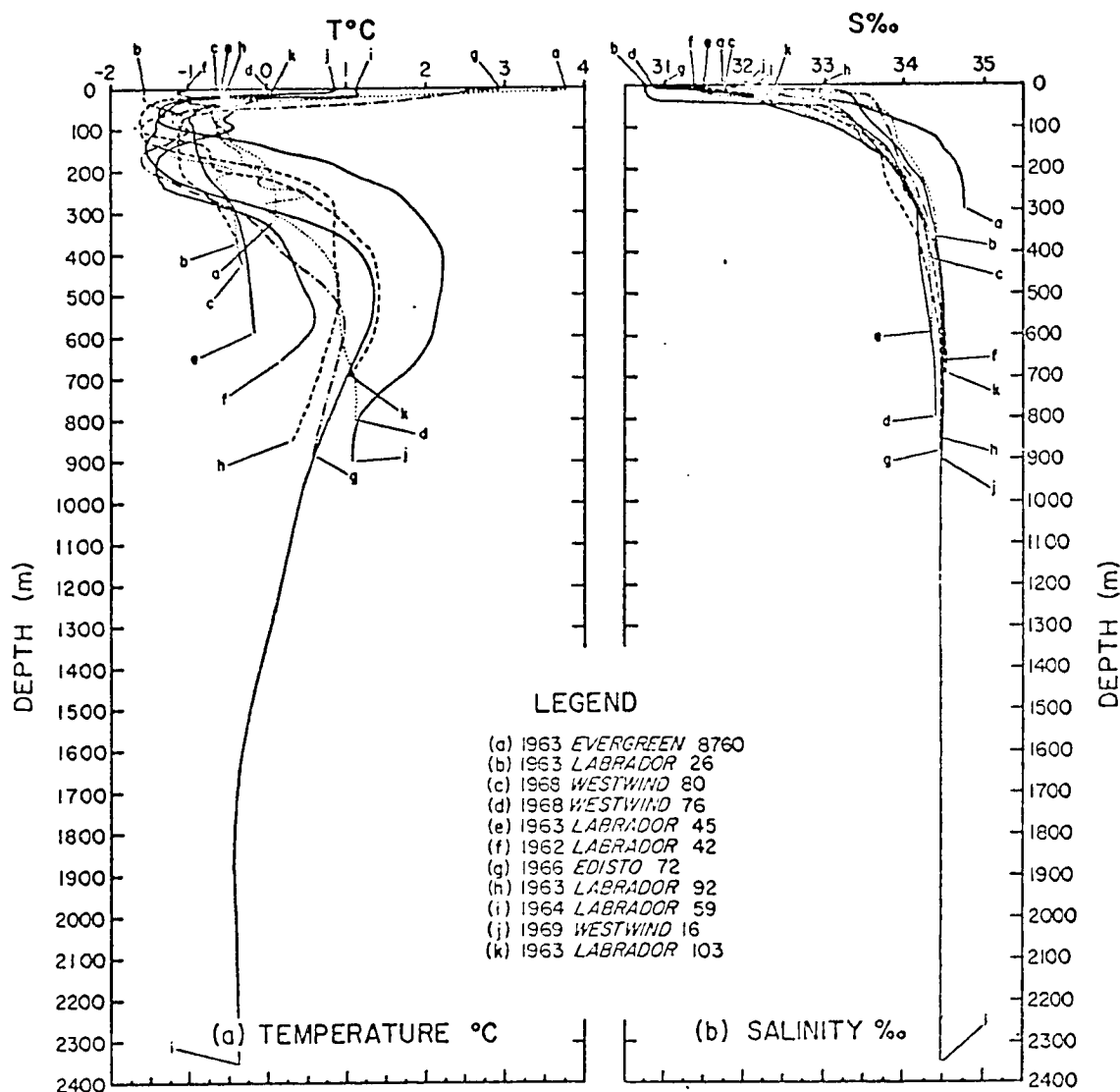


Figure 4. Selected vertical temperature and salinity profiles from the northern Baffin Bay region. Geographical locations of profiles are indicated on Figure 3.

primary control over density at the low temperatures prevailing, a strong pycnocline.

Horizontal temperature variations are indicated on the sections (Fig. 5) and may be deduced from the profiles (Fig. 4a). Horizontal salinity

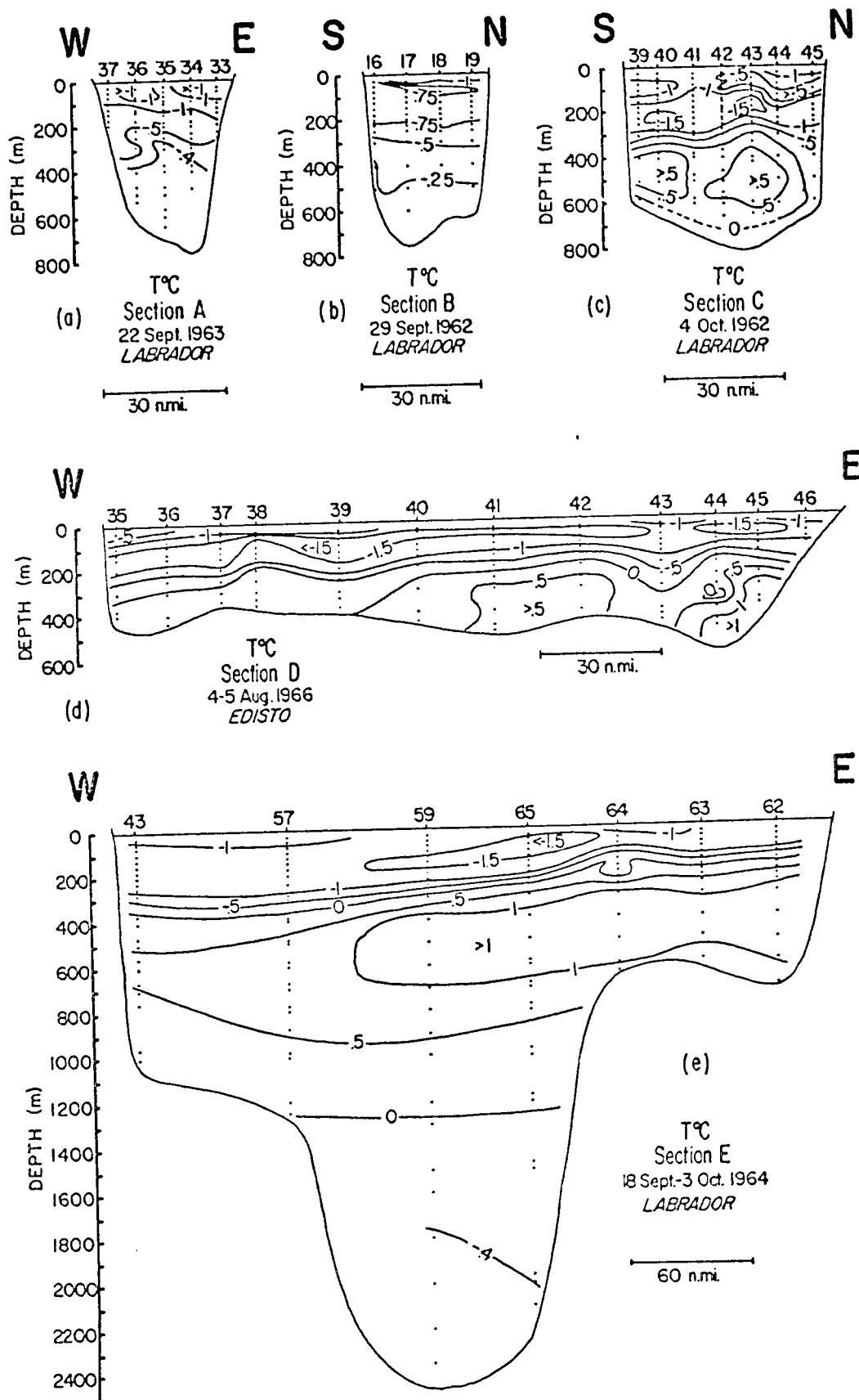


Figure 5. Selected temperature cross sections from the northern Baffin Bay region. Geographical locations of the sections are indicated on Figure 3.

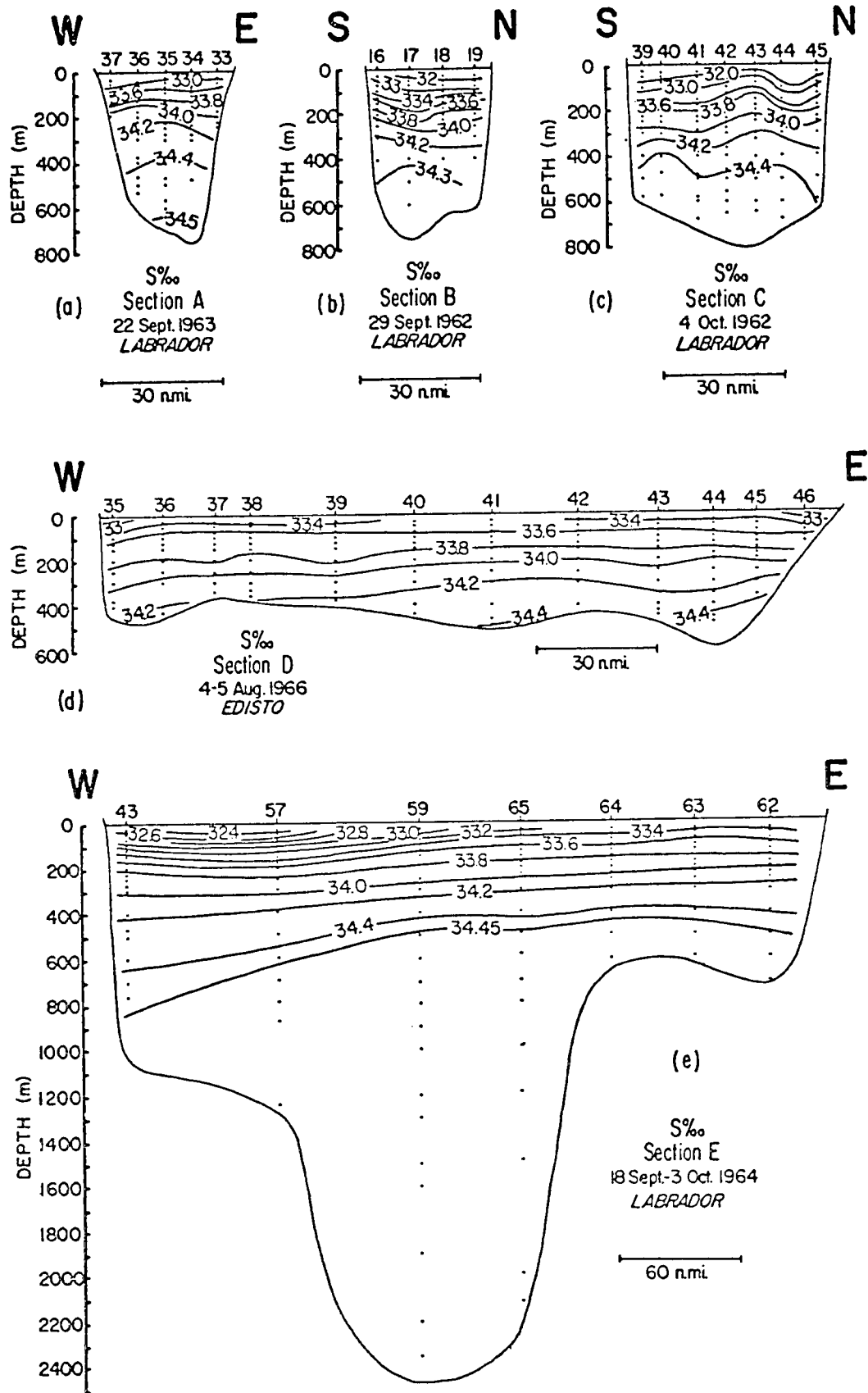


Figure 6. Selected salinity cross-sections from the northern Baffin Bay region. Geographical locations of the sections are indicated on Figure 3.

variations below the near-surface layers are small relative to the temperature variations, and are clearly indicated only on the sections (Fig. 6).

Northern Kane Basin (profile (a), Fig. 4) contained Atlantic Water, below 200 m, the relatively high salinity of which will be seen (section 2.1.3) to indicate that it originated in the Arctic Ocean rather than in Baffin Bay. The minimum temperature in the Arctic Water layer was high relative to minimum temperatures farther south. The relatively high ($>3^{\circ}\text{C}$) surface temperature appeared to be due to summer warming; the data used in constructing this profile were obtained following a period when pack ice was absent from Kane Basin.

Smith Sound (profile (b), Fig. 4; section A, Fig. 5a) contained higher minimum temperatures than occurred farther south. While higher temperatures ($>-1^{\circ}\text{C}$) shallower than 100 m on the cross-section suggest summer warming, low near-surface temperatures (-1.6°C) on the profile suggest cooling. Such near-surface variations, due to climatic effects, are typical for the entire region in summer. The high salinity water noted in Kane Basin was not present in Smith Sound, although the near-bottom salinity of $34.5\text{ }^{\circ}\text{oo}$ was higher than at the same depth farther south. The Arctic Water layer extended throughout the entire water column, i.e. down to nearly 800 m, in Smith Sound.

The channels west and east of the Carey Islands contained vertical temperature variations on a 30-50 m scale (profiles (c) and (d), Fig. 4a). No corresponding vertical salinity variations were detected (Fig. 4b). A relatively low (about -1.3°C) minimum temperature 150 m deep was accompanied west of the Carey Islands by a shallower (60 m), less well defined minimum.

The northern Baffin Bay-Smith Sound region contained a widespread cold core having relatively low ($<-1.5^{\circ}\text{C}$) temperatures (section D, Fig. 5d). The Arctic Water layer thickness was highly variable, and the layer extended much deeper on the west (450 m at stas. 35 and 36) than on the east (200 m at sta. 44). The occurrence at station 44 of a blob of cold ($<0^{\circ}\text{C}$) water at about 400 m was suggestive of the type of vertical variations noted on profiles (c) and (d).

The salinities below 200 m were lower by about $0.2\text{ }^{\circ}/\text{oo}$ in northwestern than in northeastern Baffin Bay (section D, Fig. 6d). Relatively low salinity ($<33\text{ }^{\circ}/\text{oo}$) surface wedges occurred down to about 50 m depth in extreme northeastern and northwestern Baffin Bay.

Jones Sound exhibited relatively high ($>-1^{\circ}\text{C}$) minimum temperatures (profile (e), Fig. 4a; section B, Fig. 5b). A warm ($>0.75^{\circ}\text{C}$) layer was present at 50-100 m depth on section B, but profile (e) indicated a minimum temperature at 75 m. Salinities in Jones Sound (profile e, Fig. 4b; section B, Fig. 6b) were about $0.1\text{-}0.2\text{ }^{\circ}/\text{oo}$ lower than at corresponding depths in Baffin Bay, and were similar above 300 m to those in Lancaster Sound.

Lancaster Sound (profile (f), Fig. 4a; section C, Fig. 5c) was characterized by low ($<-1.5^{\circ}$) and irregular temperatures in the Arctic Water layer, with a deep minimum at 200-250 m and a less pronounced, shallower minimum at 75-100 m. A relatively warm ($>-1^{\circ}\text{C}$) layer 100-200 m thick occurred between the minimum temperature layers. A core of warm ($>0.5^{\circ}\text{C}$) water occurred at 50-200 m depth in the northern part of the section and also appeared in the profile.

The salinities deeper than about 300 m in Lancaster Sound (profile (f), Fig. 4b; section C, Fig. 6c) were similar to those at the same depths

in northwestern Baffin Bay. Salinities above 300 m were similar to those in Jones Sound at identical depths. The surface-to-300 m layer was 0.2 to 0.4 ‰ less saline in Lancaster Sound than in northern Baffin Bay.

The minimum temperature in the Arctic Water layer was generally lower in northeastern and north central Baffin Bay than elsewhere (Fig. 4a; section E, Fig. 5e). A relatively shallow and poorly defined secondary temperature minimum, which occurred in northwestern Baffin Bay, was barely discernable in its north central portion (profile (i), Fig. 4a). The Arctic Water layer appeared relatively thin and shallow in northeastern Baffin Bay. The layer was less saline in the west as indicated by the downward slope, to the west, of the isohalines. A near-surface layer of relatively low salinity (<32.4 ‰) water occurred at station 57 (section E, Fig. 6e).

2.1.2. Horizontal and secular temperature and salinity variations in the Arctic Water

The minimum temperature core is the prominent feature of the Arctic Water layer; its horizontal temperature and salinity variations are therefore of particular interest. Unambiguous plots depicting horizontal temperature and salinity variations in the cold core for each year were impossible due to the sparsity of data. In order to combine all data, the northern Baffin Bay region was divided into 13 sub-areas (Fig. 3). The summer mean values of the parameters discussed below were computed for each sub-area for each year, and were used in turn to compute an overall summer mean value of the parameter in each sub-area. The summer mean ranges were computed in similar fashion. Results, including the number of observations used in each calculation, are given in Table I and Figure 7.

The mean values may be biased by variations of the parameters within the sub-areas. Data distributed uniformly over a given sub-area would be

TABLE I
Mean summer parameters, by area*, for the Arctic Water

| AREA | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
|------------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| \bar{T}_{\min} ($^{\circ}$ C) | -1.16 | -1.44 | -1.60 | -1.57 | -1.52 | -1.46 | -1.14 | -1.35 | -1.40 | -1.41 | -1.04 | -1.25 | -1.39 |
| Range ($^{\circ}$ C) | 0.29 | 0.21 | 0.14 | 0.13 | 0.19 | 0.27 | 0.25 | 0.32 | 0.53 | 0.33 | 0.30 | 0.30 | 0.42 |
| Total no. obs. with min. | 17 | 29 | 34 | 29 | 41 | 78 | 24 | 60 | 159 | 56 | 11 | 70 | 31 |
| No. obs. without min. | 0 | 1 | 0 | 0 | 0 | 1 | 4 | 16 | 8 | 12 | 0 | 20 | 74 |
| \bar{S}_{\min} ($^{\circ}$ /oo) | 33.70 | 33.63 | 33.70 | 33.75 | 33.77 | 33.55 | 33.57 | 33.63 | 33.68 | 33.64 | 33.60 | 33.48 | 33.04 |
| Range ($^{\circ}$ /oo) | 0.11 | 0.15 | 0.12 | 0.11 | 0.12 | 0.57 | 0.62 | 0.40 | 0.34 | 0.21 | 0.17 | 0.56 | 0.68 |
| Total no. obs. | 17 | 28 | 31 | 28 | 39 | 78 | 24 | 60 | 157 | 56 | 10 | 63 | 30 |
| \bar{t} (m) | 80 | 100 | 150 | 140 | 160 | 190 | 220 | 140 | 110 | 110 | 90 | 120 | 100 |
| Range (m) | 50 | 60 | 50 | 60 | 60 | 140 | 160 | 90 | 70 | 70 | 60 | 90 | 90 |
| Total no. obs. | 13 | 26 | 33 | 29 | 37 | 73 | 10 | 48 | 107 | 38 | 4 | 46 | 21 |

*See Figure 3 for locations of numbered areas.

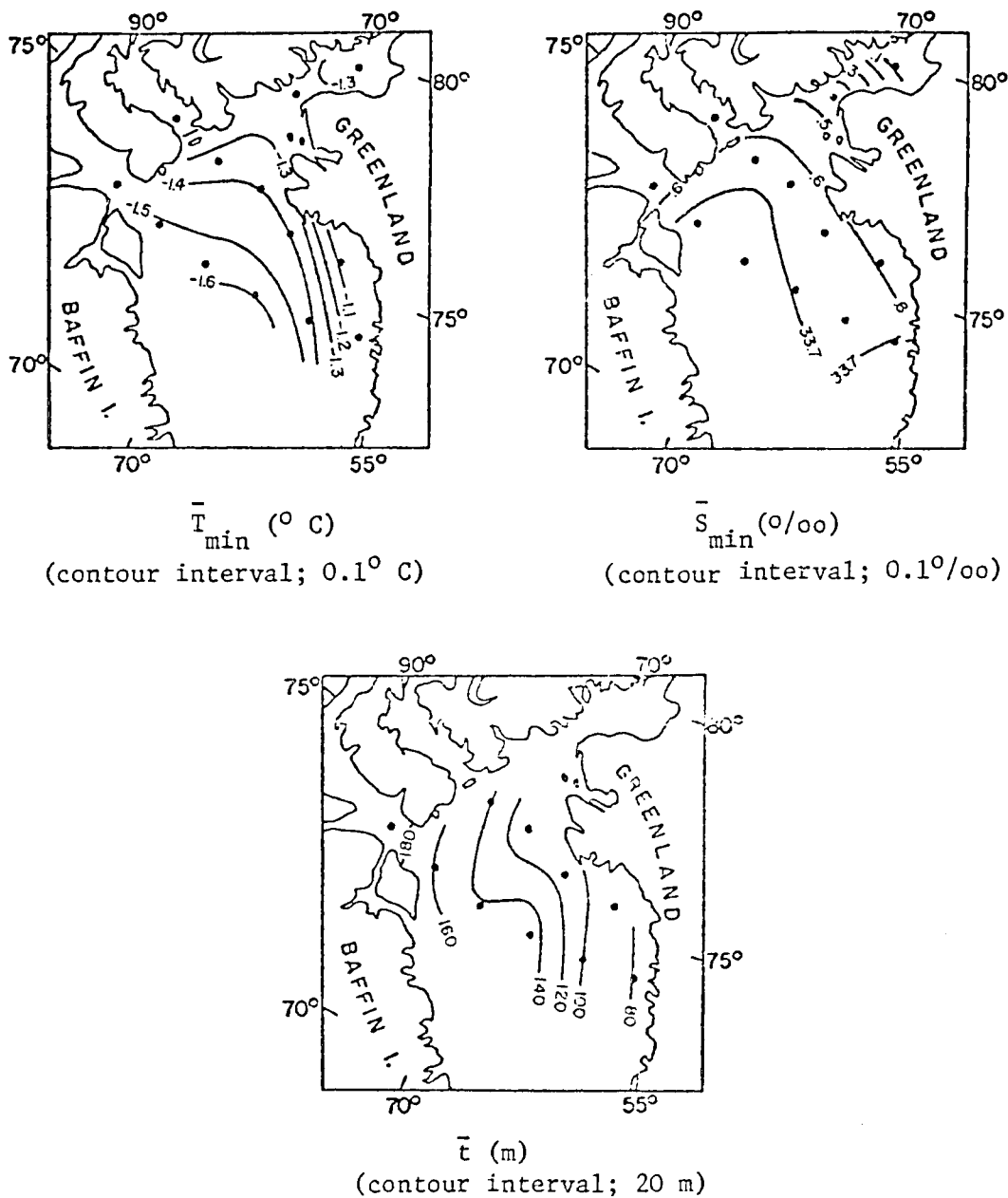


Figure 7. Graphical presentation of \bar{T}_{\min} , \bar{S}_{\min} and \bar{t} from Table I.

expected to yield a more representative mean for that area than would data from a small portion of the area, particularly where large horizontal variation occurred within the area. The mean values may, in addition, be biased by time variation within the area. It is felt that,

by combining data from different years, reliable means were obtained with the exceptions noted below.

The lowest values of \bar{T}_{\min} , the summer mean minimum temperature, occurred in north-central and northwestern Baffin Bay (areas 3-5). \bar{T}_{\min} in Lancaster Sound (area 6) was similar to that in northwestern Baffin Bay. Higher values occurred in Jones and Smith sounds (areas 7 and 12), in Kane Basin (area 13) and in the northeastern peripheral areas (areas 1 and 11). Southern Smith Sound (areas 8-10) exhibited values of \bar{T}_{\min} intermediate between the values to the north and south. The relatively small number of observations used in computations for areas 1 and 11 suggests that less confidence be placed in the values for these than for other areas.

Maximum variability of T_{\min} , as suggested by the ranges, occurred in southern Smith Sound (area 9) and Kane Basin (area 13), while the least variability occurred in northwestern and north-central Baffin Bay (areas 3-5). Values for other areas were intermediate between these two extremes. The regions of lowest variability coincided with those of lowest \bar{T}_{\min} . The large variability in area 9 will be seen (section 2.1.3) to reflect mixing between Arctic Ocean and Baffin Bay water.

The number of observations showing no minimum in the water column indicated that a clearly defined cold core was generally present in Baffin Bay, but in Smith Sound and Kane Basin (areas 12 and 13) the water column was more commonly uniformly cold without having a core. (See, e.g., the envelope of vertical temperature profiles from Kane Basin in Figure 11.)

The variations in \bar{S}_{\min} , the summer mean salinity at the cold core, throughout northern Baffin Bay (areas 1-5 and 8-11) were small and did not follow any discernable pattern. \bar{S}_{\min} was slightly lower in Lancaster,

Jones and Smith sounds (areas 6, 7 and 12) than in Baffin Bay. The lowest values occurred in Kane Basin (area 13).

The horizontal variability of S_{\min} was small and relatively constant from area-to-area in northern central Baffin Bay (areas 1-5 and 11). Greater variability occurred in Lancaster, Jones and Smith sounds (areas 6-12) and in Kane Basin (area 13).

In order to estimate the horizontal volume distribution of Arctic Water, the relative summer mean thickness \bar{t} of the Arctic Water layer was computed for each of the sub-areas. The -1°C rather than the 0°C isotherm was chosen to bound the Arctic Water for this calculation, since in many cases the 0°C isotherm was only present at the upper or lower surface of the Arctic Water layer and layer thickness using that criteria would have been indeterminate. The choice of bounding isotherm does not appreciably affect the relative layer thickness in different areas. The computations of \bar{t} will be discussed only as they apply to northern Baffin Bay and Lancaster Sound, excluding Smith and Jones sounds and Kane Basin. The latter three are filled with Arctic Water, and the relatively high minimum temperatures ($>-1^{\circ}\text{C}$) might cause computations based on the -1°C isotherm as a boundary to be misleading.

\bar{t} was largest in northwestern Baffin Bay and Lancaster Sound (areas 3-6). The northeastern areas (1, 2, 10 and 11) were characterized by thinner layers. The northeastern peripheral value (area 11) is based solely on the 1928 data and therefore less reliable than values for other areas. The horizontal variability of t was generally large relative to \bar{t} , and reached maximum values in Lancaster Sound (area 6). The areas characterized by thinner layers were also characterized, in general, by lower \bar{T}_{\min} and higher S_{\min} .

It was desirable to detect any year-to-year changes, if any, in T_{\min} , S_{\min} and t . Temperature and salinity changes at the cold core are assumed indicative of changes throughout the Arctic Water layer on the basis that vertical eddy diffusion would be expected to transmit any variations throughout the layer on the long time scales under consideration. To detect changes, sets of values of the parameters from sub-areas 1-5 and 10 (areas of minimum variability) for different years were assumed to represent different statistical populations and compared as pairs. The so-called "sign test" (see, e.g., Hoel 1962) was used because it does not presuppose knowledge of the statistical distribution of the parameters. Differences detected by this method are generally real, but due to lack of assumptions about distributions and the effects of geographical variation it may fail to detect a real difference. There was no evidence, in this case, to suggest differences other than those detected by the test.

The results are presented in Table II. The only significant changes were decreases in temperature and salinity sometime between 1964 and 1966. The large per cent significance levels accompanying the other changes suggest that they are not significant. The temperature change between 1964 and 1966 will be shown (section 2.1.3) to reflect seasonal changes. It is concluded, for purposes of the present study, that conditions in the Arctic Water layer have undergone no net changes between 1928 and the 1960's.

TABLE II

Late summer variations in temperature, salinity and thickness
of the cold core for various years

T_{\min}

During period in column A, core was (colder than)(warmer than)(of the same temperature as) during period in row B.

| A \ B | 1961 | 1962 | 1963 | 1964 | 1966 |
|-------|------|------|------|------|------|
| 1928 | C-11 | C-35 | C-35 | C-35 | W-31 |
| 1961 | | C-35 | S | S | W-10 |
| 1962 | | | C-35 | S | W-10 |
| 1963 | | | | C-35 | W-10 |
| 1964 | | | | | W-10 |

Key: C-colder than. W-warmer than. S-of the same temperature as.
Numbers following C, W or S indicate per cent significance level of temperature change as determined from the sign test.

S_{\min}

During period in column A, core was (more saline than)(less saline than)(of the same salinity as) during period in row B.

| A \ B | 1961 | 1962 | 1963 | 1964 | 1966 |
|-------|------|-------|------|------|------|
| 1928 | L-11 | L-2.5 | L-35 | L-35 | H-10 |
| 1961 | | L-35 | S | H-5 | H-10 |
| 1962 | | | H-11 | H-11 | H-10 |
| 1963 | | | | H-11 | H-10 |
| 1964 | | | | | H-10 |

Key: L-less saline than. H-more saline than. S-of the same salinity as.
Numbers following L, H or S indicate per cent significance level of salinity change as determined from the sign test.

t

During period in column A, core was (thicker than)(thinner than)(of the same thickness as) during period in row B.

| A \ B | 1961 | 1962 | 1963 | 1964 | 1966 |
|-------|------|------|------|------|------|
| 1928 | T-35 | N-35 | T-35 | N-35 | T-31 |
| 1961 | | N-35 | T-35 | N-35 | S |
| 1962 | | | T-35 | S | T-31 |
| 1963 | | | | N-19 | S |
| 1964 | | | | | T-10 |

Key: T-thicker than. N-thinner than. S-of the same thickness as.
Numbers following T, N or S indicate per cent significance level of thickness change as determined from the sign test.

2.1.3. The origins of the Arctic Water

Both Arctic Ocean Water and the waters south of Davis Strait have access to Baffin Bay and are possible sources of Baffin Bay Water. These regions are represented by the temperature-salinity curves in Figure 9, the numbers of which correspond to the areas shown on Figure 8. The curves generally fall within well defined envelopes below the relatively variable (in the Baffin Bay region) upper 0-to-200 m layer. Lancaster Sound is an exception; the relatively large temperature scatter at 500-600 m is due to inclusion of the 1954 data, which indicated abnormally high temperatures at those depths. The 1966 data were not included, since temperature-salinity analyses were carried out on these by Palfrey and Day (1968). Table III lists the data used in constructing Figure 9.

The individual temperature-salinity curves (Fig. 10) are visually determined representations of the densest portions of the families of curves in Figure 9, and will be seen below to indicate gross relations between the water masses.

The relatively high salinity (32-33 ‰) of the 0-50 m layer in northeastern Baffin Bay (area 9) compared to Arctic Ocean (areas 1 and 2) near-surface water (30-32 ‰) suggests that the northeastern Baffin Bay Arctic Water is primarily advected northward from eastern Davis Strait (area 10) by the West Greenland Current. This was also suggested by Palfrey and Day (1968) on the basis of the 1966 data. The temperature of this layer is, however, several degrees lower than that of the corresponding layer in eastern Davis Strait. Computations were carried out assuming annual formation of 2 m ice in northeastern Baffin Bay (judged a reasonable thickness on the basis both of observations and ice potential calculations)

and indicated that brine released during the freezing process is sufficient to cause downward convection to about 150 m. The resulting uniform, cold layer would have a salinity of about 33.7 ‰ and a temperature of about

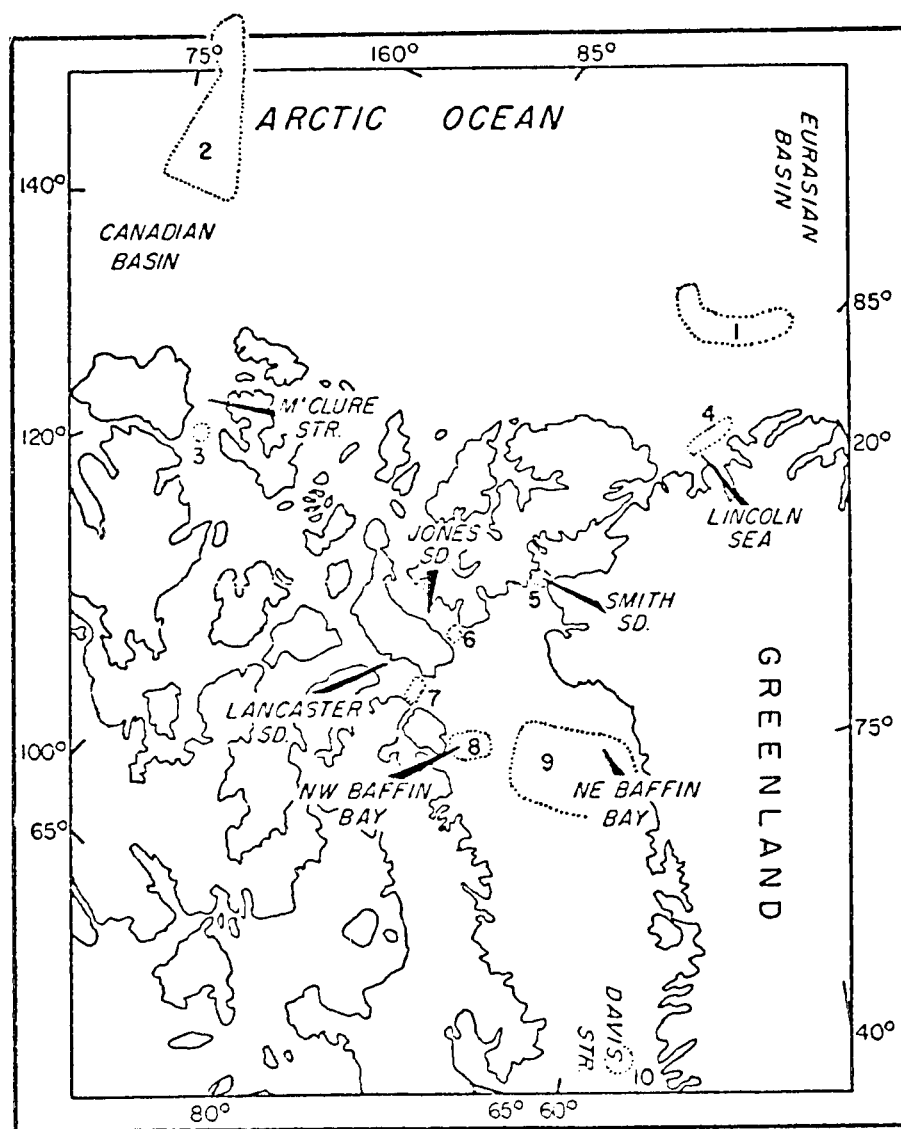


Figure 8. Geographical locations of water masses represented by the temperature-salinity curves in Figures 9, 10, 12 and 17. Numbered areas delineated by dotted lines correspond to the numbered temperature-salinity curves on those figures.

-1.83°C (the freezing point at the specified salinity), and would extend from the surface approximately to the level of the bottom of the cold core

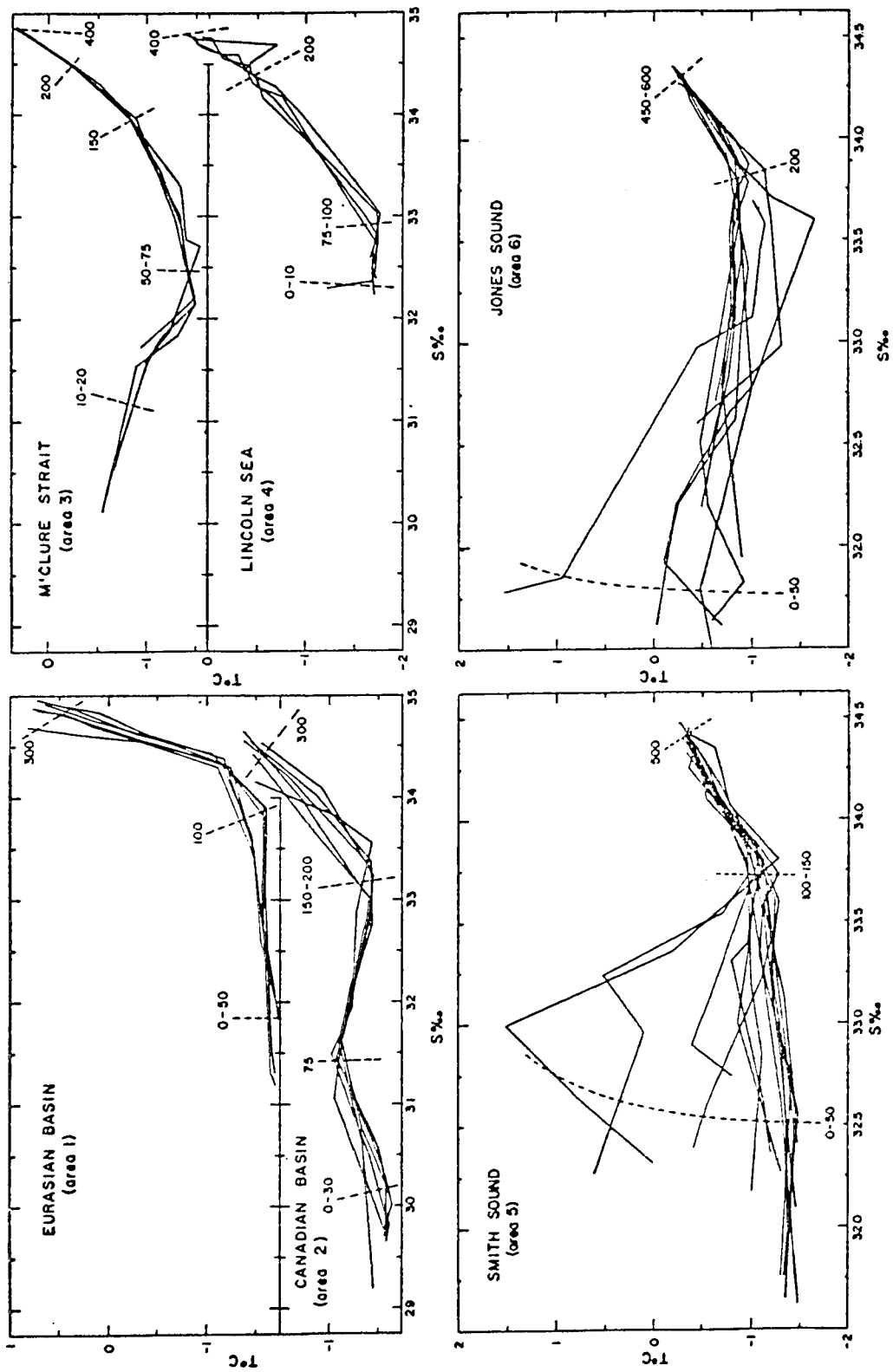


Figure 9a. Temperature-salinity curves for areas 1-6 delineated on Figure 8. Dashed lines and numbers indicate depths in meters.

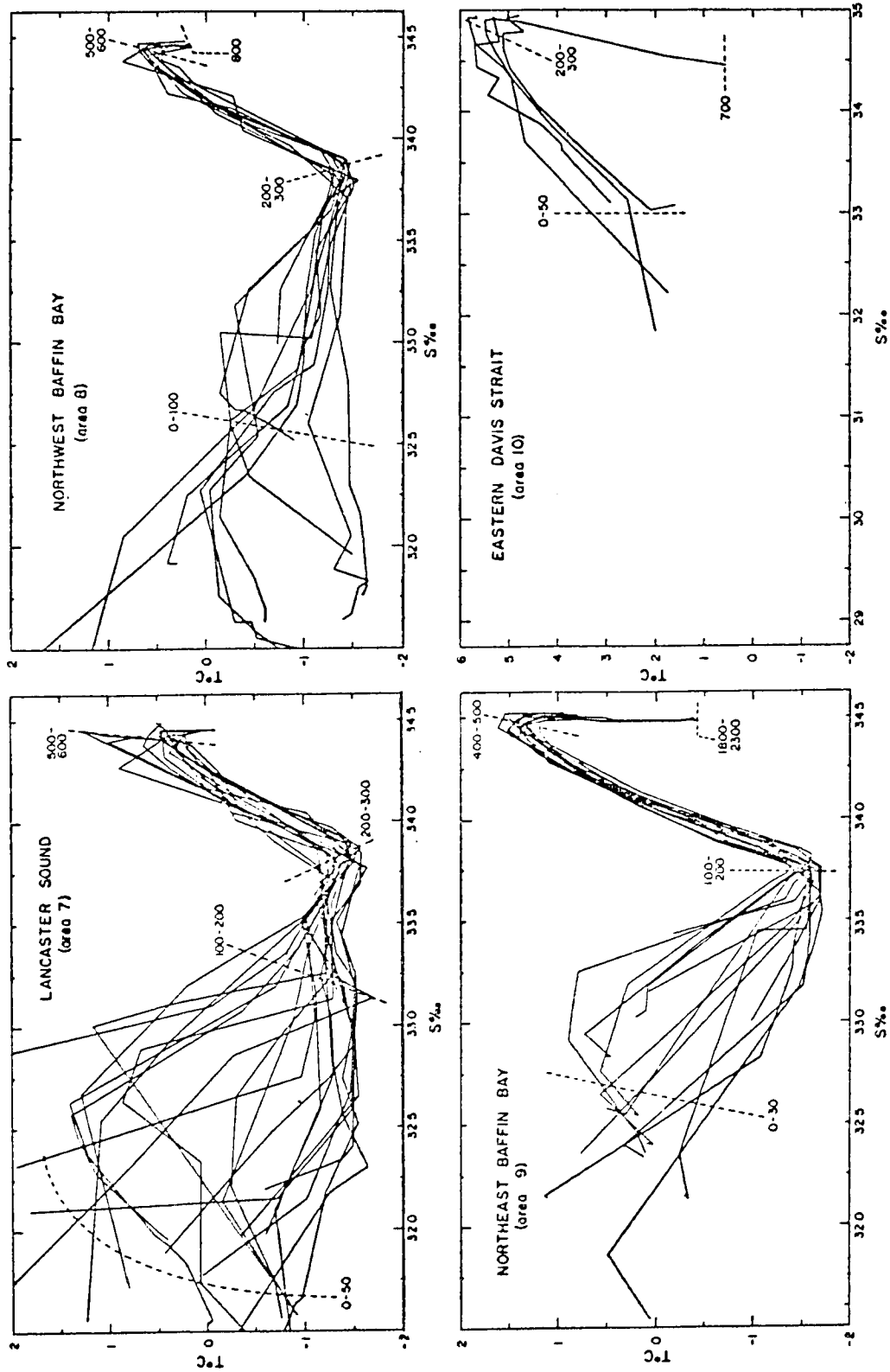


Figure 9b. Temperature-salinity curves for areas 7-10 delineated on Figure 8. Dashed lines and numbers indicate depths in meters.

TABLE III

Data used in compiling the temperature-salinity
curves in Figures 9, 10, 12 and 17

| AREA | PLATFORM | YEAR | STATIONS | MONTHS |
|------|--|--------------------------------------|---|--|
| 1 | ARLIS-II | 1964 | 1, 25, 55, 85, 109, 138, 171 | February-September |
| 2 | T-3 | 1965 | 1, 5, 35 | July, October, December April, June, September |
| 3 | LABRADOR | 1961 | 9, 11, 14, 16 | August-September |
| 4 | pack ice | 1967 | 5, 9, 12, 13 | June |
| 5 | LABRADOR LABRADOR LABRADOR LABRADOR | 1954 1962 1963 1964 | 22, 23, 24, 25 2, 3 29, 30, 31, 32 14, 15, 16, 17 | August September September September |
| 6 | LABRADOR LABRADOR LABRADOR | 1954 1962 1963 | 15, 16, 17 13, 14, 15 44, 45, 46, 47 | August September September |
| 7 | LABRADOR LABRADOR LABRADOR LABRADOR LABRADOR | 1954 1957 1960 1961 1962 | 9, 10, 11, 12 138, 139, 140, 141 52, 53, 54, 55 61, 62, 63, 64 32, 34, 36, 38 | August September September September September |
| 8 | LABRADOR LABRADOR LABRADOR LABRADOR | 1960 1961 1962 1963 | 57, 58, 59 79, 80, 81 51, 52 94, 96, 98 | September September October October |
| 9 | LABRADOR LABRADOR LABRADOR LABRADOR | 1961 1962 1963 1964 | 91, 92 57, 58 100-105 55, 59, 60-65 | September October October September-October |
| 10 | LABRADOR LABRADOR LABRADOR | 1962 1964 1965 | 71 84 116, 117 | October October October |

For data sources and cruise reference numbers, see Appendix A.

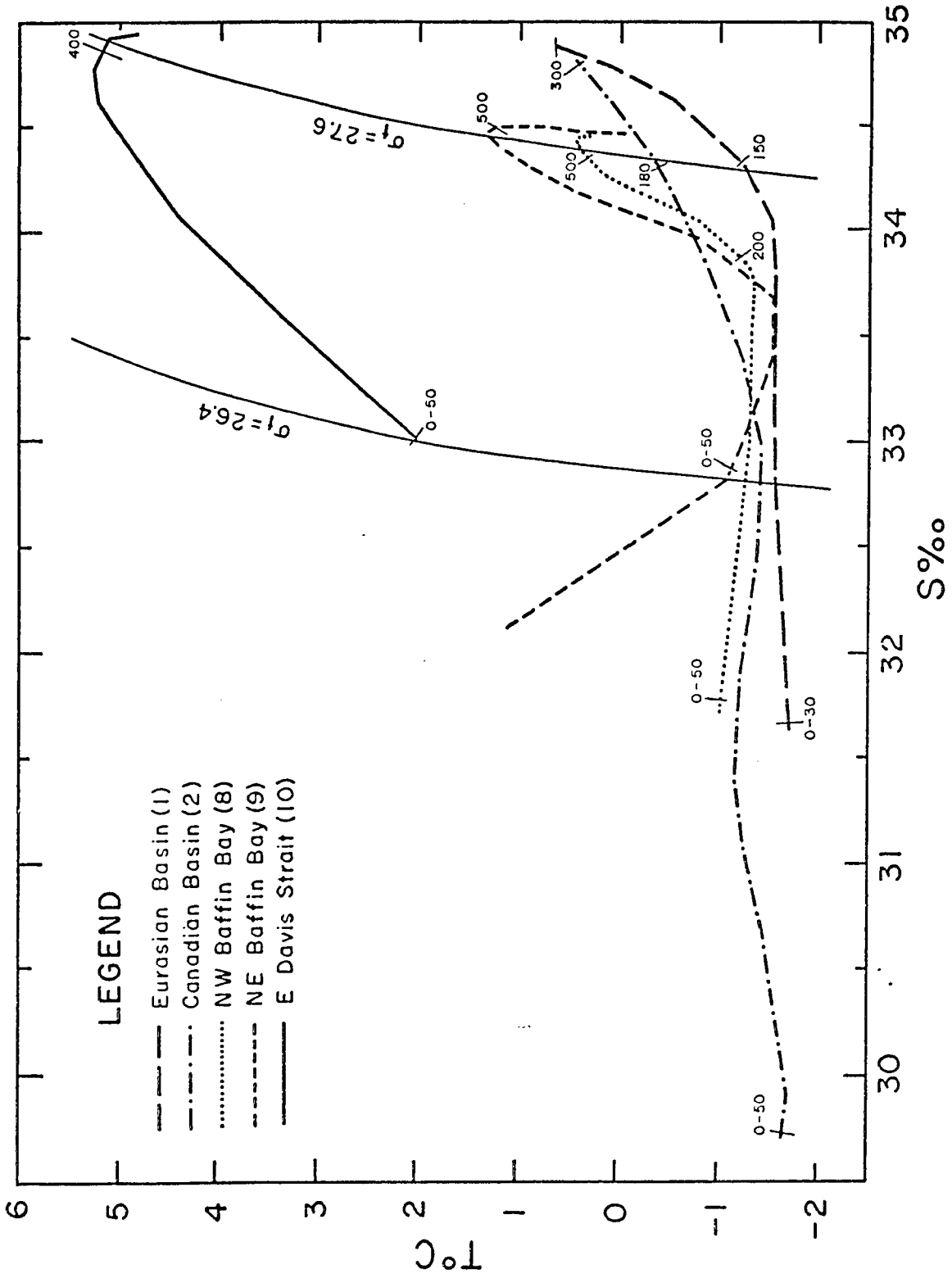


Figure 10. Temperature-salinity curves for Davis Strait, Baffin Bay and the Arctic Ocean, showing gross relations. Parenthesized number in the legend refer to areas delineated on Figure 8.

observed in late summer. Such a mechanism could account for the low temperature of the upper layer in northern Baffin Bay relative to that farther south and hence for the presence of the cold core.

Below the cold core, mixing with the warmer ($>0^{\circ}$ C) underlying Atlantic Water results in formation of a transition layer. Redfield and Friedman (1970) concluded, on the basis of deuterium-salinity ratios, that the effects of brine production due to surface freezing were detectable down to 450 m in Baffin Bay, or nearly to the warm core of the Atlantic Water. This suggests that the effects of mixing with the cold surface layer, directly enriched by brine from freezing, might extend down to that depth.

The low salinity (<32 ‰) of the near-surface layer in northwestern (area 8) compared with northeastern (area 9) Baffin Bay suggests an admixture of Arctic Ocean Water in the former location. This relatively low salinity near-surface layer is apparently a characteristic feature of the south-flowing Baffin Current, which occupies western Baffin Bay and whose waters originate primarily from the Arctic Ocean via Smith, Jones and Lancaster sounds. The different sources of the near-surface layers in northwestern and northeastern Baffin Bay are therefore responsible for the small salinity differences which result in the downward slope of isohalines to the west.

Before examining more detailed temperature-salinity relations in northern Baffin Bay, some consideration of seasonal variations is appropriate. Convection, due to winter ice formation, in northeastern Baffin Bay was suggested above to extend down to about 150 m and result in the formation of a cold, uniform layer above that depth. The summer formation of a relatively warm ($>-1^{\circ}$ C), low salinity (<33 ‰) surface layer by addition of meltwater combined with solar heating in northeastern

Baffin Bay isolates the lower portion of the winter-formed cold layer forming the observed cold core. That this cold core is not supplanted, in summer, by warmer water advected northward from Davis Strait suggests that the West Greenland Current is relatively sluggish. This is borne out (section 3.1) by the dynamics. The cold core would be expected to decay somewhat during the summer due to inward diffusion of heat from above and below. While insufficient data are available to fully document such a decay, the 1966 July-August data indicated lower minimum temperatures in northern Baffin Bay than did September-October data from the same areas but from other years.

Seasonal temperature and salinity variations in western Kane Basin have been partially documented (Fig. 11). Data obtained in May 1969 before the onset of warm weather and accompanying melting are felt to represent winter conditions; the common occurrence of supercooled water in the 0.50 m layer suggested, in fact, that sea ice was still forming. The September station represents late summer conditions.

Temperature, assumed to be a better indicator than salinity of downward convection (Aagaard and Coachman 1968) suggested occurrence of convection down to about 100 m in May. Summer warming was apparent down to about 130 m in September, although the relatively high temperature and salinity variability in Kane Basin (Table I) suggests that part of this difference may be a consequence of geographic variability rather than being purely a seasonal effect.

Seasonal temperature variations in the Arctic Ocean's near-surface layer are negligible (Fig. 9a). Heat absorbed through the surface in the summer melts ice, changing the salinity but not the temperature; since some ice cover is always present in the Arctic Ocean, no heat is available

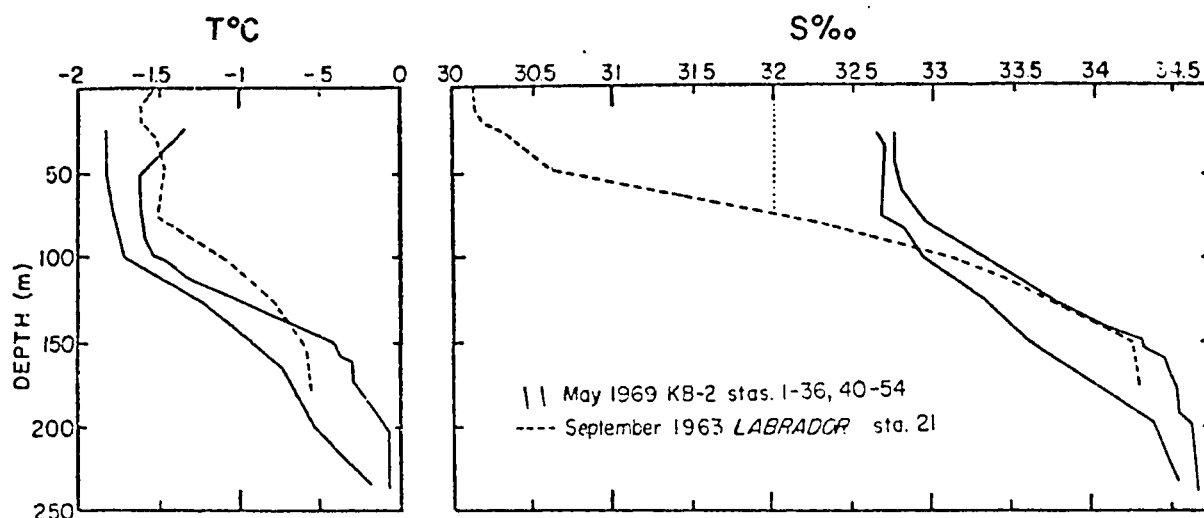


Figure 11. Observed seasonal (May to September) temperature and salinity variations in western Kane Basin. The dotted line indicates a boundary to the left of which the salinity deficit may be accounted for by fresh water runoff and to the right of which a salinity excess may be accounted for by winter brine production due to freezing (see text).

for warming the surface water. Assuming an Arctic Ocean origin, the observed temperature changes in Kane Basin water must be due to admixture in Nares Strait of relatively warm meltwater coupled with a downward surface heat flux.

The seasonal salinity variations in Kane Basin were large relative to the temperature variations and extended down to about 100 m depth. A near-bottom winter salinity increase of 0.6 ‰ in Kane Basin, suggested by Redfield and Friedman (1970) on the basis of deuterium-salinity ratios, was not evident. The Lincoln Sea and Eurasian Basin temperature-salinity curves (Fig. 9a) indicate near-surface salinities of about 32 ‰ , the Lincoln Sea Water (in June) being 1.0 ‰ more saline than the Eurasian Basin Water (over a February-September interval). The Canadian Basin exhibited slightly lower salinities and a relatively small, poorly documented seasonal change. Seasonal salinity variations, in the Arctic Ocean source water for Nares Strait and Kane Basin are therefore insufficient to

account for the observed salinity variations in Kane Basin.

Calculations reveal that winter formation of about 2 m ice in Kane Basin (a reasonable thickness on the basis of observations, made during spring 1968 and spring 1969 from the Kane Basin pack ice, and on ice potential calculations) is sufficient to raise the salinity of the 0-100 m layer from 32 to 32.7 ‰. Brine production due to freezing can therefore account for the excess salinity represented by the area to the right of the dotted line on Figure 11. The cold, uniform layer formed by such winter convection is too fresh (about 32.7 ‰) to contribute to the cold core (which has a mean salinity of about 33.7 ‰ in Baffin Bay. The layer is seen rather as a poorly defined, shallow cold core at 50-100 m on the vertical temperature profiles in western Smith Sound, Lancaster Sound and northwestern Baffin Bay (Fig. 4a).

The salinity deficit in late summer relative to the assumed base salinity of 32 ‰ (represented by the region to the left of the dotted line on Figure 11) may be accounted for approximately by terrestrial runoff from the Greenland and Ellesmere Island coastal regions. Annual precipitation figures for Greenland from Bader (1961) were, in the absence of more accurate data, extrapolated to Ellesmere Island in a first approximation to total fresh water runoff. Calculations assuming a southward surface current speed in Nares Strait of 15 cm/sec, based on geostrophic calculations and current measurements (section 3.1), suggested that runoff would be sufficient to maintain the salinity deficit for 2-3 months. Due to the crudeness of both the precipitation data and knowledge of the flow through Nares Strait, these figures can be regarded only as estimates.

Seasonal temperature and salinity variations in Jones and Lancaster sounds have not been documented. Neither annual precipitation in the

Canadian Islands nor current patterns through them are sufficiently well known to allow an estimate of the effect of runoff. Redfield and Friedman (1970) have suggested, on the basis of deuterium-salinity ratios, that the presence of summer meltwater is evident as deep as about 200 m in Lancaster Sound. The effect of winter ice formation, based on the characteristics of Arctic Ocean (Canadian Basin) water, is suspected to be similar to that in Kane Basin with a convective layer extending downward to about 100 m.

Further temperature and salinity variations occur within northern Baffin Bay itself. The temperature-salinity curves in Figure 12 are visually determined representations of the densest portions of the families of curves in Figure 9. Prior to considering temperature-salinity relations within northern Baffin Bay, however, the probable source waters must be more clearly specified.

A comparison between Lincoln Sea Water (area 4) and Arctic Ocean Water (areas 1 and 2) suggests that Lincoln Sea Water between 100 and 300 m is a mixture of Canadian and Eurasian basin waters, the former being predominant. Lincoln Sea Water above 100 m appears to have been derived from the near-surface layer of the Eurasian Basin; its uniformly low temperature ($<-1.5^{\circ}\text{C}$) and relatively high salinity ($>32.25\text{‰}$) suggest modification by local cooling and brine formation. Such modification would be expected to be seasonal, however, and is observed here in June (i.e. late winter) data. Reliable late summer Lincoln Sea data were not available.

These conclusions regarding the origin of Lincoln Sea Water agree with those of Dunbar, Dunbar and Nutt (1967), who suggested that source water north of Nares Strait was a combination of Canadian and Eurasian basin water. Similar conclusions were drawn by Seibert (1968), who

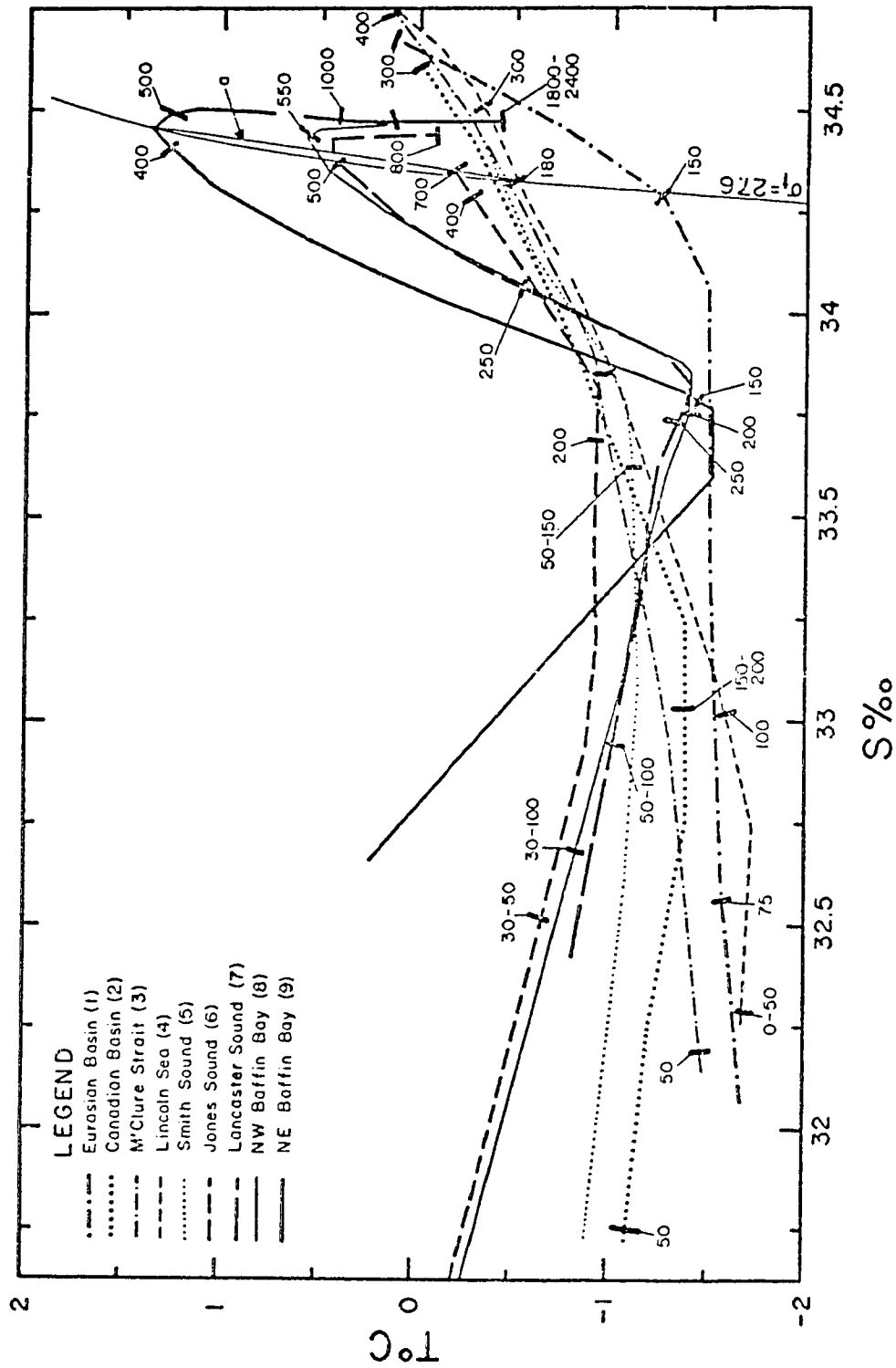


Figure 12. Temperature-salinity curves for the northern Baffin Bay region. Numbers on the legend correspond to the areas delineated on Figure 8. Ticks with numbers on the temperature-salinity curves indicate depths in meters.

suggested on the basis of the 1967 Lincoln Sea data that water there was a mixture, modified by vertical turbulence, of Eurasian Basin sub-surface water, Lincoln Sea surface water and (Arctic Ocean) Atlantic Intermediate Water.

M'Clure Strait (area 3; see Fig. 8) water below about 150 m is similar to Lincoln Sea Water, suggesting a similar origin. Above 150 m, 0.1 to 0.4° C higher temperatures suggest a greater proportion of Canadian Basin Water than was apparent in the Lincoln Sea, and probably some summer warming since the M'Clure Strait data were obtained during the late summer. Additional data are required to further clarify the origins of M'Clure Strait Water.

The net flow through Smith, Jones and Lancaster sounds (areas 5-7) as deduced both from the presence of Arctic Ocean Water in Baffin Bay and from the dynamics (section 3.1), is towards Baffin Bay. Mixing of this inflowing Arctic Ocean Water with 0.5° C colder Baffin Bay Water results in the marked attenuation of the Baffin Bay cold core in northwestern Baffin Bay (area 8) and Lancaster Sound (area 7). The fact that this attenuation appears more marked in northwestern than northeastern Baffin Bay suggests that the inflowing Arctic Ocean Water is primarily advected southward by the Baffin Current and is not present in appreciable quantities in northeastern Baffin Bay. The upper 100-150 m layer in Lancaster Sound and northwestern Baffin Bay appears to be, in fact, Arctic Ocean Water which is practically unmodified save for some local heating and fresh water admixture near the surface.

Water below 100-150 m in Smith Sound is identical in temperature and salinity (-0.3 to -1.2° C and 33.75 to 34.4 ‰) to Lincoln Sea Water, further suggesting a net southward flow via Nares Strait of water from

the Lincoln Sea. A similar conclusion was arrived at by Muench (1966) based on data through 1964 and also by Palfrey and Day (1968) based on the 1966 data. This southward flow explained the presence in Kane Basin of the relatively high salinity water found deeper than about 100 m.

Palfrey and Day (1968) also noted the presence in Smith Sound of a cold core at 100-150 m depth having a salinity of about 33.7 ‰. They attributed the presence of this core to southward flow from the Lincoln Sea, whereas it was more probably a result, due to its high salinity, of a northward pulse of Baffin Bay Water into Smith Sound prior to the acquisition of the data. Such northward pulses in Nares Strait are suspected to be common (sections 3.1 and 3.2).

Water below 200 m in Jones Sound is similar to that at shallower depths in the Lincoln Sea and M'Clure Strait. A 0.2 to 0.3° C temperature elevation below about 250 m, relative to Canadian Basin Water, suggests some admixture of Baffin Bay Atlantic Water into Jones Sound. The uniformity of the water between 450 and 700 m (-0.2° C and 34.35 ‰) reflects the 450 m limiting sill east of Jones Sound. Recalling that the sill depth to the west is only 180 m, the bathymetry in Jones Sound forms a cul-de-sac for water deeper than 450 m and thus which can communicate only with the overlying water via vertical motions. The water at 600-700 m in Jones Sound falls approximately along a straight line (marked "a" on Figure 12) between water at 450 m in Baffin Bay and water at 180 m in M'Clure Strait, arguing for its formation from mixing of these two source waters and further corroborating presence of the sills to the east and west. Comparison of the density of the straight-line mixture with that of the contributing waters (represented roughly by the $\sigma_t=27.6$ curve) suggests a density increase due to cabballing of about 0.02 sigma-t units.

The relatively large temperature and salinity variations above about 200 m in Jones Sound (Fig. 9a) are probably the effects of local summer warming and meltwater addition. Flow reversals, leading to the occasional presence in Jones Sound of Baffin Bay rather than Arctic Ocean surface water, are also possible. Palfrey and Day (1968) hypothesized an eastward flow of water, including that of the Baffin Bay cold core which is too saline to have originated in the Arctic Ocean, through Jones Sound. The 1966 data were obtained, however, from the eastermost portion of Jones Sound and would be expected to reflect the known cyclonic circulation (see section 3.1) of Baffin Bay water there.

Lancaster Sound contains both Arctic Ocean Water from M'Clure Strait and Baffin Bay Water which participates in a cyclonic circulation in eastern Lancaster Sound (section 3.1). The relatively cold (-1.4°C), saline ($33.7\text{ }^{\circ}/\text{oo}$) Baffin Bay cold core is the indicator there of Baffin Bay Water. A less saline ($33.3\text{ }^{\circ}/\text{oo}$), relatively poorly defined cold core, about 100 m shallower, is probably a remnant of the preceding winter's cold Arctic Ocean surface layer. This core may have been advected either from the west or southward from Smith Sound. The layer between about 100 and 250 m, 0.1 to 0.2°C warmer than either of the two cold cores, seems to be a remnant of the Baffin Bay summer warmed surface layer which has been advected along isopycnal surfaces from northeastern to northwestern Baffin Bay and into Lancaster Sound. The Baffin Bay (i.e. the deeper) cold core is about 0.2°C warmer than farther east, apparently due to some mixing with the relatively warm Arctic Ocean Water from the west. This attenuation is small because the greater proportion of Arctic Water is less saline, hence less dense, than Baffin Bay Water; the Arctic Ocean Water largely flows over the cold core and its overlying warm water layer,

resulting in the 100-150 m thick near-surface layer of cold ($<0^{\circ}$ C) water present there.

The 50-100 m deep warm core in Lancaster Sound, a relatively minor feature, is of interest because of its persistence and because it has been suggested (Bailey 1957) to play a part in the formation of the North Water. Water in this core is identical to water found in a warm surface lens east of Devon Island (Fig. 13). The surface lens coincides with a cyclonic gyre (see section 3.1). The computed downward heat flux over an entire summer (see Table VIII) is sufficient to warm a 50 m thick layer 5° C and may account for the lens. Presence of the warm water in Lancaster Sound is due to its advection along isopycnal surfaces beneath the surface layer of inflowing Arctic Ocean Water. The fact that it is more prominent in northern than in southern Lancaster Sound (see, e.g., Figure 5c) suggests that it loses its identity relatively rapidly through mixing with the colder eastward-flowing Arctic Ocean Water.

2.1.4. Summary of the Arctic Water

The Baffin Bay Arctic Water originates both from eastern Davis Strait and from the upper layer of the Arctic Ocean via Smith, Jones and Lancaster sounds. Modification of the water from Davis Strait by winter cooling and freezing in Baffin Bay leads to the formation of a prominent cold core which occurs at shallower depths in northeastern and north central than in northwestern Baffin Bay and exhibits minimum temperatures. That Arctic Water of Arctic Ocean origin occurs in Smith, Jones and Lancaster sounds and in northwestern Baffin Bay, where it overlies the denser Arctic Water from Davis Strait.

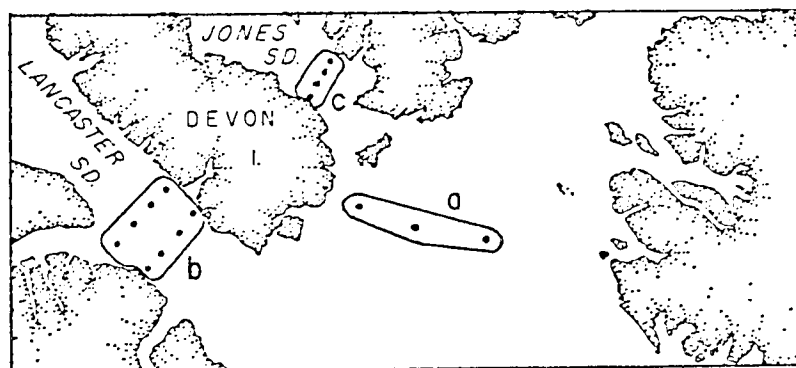
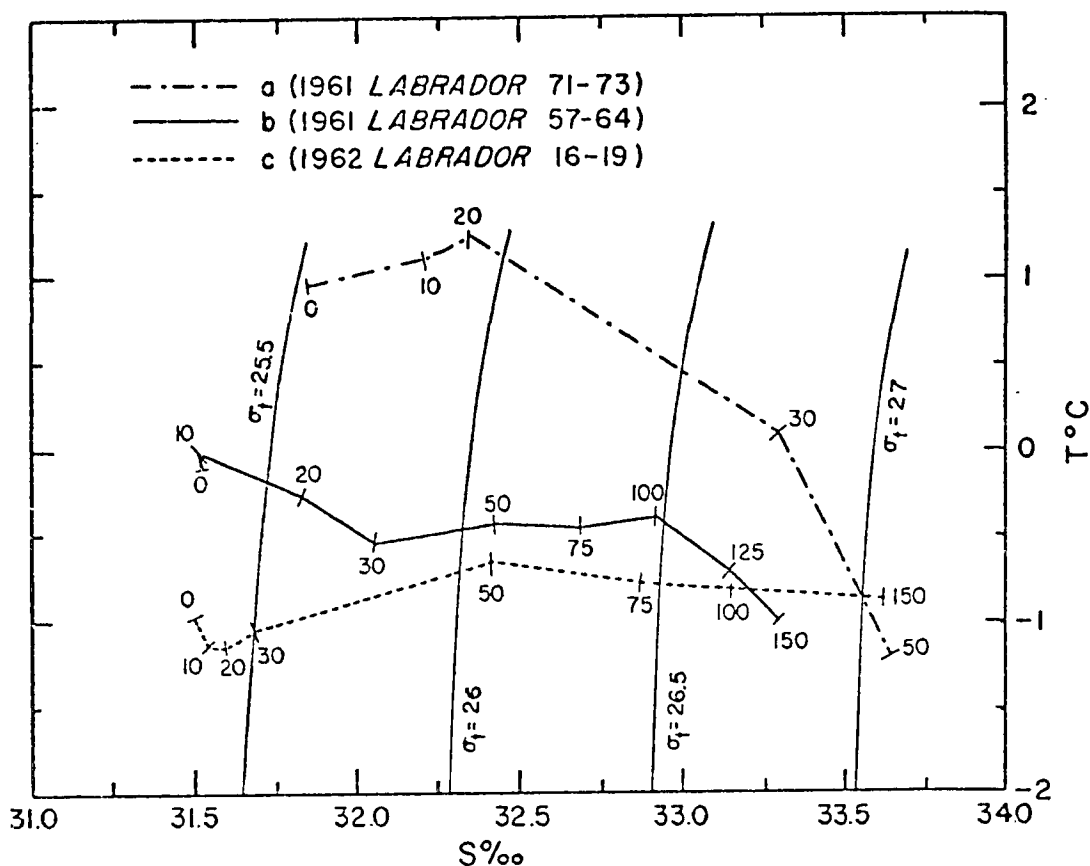


Figure 13. Temperature-salinity curves and location key illustrating the origin of the 50-100 m depth warm core observed in Lancaster Sound and less regularly in Jones Sound.

Two temperature minima occur in the Baffin Bay Arctic Water. The deeper and better defined is a consequence of isolation of the lower portion of the convective layer formed during the winter within Baffin Bay itself. A shallower minimum in Smith, Jones and Lancaster sounds and

in northwestern Baffin Bay is a remnant of the preceding winter's convective layer in the Arctic Ocean.

Net year-to-year temperature and salinity changes in the Baffin Bay Arctic Water layer appear to have been negligible from 1928 to 1966 except for the abnormally low temperatures and salinities observed in 1966; however, these may have been an artifact of sampling earlier in the season in 1966.

2.2 The Atlantic Water

2.2.1 Temperature and salinity distributions in the Atlantic Water from vertical cross-sections and profiles

The profiles (Fig. 4) indicate the nature of the vertical temperature and salinity structure in northern Baffin Bay's intermediate depth Atlantic Water layer, whose source is the Atlantic Ocean via the West Greenland Current. Profiles (f), (g), (h), (i), and (k) indicate the presence of the temperature maximum, while profiles (a) and (d) intersect the bottom before reaching the maximum. Only profile (i) shows the lower extent of the Atlantic Water layer; the other profiles are too shallow.

The Atlantic Water layer is characterized in general by temperatures above 0°C (Fig. 4a). It extends from 200-400 m down to 700-1400 m, and is generally deeper in northwestern than in northeastern Baffin Bay. Maximum temperatures occur in a prominent warm core at 450-600 m depth, and have been observed to vary from just above 0°C to higher than 2°C within northern Baffin Bay. The salinity corresponding to the maximum temperature is about 34.4 ‰ (Fig. 4b). Temperature and salinity both decrease upward from the warm core to the upper boundary of the layer.

Downward from the warm core, the temperature decreases while the salinity increases to a maximum value in northern Baffin Bay of about $34.5 \text{ }^{\circ}/\text{oo}$.

Horizontal temperature variations can be seen in the vertical profiles (Fig. 4a) but show more clearly in the section (Fig. 5). The warm core occurred at shallower depths (450 m) and was characterized by high maximum temperatures ($>1^{\circ} \text{ C}$) in northeastern and north central Baffin Bay (profiles (h)-(k), Fig. 4a). In the western portions and in Lancaster Sound (profiles (g) and (f), Fig. 4a) a maximum temperature of only about 0.5° C occurred at 500-600 m depth. Atlantic Water was present only in the eastern portion of the channel south of Smith Sound (section D, Fig. 5d) and exhibited variable maximum temperatures relative to those farther south in north central Baffin Bay.

Neither Smith Sound (profile (b), Fig. 4a; section A, Fig. 5a) nor Jones Sound (profile (e), Fig. 4a; section B, Fig. 5b) contained Atlantic Water.

Horizontal salinity variations can be seen in the cross-sections (Fig. 6) but are too small to discern clearly on the profiles (Fig. 4b). Salinities were generally about $0.1 \text{ }^{\circ}/\text{oo}$ lower between 400 and 800 m in northwestern than in north Baffin Bay (section E, Fig. 6e). Below 800 m the horizontal and vertical variations are within $0.05 \text{ }^{\circ}/\text{oo}$. Farther north, Atlantic Water at about 400 m in the eastern portion of the channel was accompanied by salinities $0.2 \text{ }^{\circ}/\text{oo}$ higher than in the western portion at the same depth (section D, Fig. 6d).

Only relatively small scale salinity variations occurred in Lancaster Sound (section C, Fig. 6c), where salinities were similar to those in northwestern Baffin Bay.

2.2.2. Horizontal and secular temperature and salinity variations in the Atlantic Water

Oceanographic data acquired during 1960-to-1964 and 1966 allowed the determination, for each of those years, of horizontal variations in the maximum warm core temperature T_{\max} in northern Baffin Bay. Year-to-year variations were sufficiently large to preclude combination of data from

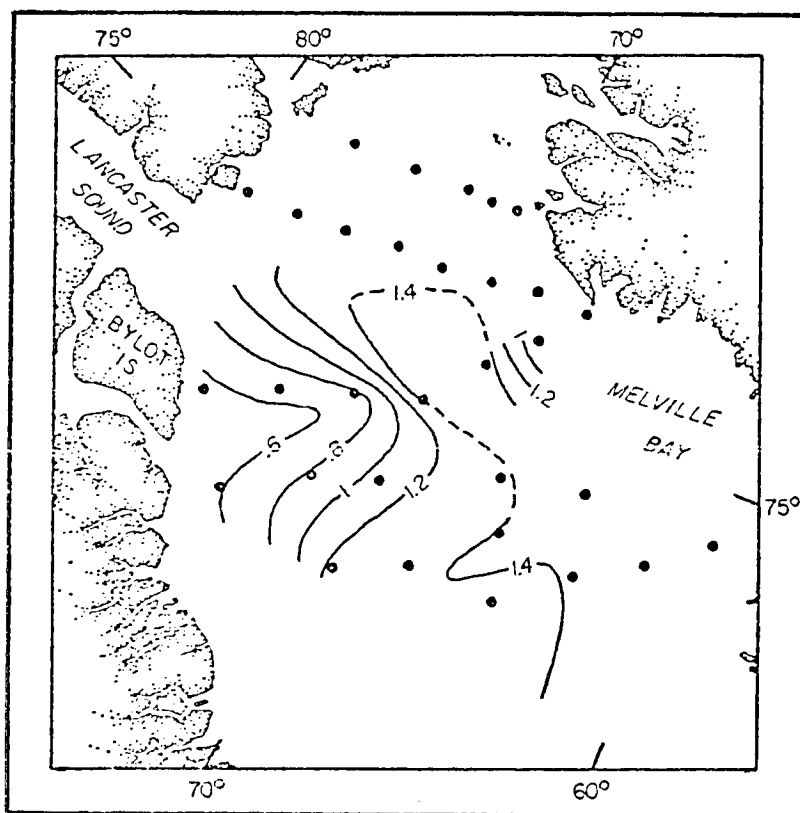


Figure 14. Horizontal distribution of T_{\max} , based on the 1964 LABRADOR data.

different years, although the yearly distribution patterns agreed with one another in their essential features. The 1964 data allowed the most complete determination of the horizontal distribution of T_{\max} (Fig. 14). A relatively warm ($>1.4^{\circ}\text{C}$) tongue extended westward from Melville Bay toward Lancaster Sound at about 75°N , a colder ($<1.2^{\circ}\text{C}$) tongue extended eastward from Bylot Island towards Melville Bay and another warm ($>1.4^{\circ}\text{C}$)

tongue extended southwestward in north central Baffin Bay.

Dashed lines represent the isotherms (Fig. 14) in northernmost Baffin Bay and indicate relatively wide variation in the temperature distribution there. Data acquired from that area in 1928, 1961-to-1964, 1966 and 1968-to-1969 were sufficient to roughly document the northward extent of Atlantic Water during each of these years (Fig. 15). No Atlantic Water was observed north of Thule in 1928, 1961 or 1962. In 1963 it was observed as far north as the mouth of Hvalsund in the eastern portion of the channel, and was registered at one station off Jones Sound. In 1964 it was observed as far north as Thule only in the extreme eastern portion of the channel. In 1966 it appeared north of Hvalsund and occupied the eastern half of the channel. A similar distribution, with slightly greater northward extent, was observed in 1968. The data were insufficient to establish the Atlantic Water's northward extent in 1969, but its zonal extent west of Thule suggested a northward extent similar to that observed in 1968.

The prominent feature of the Atlantic Water distribution in the southern Smith Sound region is its occurrence primarily along the eastern portion of the channel, with a lesser tongue observed occasionally off Jones Sound. There is a tendency for the Atlantic Water to occur farther northward in more recent years, the maximum northward extent having been observed in 1968 and, possibly, 1969. It is noteworthy that 1966, 1968 and 1969 were abnormally light ice years in the eastern Arctic, as observed during summer 1966 by Palfrey and Day (1968) and during the summers of 1968 and 1969 from CGC WESTWIND. There are, however, insufficient data to deduce any relations between northward extent of the Atlantic Water and severity of the ice season.

In order to combine all data, with the exception noted below, a statistical analysis similar to that applied above (section 2.1.2) to

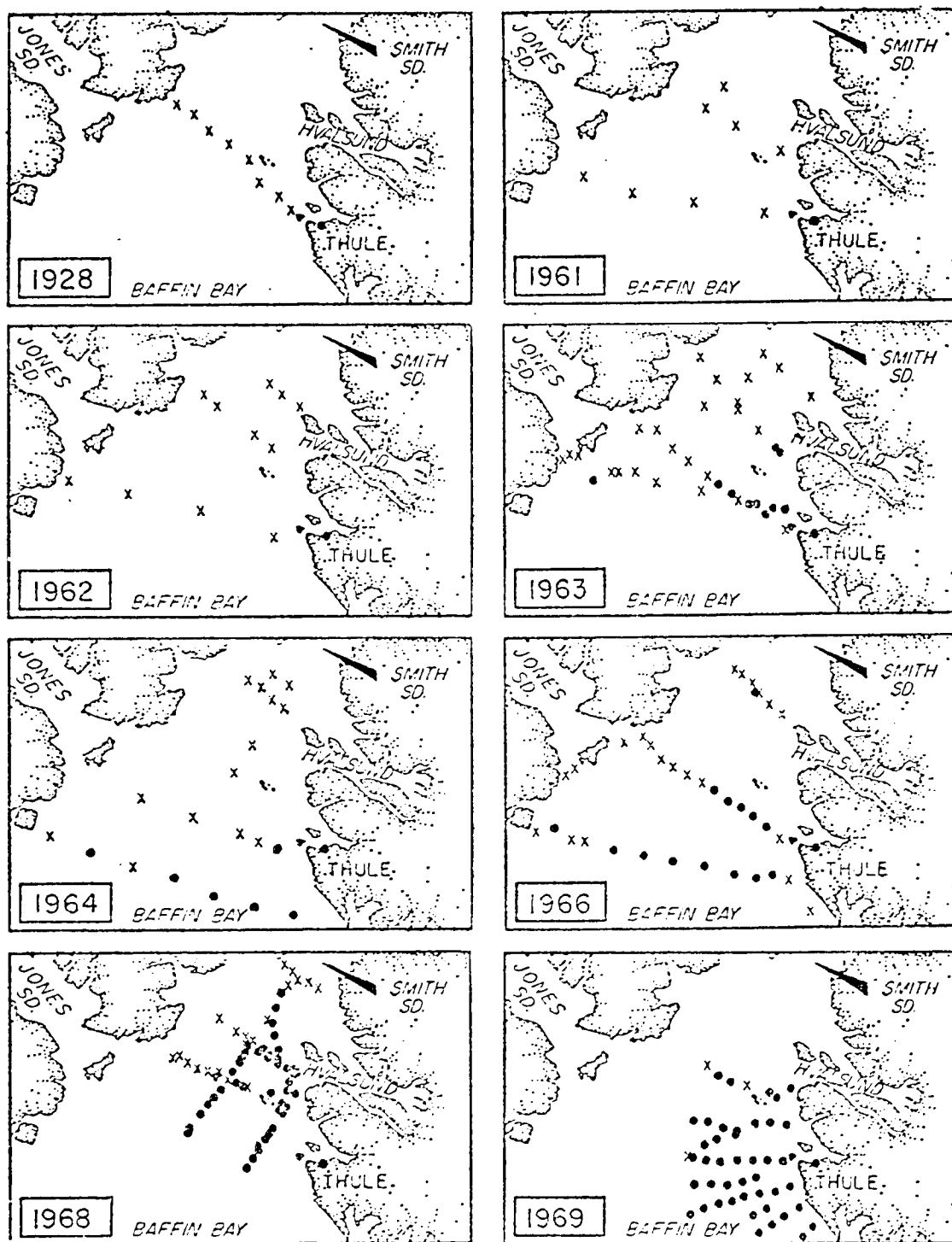


Figure 15. Extent of occurrence of Atlantic Water in the Smith Sound region for various years. Crosses indicate absence of Atlantic Water, while dots indicate that Atlantic Water was present.

the Arctic Water was carried out. Mean summer values of maximum temperature (\bar{T}_{\max}), salinity at the warm core (\bar{S}_{\max}) and Atlantic Water thickness (\bar{h}) were computed for each of 13 sub-areas (Fig. 3). The results are presented in Table IV and Figure 16.

Data acquired during 1969 (see, e.g., profile (j), Fig. 4a) suggest that the maximum temperature T_{\max} was anomalously high in that year. The 1969 data were acquired only from northeastern Baffin Bay and southern Smith Sound, and were not used in the computations in order to avoid possible biasing of the means for northeastern relative to those for the rest of northern Baffin Bay.

The horizontal distribution of \bar{T}_{\max} agrees generally with the distribution of \bar{T}_{\max} in 1964 (Fig. 14). The highest values occurred in northeastern and north central Baffin Bay (areas 1-3, 10), while values are less westward (areas 4 and 5) and in Lancaster Sound (area 6). At no time has Atlantic Water been observed in Smith Sound (area 12) or Jones Sound (area 7). The value of \bar{T}_{\max} for area 8 was based on two observations barely west of the boundary with area 9, therefore area 8 contained essentially no Atlantic Water. T_{\max} was observed only once in Kane Basin (area 13); water warmer than 0° C was, however, not uncommon.

Variability, deduced from the ranges, was relatively small in the northeastern and central regions (areas 1-3 and 10); these were also the regions having maximum \bar{T}_{\max} . The variability was larger in the west (areas 4-5) but smaller in Lancaster Sound (area 6). A relatively large variability occurred in area 9, as expected considering the temperature irregularities in section D (Fig. 5d) and profile (d) (Fig. 4a).

\bar{S}_{\max} was maximum and uniform (to within the accuracy of its determination) throughout northeastern and north central Baffin Bay (areas 1-4 and

TABLE IV
Mean summer parameters, by area*, for the Atlantic Water

| AREA | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
|----------------------------------|-------|-------|-------|-------|-------|-------|----|-------|-------|-------|-------|----|-------|
| \bar{T}_{\max} ($^{\circ}$ C) | 1.30 | 1.33 | 1.30 | 1.05 | 0.75 | 0.65 | | 0.25 | 0.69 | 1.28 | 0.80 | | 0.07 |
| Range | 0.09 | 0.28 | 0.36 | 0.50 | 0.49 | 0.36 | | 0.30 | 0.88 | 0.29 | 0.09 | | |
| Total no. obs. with max. | 7 | 10 | 21 | 28 | 27 | 39 | 0 | 2 | 48 | 10 | 2 | 0 | 1 |
| No. obs. without max. | 11 | 12 | 4 | 2 | 6 | 27 | 0 | 3 | 30 | 17 | 2 | 0 | 14 |
| No. obs. without T_{00} C | 0 | 3 | 0 | 0 | 5 | 13 | 34 | 64 | 43 | 7 | 0 | 89 | 24 |
| \bar{S}_{\max} (‰/‰) | 34.43 | 34.45 | 34.46 | 34.44 | 34.40 | 34.37 | | 34.08 | 34.25 | 34.43 | 34.40 | | 34.73 |
| Range (‰/‰) | 0.05 | 0.04 | 0.05 | 0.07 | 0.12 | 0.16 | | | 0.38 | 0.07 | 0.06 | | |
| Total no. obs. | 5 | 19 | 26 | 25 | 26 | 40 | 0 | 3 | 84 | 23 | 4 | 0 | 1 |
| \bar{h} (m) | 410 | 580 | 560 | 390 | 240 | 190 | | | 110 | 400 | | | |
| Range (m) | 70 | 120 | 50 | 260 | 160 | 210 | | | 170 | | | | |
| Total no. obs. | 2 | 8 | 25 | 21 | 19 | 17 | 0 | 0 | 23 | 1 | 0 | 0 | 0 |

*See Figure 3 for locations of numbered areas.

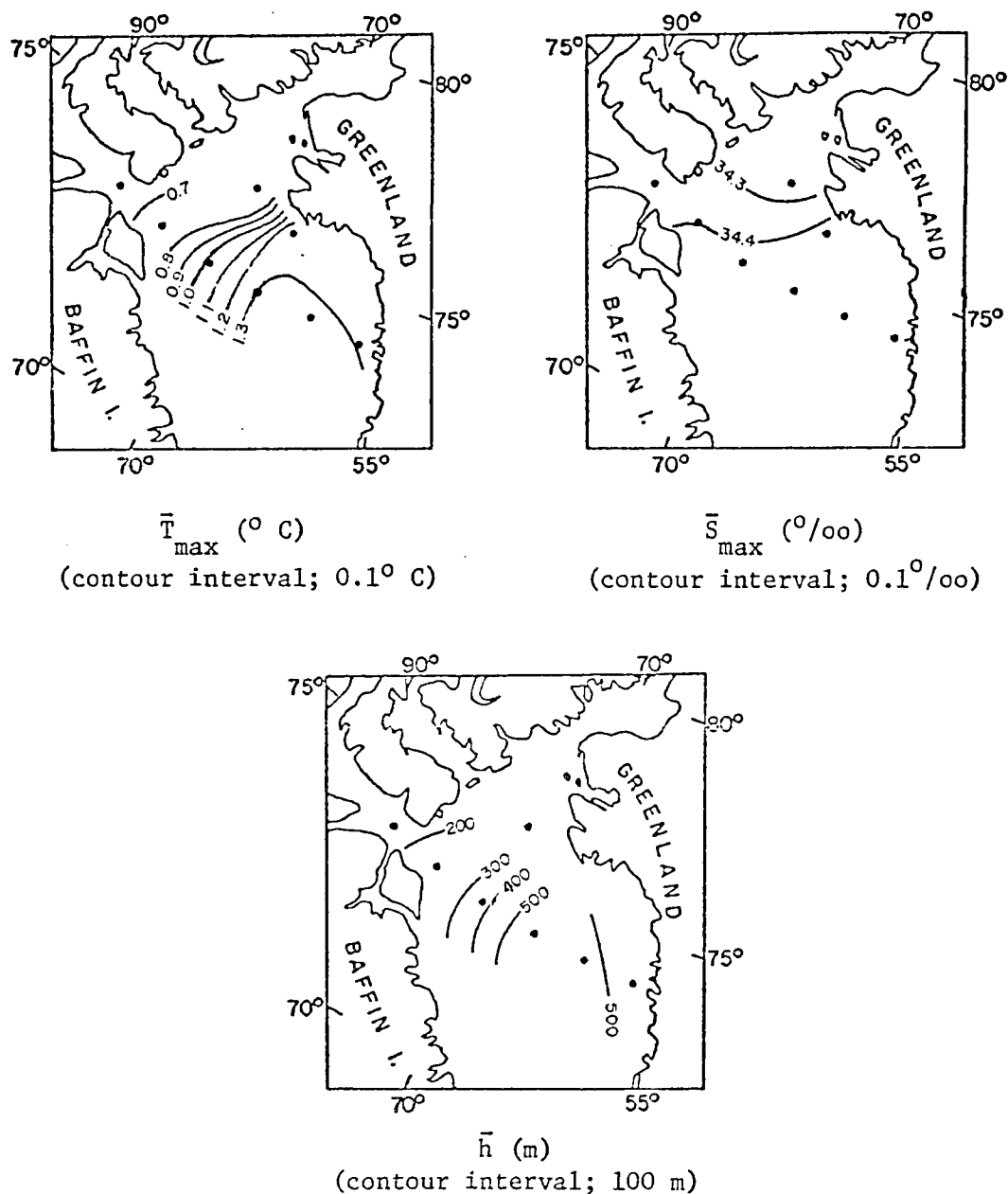


Figure 16. Graphical presentation of \bar{T}_{\max} , \bar{S}_{\max} and \bar{h} from Table IV.

10-11). There seemed to be a tendency for \bar{S}_{\max} to be less in the western portion (area 5), and it is definitely smaller to the north (area 9).

\bar{S}_{\max} was relatively low in Lancaster Sound (area 6). The high value in Kane Basin (area 13) agrees with that noted on profile (a) (Fig. 4b).

Salinity variability was relatively small throughout the northeastern and north central portions (areas 1-4 and 10-11), but was greater towards and into Lancaster Sound (area 6). The smallest variabilities generally coincided with the largest \bar{S}_{\max} . The maximum observed salinity variability occurred in southern Smith Sound (area 9), as did the maximum temperature variability, and coincided with a relatively low \bar{S}_{\max} .

To determine the relative horizontal distribution of Atlantic Water, the layer was redefined to include water bounded by the 0.5°C rather than the 0°C isotherm. The bounding isotherm does not affect the relative layer thickness in different areas. This choice was made, as for the Arctic Water (see section 2.1.2), in order to optimize the proportion of data from which thickness could be determined. Using this criteria, the Atlantic Water was thickest in northeastern and north central Baffin Bay (areas 2-3). Thickness decreased westward into Lancaster Sound (areas 4-6) and northward toward Smith Sound (area 9). Extreme eastern and northeastern Baffin Bay (areas 1 and 10) was characterized by a layer quite thin relative to that in the north central portion. A sparsity of data from extreme eastern Baffin Bay suggests that the thickness value there is subject to greater error than in other areas. Large variations in the layer thickness occurred, particularly in Lancaster Sound (area 6) and south of Smith Sound (area 9).

It seemed desirable to detect any secular changes in parameters of the warm core. Data of sufficiently wide geographic coverage to yield means unbiased by horizontal variations were obtained only during 1928, 1961 to 1964 and 1966. Values from areas 1-5 and 10 were analyzed using the "sign test" discussed in section 2.1.2. The results are presented in Table V. Significant changes were a warming sometime between 1928 and

TABLE V

Late summer variations in temperature, salinity and thickness
of the warm core for various years

T_{\max}

During period in column A, core was (colder than) (warmer than) (of the same temperature as) during period in row B.

| A \ B | 1961 | 1962 | 1963 | 1964 | 1966 |
|-------|-------|------|-------|-------|------|
| 1928 | C-2.5 | C-10 | C-2.5 | C-11 | C-10 |
| 1961 | | W-31 | C-2.5 | S | C-31 |
| 1962 | | | C-31 | W-31 | S |
| 1963 | | | | W-2.5 | S |
| 1964 | | | | | C-31 |

Key: C-colder than. W-warmer than. S-of the same temperature as.
Numbers following C, W or S indicate per cent significance level of temperature change as determined from the sign test.

S_{\max}

During period in column A, core was (more saline than) (less saline than) (of the same salinity as) during the period in row B.

| A \ B | 1961 | 1962 | 1963 | 1964 | 1966 |
|-------|------|------|------|-------|------|
| 1928 | L-31 | L-10 | L-11 | S | S |
| 1961 | | S | L-31 | H-13 | S |
| 1962 | | | L-31 | H-31 | S |
| 1963 | | | | H-2.5 | H-31 |
| 1964 | | | | | S |

Key: L-lower salinity than. H-higher salinity than. S-of the same salinity as. Numbers following L, H or S indicate per cent significance level of salinity change as determined from the sign test.

h

During period in column A, core was (thicker than) (thinner than) (of the same thickness as) during period in row B.

| A \ B | 1961 | 1962 | 1963 | 1964 | 1966 |
|-------|------|------|------|------|------|
| 1928 | N-31 | S | N-50 | N-13 | N-13 |
| 1961 | | T-31 | N-50 | N-50 | N-13 |
| 1962 | | | N-13 | N-13 | N-13 |
| 1963 | | | | N-31 | N-13 |
| 1964 | | | | | N-13 |

Key: T-thicker than. N-thinner than. S-of the same thickness as.
Numbers following T, N or S indicate per cent significance level of thickness change as determined from the sign test.

1961 and an abnormally thick layer in 1966. The latter feature coincided with the abnormally low temperature and salinity of the Arctic Water in 1966 (Table II).

2.2.3. The origins of the Atlantic Water

Water of the correct temperature ($>0^{\circ}$ C) and salinity ($<34.5^{\circ}/\text{oo}$) to contribute to the Baffin Bay Atlantic Water does not occur in the Arctic Ocean. It is present, rather, in eastern Davis Strait from whence it must be advected northward by the West Greenland Current (Fig. 10). Its low temperature in northeastern Baffin Bay (<1 to 2° C) relative to that in Davis Strait (about 5° C) appears to be a consequence of mixing with relatively cold ($<0^{\circ}$ C) Arctic Ocean Water approximately along the $\sigma_t=27.6$ surface. That Baffin Bay Atlantic Water originates from the south via Davis Strait has, in fact, long been recognized (Kiillerich 1939; Dunbar 1951; Bailey 1957; Palfrey and Day 1968).

The Atlantic Water undergoes modification within northern Baffin Bay (Fig. 12). Atlantic Water has never been observed in Smith or Jones sounds; the net flow through these channels into northern Baffin Bay apparently prevents Atlantic Water from entering them. The common presence of Atlantic Water in Lancaster Sound seems to be a consequence of the cyclonic circulation of Baffin Bay Water there; the warm core water in Lancaster Sound is approximately a 50-50 mixture between water from the northeastern Baffin Bay warm core and water from 180 m in either the Lincoln Sea or M'Clure Strait. It will be seen (section 2.4) that such mixing between the Baffin Bay warm core and Arctic Ocean Water occurs primarily in the southern Smith Sound region, and that only a minor amount of mixing between these waters occurs in Lancaster Sound. This is not surprising, considering

the relatively shallow sill, west of Lancaster Sound, which prevents eastward flow of Arctic Ocean Water below 180 m which could mix with Baffin Bay Atlantic Water; the 250 m sill in Kane Basin allows the southward flow of Arctic Ocean Water which can mix with Baffin Bay Atlantic Water.

A similarity between temperature-salinity characteristics in northwestern Baffin Bay and in Lancaster Sound suggests a common origin for water at both locations; apparently the water circulates freely between northwestern Baffin Bay and Lancaster Sound (see section 3.1). The 0.1°C higher temperature maximum in northwestern Baffin Bay than in Lancaster Sound might be due to: (a) admixture of Arctic Ocean Water into the warm core in Lancaster Sound, lowering the warm core temperature relative to that in northwestern Baffin Bay; (b) admixture of warm water from northeastern Baffin Bay, raising the maximum temperature in northwestern Baffin Bay; (c) diffusion of heat out from the warm core during circulation around northern Baffin Bay; or (d) some combination of the above.

In an attempt to estimate the effect of vertical diffusion on the warm core, an eddy conductivity coefficient of $10\text{ cal/cm-sec-}^{\circ}\text{C}$ (based on Table 65, Sverdrup et al 1942) was chosen. A mean speed at the warm core of 2 cm/sec was selected, on the basis of geostrophic calculations (section 3.1), which yielded a circulation time around northwestern Baffin Bay and Lancaster Sound of about $0.3 \times 10^7\text{ sec}$. Using the vertical temperature profiles for the area (Fig. 4a), a decrease in temperature by 0.1°C was computed which, within the possible errors in choice of eddy coefficient and speed, is of the same order as the observed difference between northwestern Baffin Bay and Lancaster Sound. It therefore appears that vertical diffusion is important in accounting for the observed maximum temperature variation.

Water occurring at the bottom of the Atlantic Water layer in Lancaster Sound (i.e. at about 800 m depth) is nearly identical to water at 1200-1300 m in Baffin Bay (-0.1° C and 34.4 ‰). This water was probably formed in southern Smith Sound as a mixture of Baffin Bay Atlantic Water and Arctic Ocean Water. Such a mixture may sink and contribute to Baffin Bay Deep Water (section 2.3).

The greater depth, by 100-150 m, of the warm core in Lancaster Sound and northwestern Baffin Bay than in northeastern Baffin Bay is a consequence primarily of the downward westward slope of the isohalines (hence also the isopycnals) due to presence of overlying relatively low salinity Arctic Ocean Water in the western portions; the warm core, advected primarily along isopycnal surfaces, is constrained by the sloping isopycnals to descend to greater depths in the western portions. (The observed cabballing effects can increase the density sufficiently in Lancaster Sound and northwestern Baffin Bay to account for only about 10% of the observed depth difference.)

2.2.4. Summary of the Atlantic Water

Baffin Bay Atlantic Water is characterized by temperatures above 0° C and maximum temperatures in the water column which produce a pronounced core. Atlantic Water originates in Davis Strait and is advected to northern Baffin Bay by the West Greenland Current. The Atlantic Water's temperature is lowered, during the course of this advection, primarily by mixing with Arctic Ocean Water which has flowed into Baffin Bay from the north. Further temperature attenuation occurs between northeastern and northwestern Baffin Bay due to admixture with Arctic Ocean Water which originates

primarily from Smith Sound. The Atlantic Water occurs at greater depths in northwestern than in northeastern Baffin Bay because it is advected along isopycnals which slope downward to the west. This slope is due to the near-surface presence in the west of relatively low salinity (therefore low density) Arctic Ocean Water from Smith, Jones and Lancaster sounds.

An overall warming of the Atlantic Water has occurred sometime between 1928 and 1966.

2.3 The Deep and Bottom Waters

Baffin Bay Deep Water occurs between 1200 m depth and the bottom. Temperatures within the Deep Water decrease from 0° C at 1200 m down to about -0.40° C at 1800 m. The temperature is nearly uniform below 1800 m; the slight observed vertical variation is of the proper magnitude to be an adiabatic effect (profile (i), Fig. 4a; section E, Figs. 5e, 6e). It was impossible to identify, due to the uniform temperature and salinity, any horizontal variation pattern in the Deep Water. Salinities are nearly uniform between 1300 m and the bottom, varying between 34.46 and $34.52^{\circ}/\text{oo}$.

The Deep Water below about 1800 m is noteworthy because its temperature and salinity are uniform nearly to within the accuracy of their measurement, i.e. to within $\pm 0.02^{\circ}$ C and $\pm 0.02^{\circ}/\text{oo}$. No measurable changes in temperature or salinity have occurred between 1928 and 1966. This extremely uniform, deep layer of water is the Baffin Bay Bottom Water.

Water at 250-300 m depth in Davis Strait has the correct salinity to contribute to the Deep Water, but too high a temperature (Fig. 10). A more probable origin for the Deep Water is suggested by a comparison of potential

temperature-salinity characteristics for the Arctic Ocean and Baffin Bay (Fig. 15). Single stations were selected to represent temperatures and salinities below 900 m in Baffin Bay, the Canadian and Eurasian basins of the Arctic Ocean and Smith Sound. The use of single stations can be justified by the small temperature and salinity variations at the depths under consideration. The Lincoln Sea curve was constructed by averaging the values of potential temperature and salinity at each depth for four stations.

Potential rather than in situ temperature was used because adiabatic effects become significant at the relatively great depths of the water mass under consideration.

Deep Water between the 0° C isotherm (at 1200 m) and 1500 m appears to be a mixture of Lincoln Sea water from about 225 m depth and water from the overlying Baffin Bay Atlantic Water layer. The Lincoln Sea water apparently retains its identity below 200 m as far south as Smith Sound, as noted above (section 2.1.3), while the region in southern Smith Sound is the site of mixing between this water and Baffin Bay Atlantic Water (see section 2.4). Water of the proper potential temperature and salinity to contribute to Baffin Bay Deep Water occurs in the Canadian Basin but is prevented from flowing into Baffin Bay by the 180 m limiting sill depth in Jones and Lancaster sounds.

Baffin Bay Deep Water below about 1500 m, including the Bottom Water, is of uncertain origin. Eurasian Basin Water at about 180 m depth is identical in potential temperature and salinity to Baffin Bay Bottom Water. Such water has been observed only once, however, in the Lincoln Sea-Nares Strait system; in Smith Sound in 1928 (GODTHAAB station 98). It was not detected in the 1967 Lincoln Sea data, apparently having lost its identity

by mixing with other water from both the Eurasian and Canadian Basins.

Previous writers (Bailey 1956; Collin 1965; Palfrey and Day 1968) concluded erroneously, due to their use of in situ rather than potential temperature, that water of the proper type to contribute to Baffin Bay Bottom

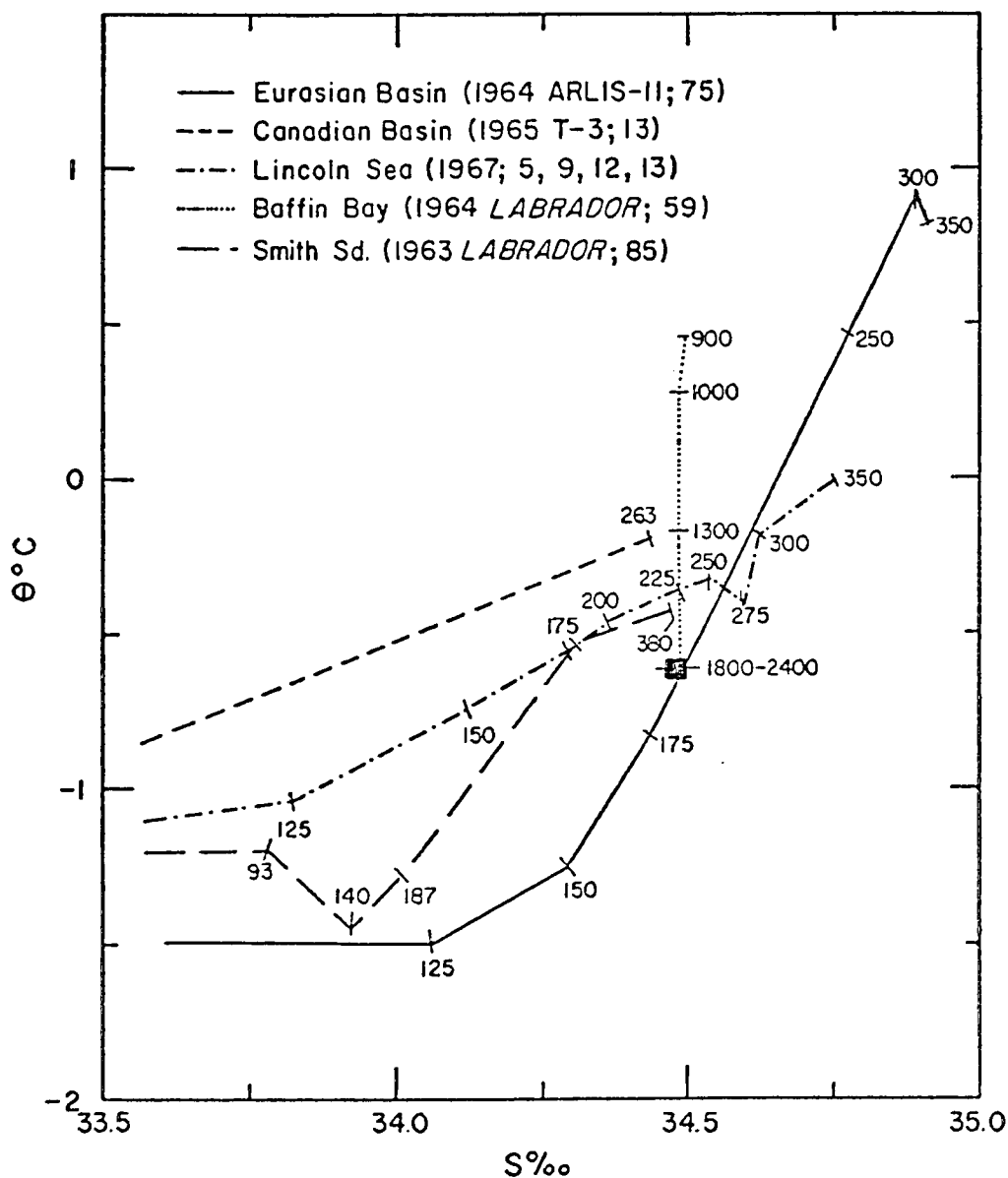


Figure 17. Potential temperature-salinity diagram showing relations between Baffin Bay Deep Water and Arctic Ocean Water. The Baffin Bay Bottom Water is represented by the black rectangle. Numbered ticks give depths in meters.

Water was commonly present in Nares Strait as far south as Kane Basin. They hypothesized that a relatively small southward pulse through Nares Strait would cause flow of this water into Baffin Bay. Potential temperatures reveal, however, that water of the proper salinity in Nares Strait is at least 0.2°C too warm to contribute to the Bottom Water. A southward pulse would have to be sufficient to advect water of the proper type into the Lincoln Sea and Nares Strait before it could reach Baffin Bay. Such pulses seem possible (see section 3.1), although it appears that replenishment of Bottom Water by this means must be less common than previously thought.

Sverdrup, Johnson and Fleming (1942) suggested an alternative mode of Deep and Bottom Water formation. They hypothesized that it was formed by a mixture of Labrador Sea deep water and Baffin Bay surface water whose salinity had been increased sufficiently by freezing to cause the water to sink. Redfield and Friedman (1970) suggested on the basis of deuterium-salinity ratios that Baffin Bay Deep and Bottom Waters contain excess brine which can only have been introduced by surface freezing. This information tends to uphold the mode of formation suggested by Sverdrup, Johnson, and Fleming.

It is impossible to demonstrate, on the basis of summer observations, that brine enrichment due to freezing is sufficient for creation of water dense enough to contribute to Bottom Water formation by the above means. Formation of 2 m ice leads to convection only down to the bottom of the cold core, i.e., down to 150 m (section 2.1.3); there is no basis for supposing that winter conditions would make deeper convection possible. Brine formation due to combined freezing and evaporation from open water

in the North Water area also appears insufficient for creation of dense enough water to contribute to the Bottom Water.

That the Bottom Water is replenished either intermittently or very slowly is suggested by the low dissolved oxygen concentrations observed there; these vary by less than 0.5 ml/l from 3.5 ml/l (about 40% saturation). They appear to have remained unchanged since 1928, although too few measurements of dissolved oxygen are available for a rigorous investigation of changes. The present suggestion of very slow and intermittent replenishment makes direct observation of such replenishment unlikely, particularly considering the relatively few data available from the Deep Water.

2.4 Some detailed studies in southern Smith Sound

It has been suggested (sections 2.2.3, 2.3) that the southern Smith Sound region is the primary site of mixing between south-flowing Arctic Ocean Water and Baffin Bay Atlantic Water. This region is, moreover, the site of the North Water. Oceanographic investigations in northern Baffin Bay during 1968 and 1969 were designed, under the auspices of the Baffin Bay-North Water Project, to investigate this area in detail (Muench, in press; Muench and Tidmarsh 1969). The data acquired allow the following more detailed discussion of water masses and mixing in the northern Baffin Bay-Smith Sound region.

It has been noted (section 2.2.2) that Atlantic Water was observed particularly far north in September 1968. Two separate surveys of the southern Smith Sound region were conducted at that time. The data from the first of these were sufficient to allow determination of the

temperature distribution on the $\sigma_t=27.5$ surface (Fig. 18). This so-called "isentropic diagram" (see Montgomery 1938) was constructed because it allows deduction of flow patterns in addition to indicating gross temperature variations. A mass of relatively warm ($>0^\circ\text{C}$) water occurred throughout the eastern portion of the channel and exhibited maximum temperatures ($>1^\circ\text{C}$) southeast of the Carey Islands. The boundary between the cold ($<0^\circ\text{C}$) water in the western part of the channel and the warmer water in the eastern part was characterized by horizontal wave-like undulations. Data from the second survey were insufficient to detect temperature variations on an isopycnal surface, but indicated the presence of time changes which precluded combination of data from the two surveys on a single plot.

A longitudinal temperature cross-section from Cape York to Smith Sound indicated considerable temperature variation throughout the water column (Fig. 19). A pronounced zone of relatively high horizontal temperature gradient between stations 28 and 31 marks the intersection of the cross-section with the boundary of the warm water mass noted on the $\sigma_t=27.5$ surface. This warm mass apparently filled the topographic depression east of the Carey Islands. It was present between about 200 m and the bottom from station 31 northward to the topographic rise at stations 39-40, and was characterized by irregular temperatures as typified by the cold blob of water 300 m deep at station 33. A layer of relatively warm ($>0^\circ\text{C}$) water about 300 m deep at stations 40-42 appeared to be a northward extension from the warm water mass. The overlying layer was colder; water colder than -1°C was uniformly present north of station 41 but was discontinuous south of that station. A 50 m thick surface layer of warm ($>0^\circ\text{C}$) water was present between stations 34 and 40.

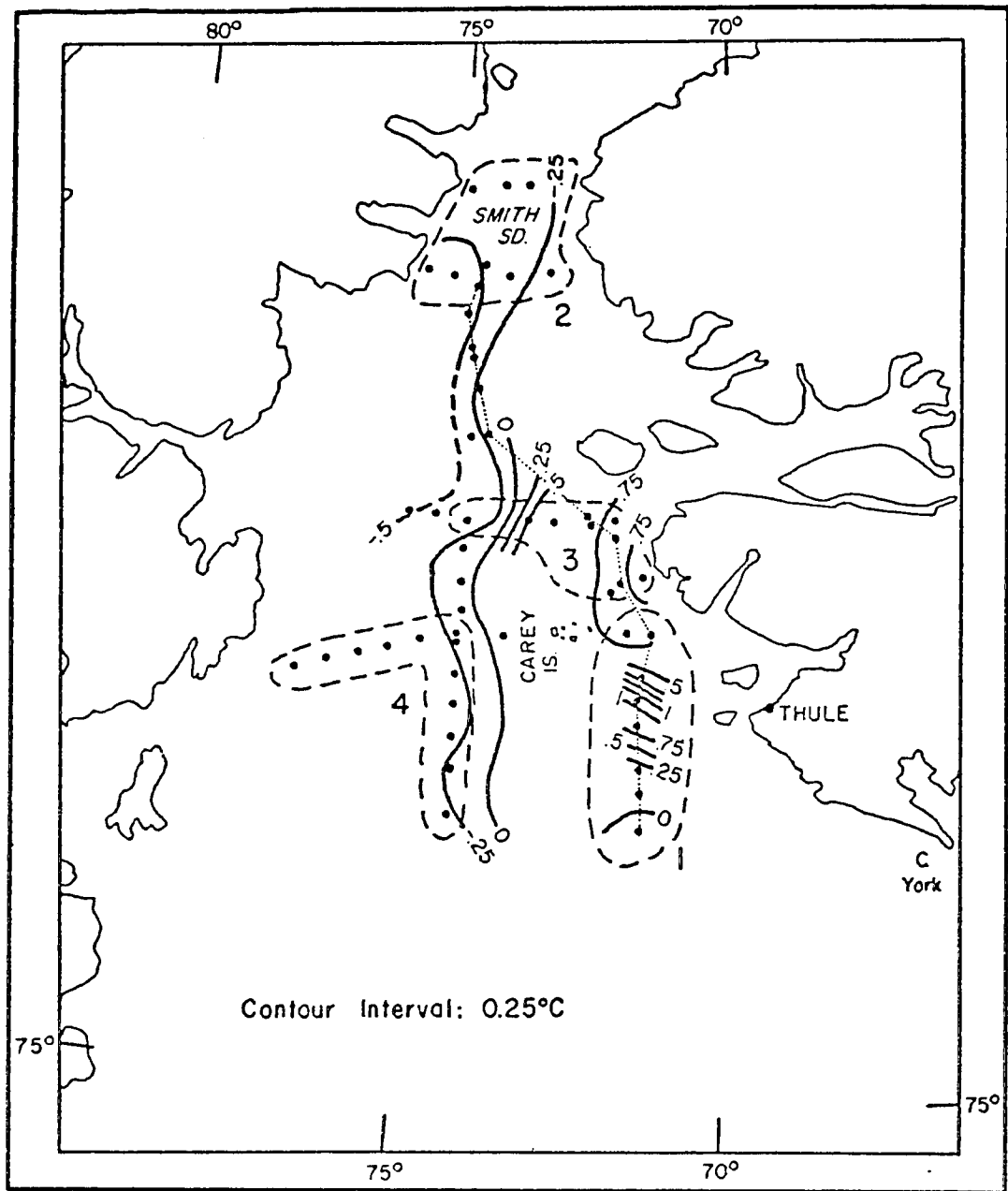


Figure 18. Temperature on the $\sigma_t = 27.5$ surface in the Smith Sound-Cape York region, 8-15 September 1968^t. Numbered areas delineated by dashed lines indicate locations of temperature-salinity curves in Figure 20. Dotted line indicates the location of the temperature cross-section in Figure 19.

The temperature-salinity curves (Fig. 20) represent the water masses in the areas delineated on Figure 18. The salinity values were, as

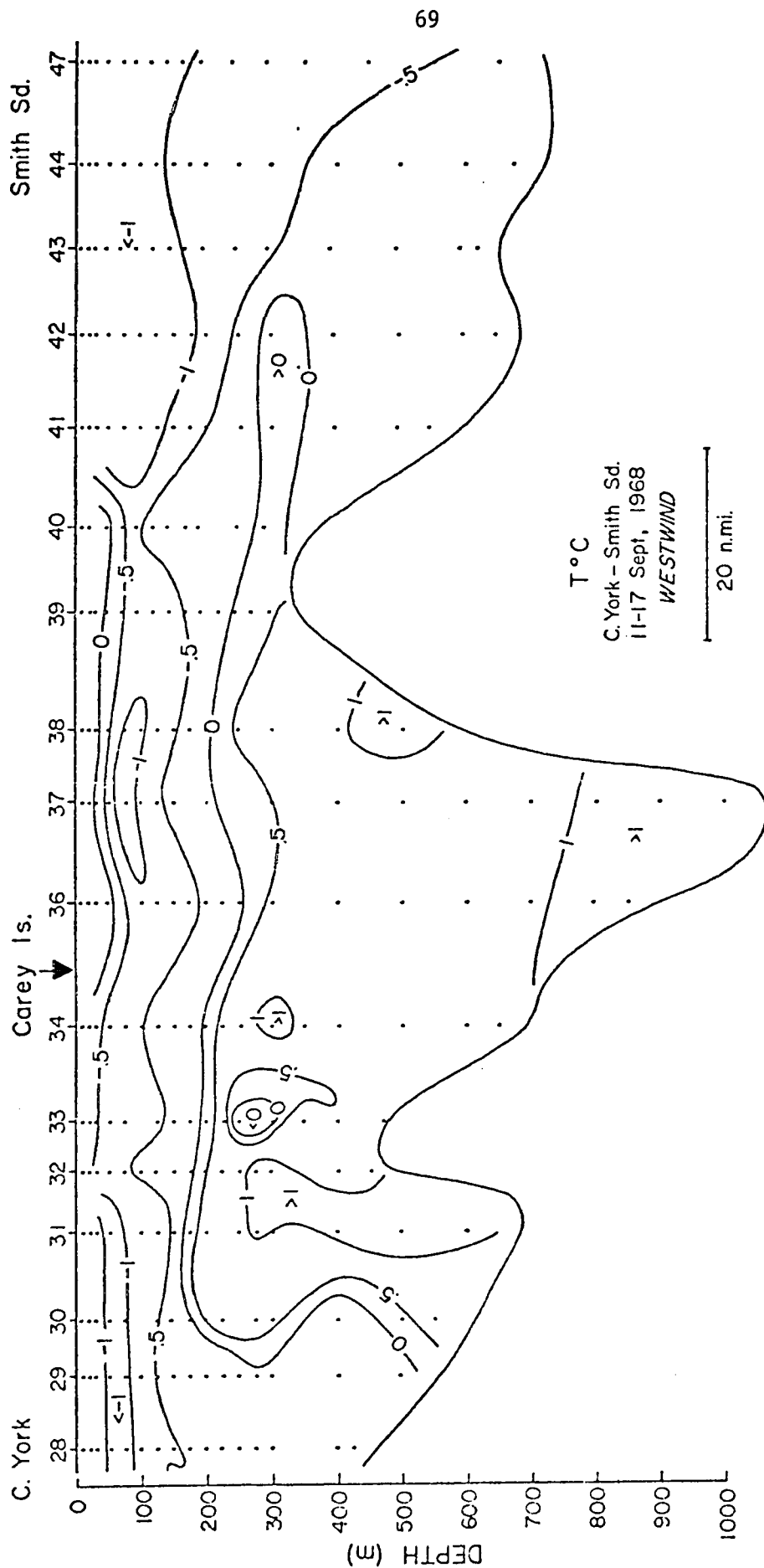


Figure 19. Longitudinal temperature distribution along the section indicated on Figure 18.

suggested by their wide scatter, unreliable and therefore temperature is regarded here as a more dependable water mass tracer.

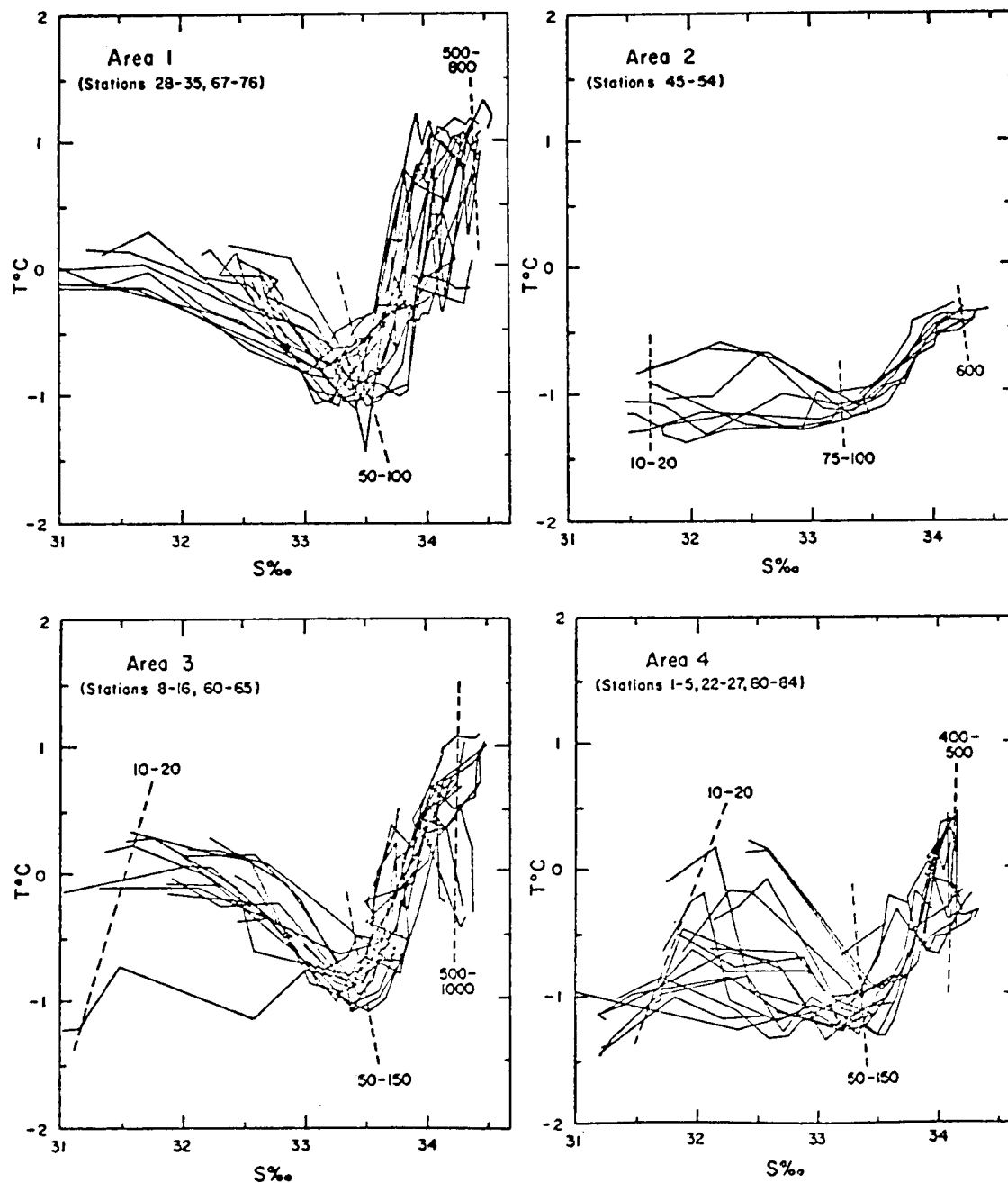


Figure 20. Temperature-salinity curves for the areas delineated on Figure 18, constructed using 1968 WESTWIND data.

Baffin Bay Atlantic Water occurred in area 1. The water in northern Smith Sound (area 2) is of Arctic Ocean origin, due to its relatively low

temperature (0°C to -1.5°C) and salinity ($<34.4^{\circ}/\text{oo}$). The remaining areas (3 and 4) contained mixtures of Baffin Bay and Arctic Ocean waters, corroborating the occurrence of mixing in this region. Maximum warm core temperatures (1.3°C) occurred southeast of the Carey Islands (area 1). Warm core temperatures were slightly lower (1.1°C) east and north of them (area 3) and had decreased still further west of them (area 4), down to 0.5°C . This decrease suggested a cyclonic circulation about the Carey Islands, during the course of which the warm core was attenuated by mixture with the colder Arctic Ocean Water. Attenuation of the warm core in the southern Smith Sound region due to mixing with Arctic Ocean Water, coupled with the overall cyclonic circulation pattern in northern Baffin Bay, is probably primarily responsible for the relatively low temperature (about 0.6°C) of the warm core in Lancaster Sound and northwestern Baffin Bay.

Water of the proper temperature ($<0^{\circ}\text{C}$) and salinity ($34.5^{\circ}/\text{oo}$) to contribute to the Deep Water occurred at a single station in Smith Sound (area 2). This water was too warm ($>-0.4^{\circ}\text{C}$) however, to contribute to the Bottom Water.

Oceanographic data obtained in 1969 from the Carey Islands-Melville Bay region allowed tracing to the south and southeast of the features observed in 1968. Two separate surveys were made of approximately the same area, allowing construction of two isentropic diagrams (Fig. 21). Temperature on the $\sigma_t=27.5$ surface was minimum ($<-0.25^{\circ}\text{C}$) west of the Carey Islands and delineated a cold ($<0^{\circ}\text{C}$) tongue west of Cape York which appeared, on the later date, to be detaching to form a cold blob. A relatively warm ($>0.5^{\circ}\text{C}$) tongue, similar to that observed in 1968, occurred east of the Carey Islands but the data were insufficient to delineate its

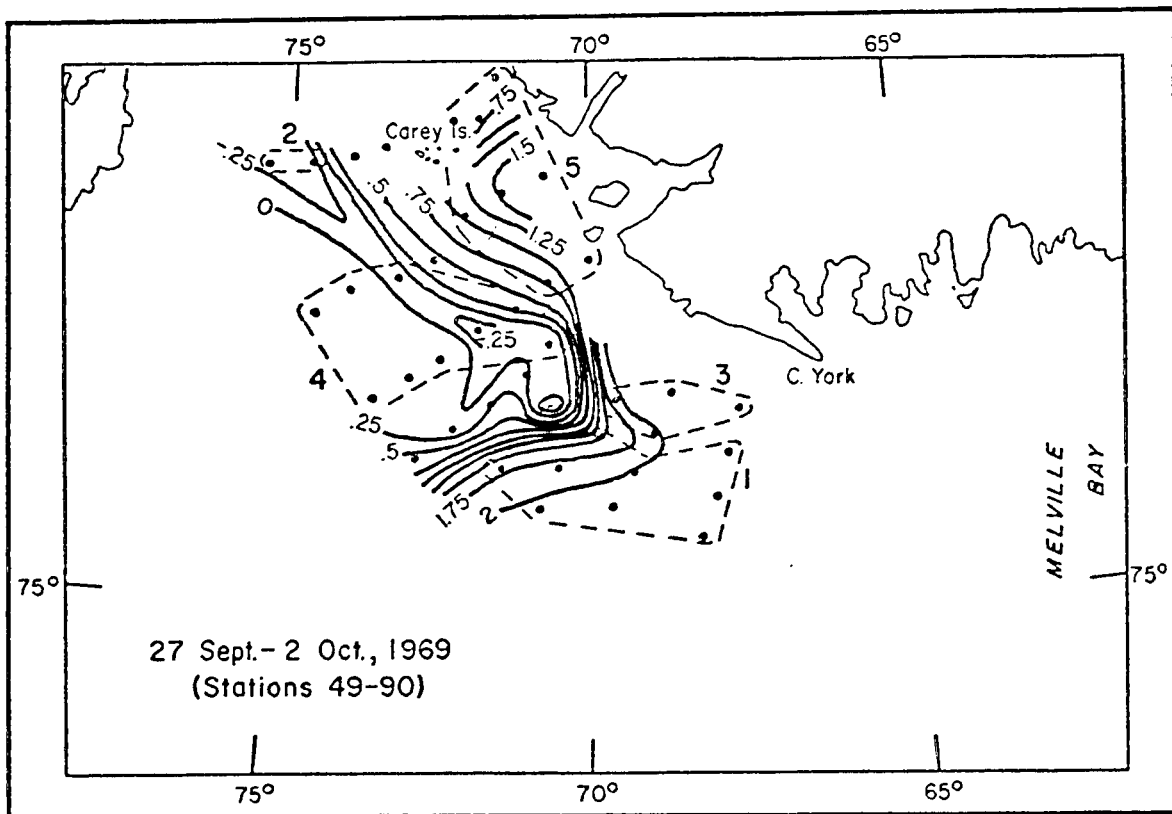
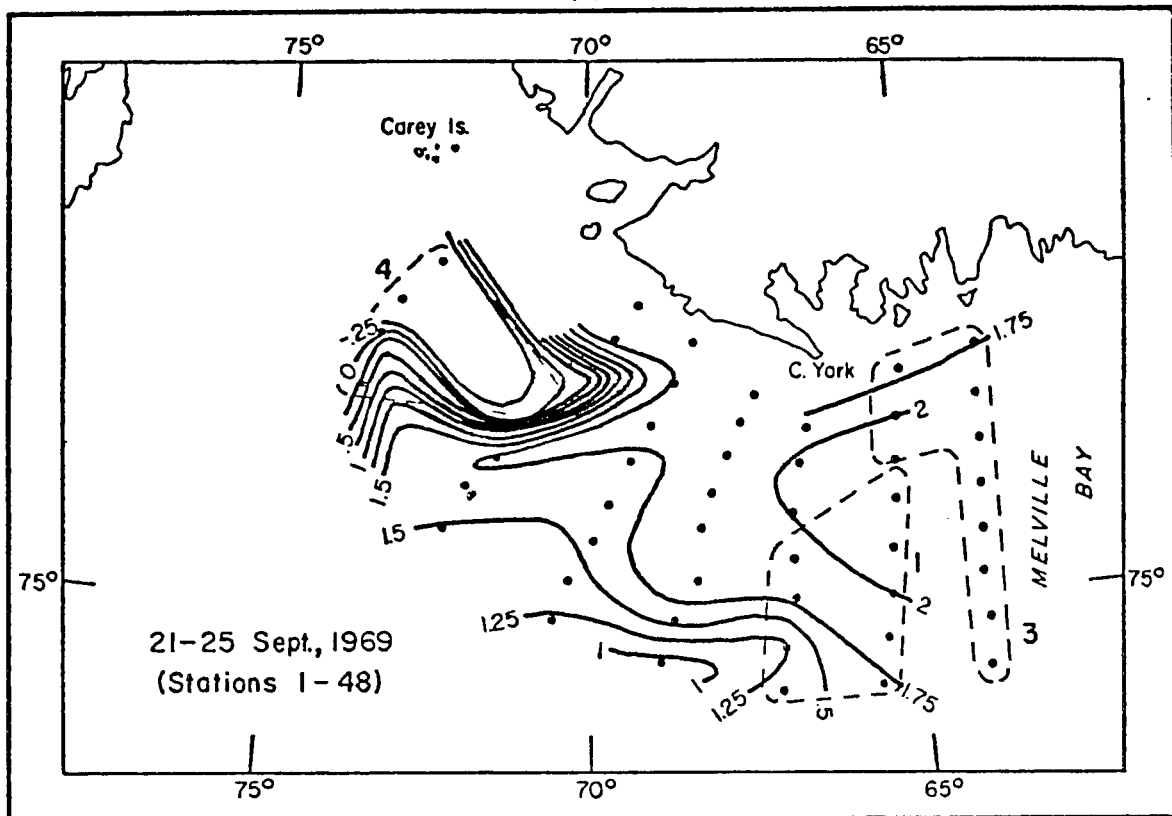


Figure 21. Temperature on the $\sigma_t = 27.5$ surface in the Smith Sound-Melville Bay region. Contour interval is 0.25°C , and data are 1969 WESTWIND. The numbered areas delineated by dashed lines correspond to the temperature-salinity curves in Figure 22.

northward extent. A still warmer ($>1.75^{\circ}\text{C}$) tongue extended westward from Melville Bay. The conjunction of the warm and cold tongues west of Cape York was associated with strong horizontal temperature gradients. As in 1968, time changes in the temperature distribution between the two surveys were sufficient to preclude combining the data on a single diagram.

The temperature-salinity curves (Fig. 22) represent waters in the areas indicated on Figure 21, and indicate the water in the southeastern portion (area 1) to be Baffin Bay Water. The maximum temperature ($>2^{\circ}\text{C}$) appeared to be, as noted (section 2.2.2), anomalously high during 1969. Water from the extreme northwest portion (area 2) appeared to have originated from Smith Sound; it was identical to Arctic Ocean Water except for some admixture of Baffin Bay Water resulting in elevated temperatures ($>0^{\circ}\text{C}$) below about 250 m.

Water in the northeastern portion (area 3) was Baffin Bay Water which had 0.5°C lower maximum and 0.7°C higher minimum temperatures than to the southwest (area 1). The currents in this region suggest (section 3.1) that the temperature difference may have been due to advection into area 3 of water which had slightly different temperatures than those in area 1. The difference would have been maintained by lack of communication between the waters below about 150 m due to the bank in Melville Bay. Insufficient data were available, however, to fully justify this explanation.

Mixtures of Arctic Ocean and Baffin Bay waters occurred east (area 5) and southwest (area 4) of the Carey Islands. Maximum temperatures were lower southwest of the Carey Islands than east of them, as in 1968. This further corroborated the cyclonic circulation, suggested on the basis of the 1968 temperature distribution, about these islands.

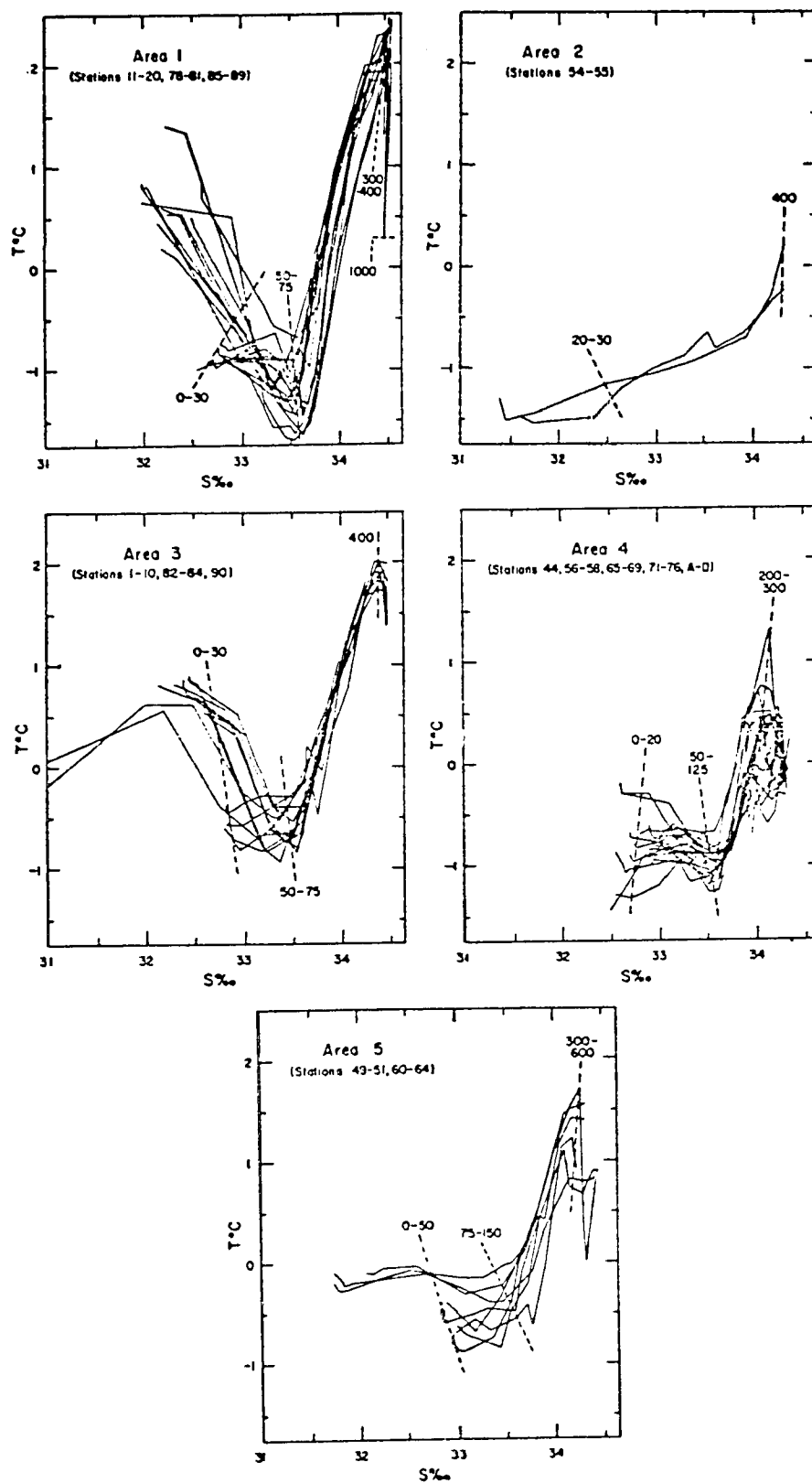


Figure 22. Temperature-salinity curves representing the areas indicated on Figure 21, constructed from 1969 WESTWIND data.

Water southwest of the Carey Islands was of proper temperature ($<0^{\circ}\text{C}$) but of too low a salinity ($<34.5^{\circ}/\text{oo}$) to contribute to Baffin Bay Deep Water. The 1969 salinity data were accurate, however, only to $\pm 0.05^{\circ}/\text{oo}$, so some of the water may have been more or less saline than indicated. It is also possible that the proper type water to contribute to the Deep Water occurred farther west, where no data were acquired.

The 1968 and 1969 data suggest that little water of proper type to contribute to the Baffin Bay Deep and Bottom Waters occurs as far south as the Carey Islands during the late summer. Most of the water flowing south from Smith Sound apparently mixes with Baffin Bay Water to form a mixture of too low a salinity to contribute to the Deep Water. The proper type water was found only at one station in Smith Sound. The small volume of such water suggests a relatively slow process of late summer replenishment for the Deep Water, in agreement with conclusions reached in section 2.3. No water of proper type to contribute to the Bottom Water was observed.

Data from any other single year were insufficient to delineate horizontal temperature or salinity distributions on the relatively small scale over which variations occur in the southern Smith Sound region. Year-to-year changes preclude combining the data. At no time, however, did data from other years contradict the distributions observed in 1968 and 1969. Maximum water temperatures have always, in the late summer, been higher east than west of the Carey Islands and the water in Smith Sound has always been Arctic Ocean Water. Apparently, general oceanographic conditions in the southern Smith Sound region have remained at least qualitatively constant, although the occurrence of the Atlantic Water farther northward in more recent years suggests that relatively large variations in quantities may be involved.

III. THE CIRCULATION

3.1 Circulation patterns in northern Baffin Bay

The large-scale surface circulation pattern in northern Baffin Bay has been known, primarily as a result of ship and ice drift observations, since the late 1800's. Information concerning these drifts has been compiled by Nutt (1966) and is summarized in Figure 23. The drifts of the INVESTIGATOR and the ENTERPRISE, the ADVANCE and the RESCUE and the RESOLUTE each indicated a net easterly surface flow through Lancaster Sound. The drift of the POLARIS party and of ice island WH-5 indicated a net southerly surface flow through Smith Sound, although the drift of WH-5 also suggested flow reversals there. The drifts of the ADVANCE and the RESCUE, the FOX and the POLARIS party indicated, in addition, a net southward surface flow within Baffin Bay during the winter months. The FOX drift indicated that a southward surface flow was occurring farther east than commonly observed during the summer. The more recent drift of WH-5 indicated a southward surface flow, during the winter, along the southeastern Baffin Island coast.

The first oceanographic data sufficient to allow deduction of the baroclinic circulation on the basis of dynamic computations were acquired by the GCDTHAAB in 1928. In 1940 the NORTHLAND completed two sections across the Baffin Current, where the GODTHAAB had been unable to venture due to heavy pack ice. The dynamic topography of the surface with respect to the 1500 db level was constructed, using the 1928 and 1940 data, by Barnes (1941) and represents all knowledge of the baroclinic circulation in Baffin Bay prior to the 1960's (Fig. 24).

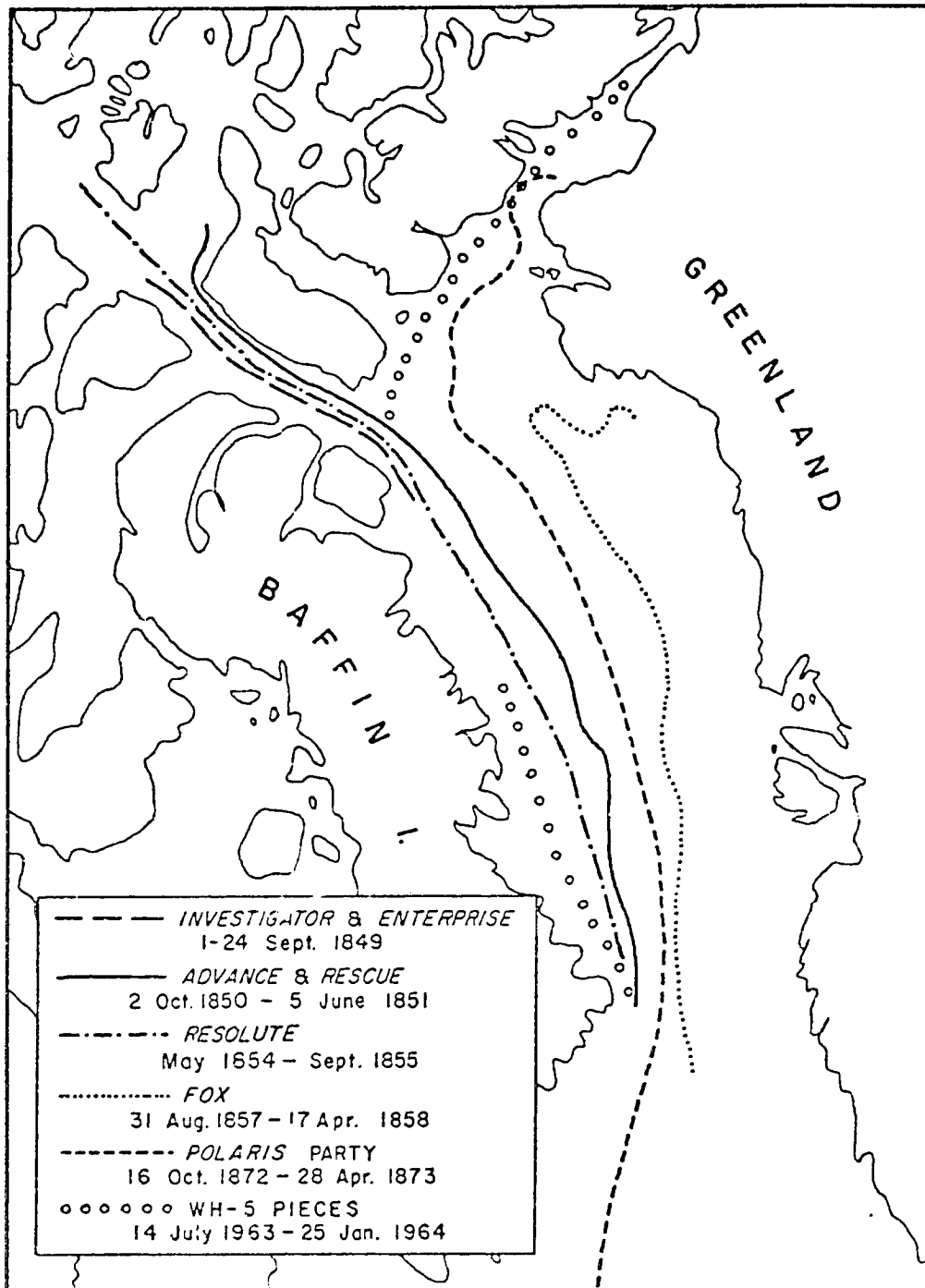


Figure 23. Recorded ship and ice drift tracks in the northern Baffin Bay region (after Nutt 1966).

Dynamic topographies of the surface relative to the 500 and 1000 db levels

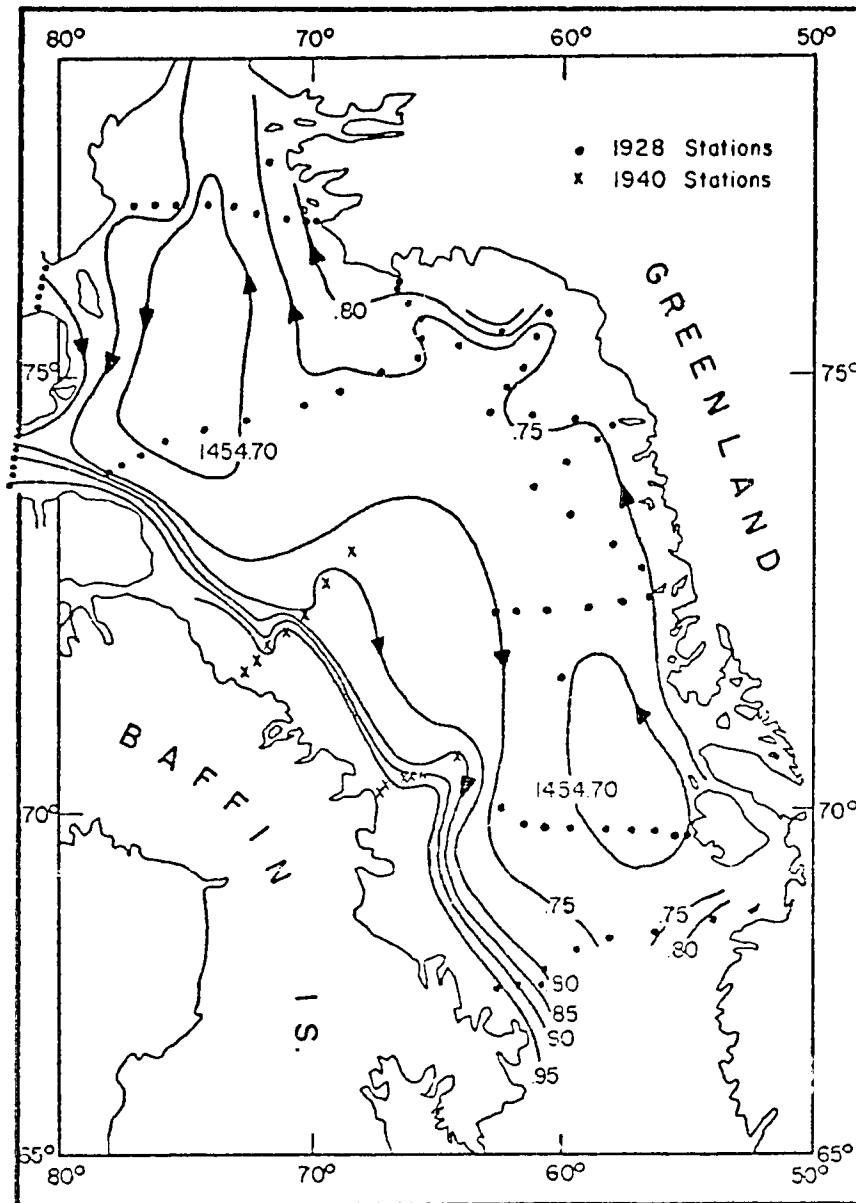


Figure 24. Dynamic topography of the surface with respect to 1500 db (from Barnes 1941). Contour interval is 0.05 dyn. m.

were constructed using the 1928 data by Kiilerich (1939); their similarity to the surface topography relative to 1500 db suggested that the baroclinic circulation in Baffin Bay occurs primarily above the 500 m level. Occurrence of the major portion of the baroclinic currents above 500 m is not surprising; the major salinity, hence density, variations occur above that depth.

The essential features of the baroclinic circulation in northern Baffin Bay are (Fig. 24) the northward West Greenland Current, which becomes westward in Melville Bay and south of Cape York; the stronger southward Baffin Current east of Bylot Island; and the cyclonic gyre in the region between Jones and Lancaster sounds and Cape York. The southern boundary of the cyclonic gyre is marked approximately by the Baffin Current.

Depths in the northern Baffin Bay region are generally less than 1500 m. (Fig. 2), therefore the circulation depicted in Figure 24 required extensive extrapolation of the reference level into regions shallower than 1500 m. This was done using the method of Helland-Hansen (1934), which assumes zero current velocity at the bottom and depends strongly upon the slopes of the isopycnals where they intersect the bottom. The depicted current speeds and directions may not be quantitatively correct because: (a) There was no reason for assuming that the baroclinic currents were zero near the bottom of northern Baffin Bay; (b) Error may have been introduced as a consequence of extrapolation of the reference level over large horizontal distances; and (c) This method gives no indication of the barotropic contribution.

Palfrey and Day (1968) chose, using the method of Defant (1961), a 700 db level for the 1966 data. The resulting surface dynamic topography agreed well with that of Barnes in its gross features and required less extrapolation of the reference level.

In further investigating the baroclinic circulation in northern Baffin Bay by means of the dynamic method, it was desired to minimize the necessity for extrapolation of reference levels. The method of Defant (1961) was used to determine suitable levels. Computations of geostrophic

currents using these reference levels revealed that computed velocities at 500 m were only about 2% of the surface velocities. A mean error of about $\pm 2\%$ would therefore be introduced into surface current calculations based on the 500 db level. Use of this level as a reference allowed, however, the construction of relatively complete dynamic topographies in northwestern Baffin Bay, Jones and Lancaster sounds without the need for extrapolation. This is felt to compensate for the known error in neglecting baroclinicity below 500 m.

Surface dynamic topographies were constructed relative to 500 db using data acquired during 1960-to-1963 and 1966 (Fig. 25). It is readily apparent that year-to-year and shorter term variations precluded combination of the data from different years, although certain features persisted. A relatively concentrated current east of Bylot Island (the Baffin Current) was always present. It flowed eastward from Lancaster Sound, and exhibited a tendency to continue eastward rather than turning south and paralleling the Baffin and Bylot island coasts. In 1961, 1963 and 1966 this eastward current appeared to coincide with the southern boundary of the cyclonic gyre east of Devon Island (Fig. 24). The 1960 and 1962 data were insufficient for delineation of the current at those times.

Topographic effects exert significant control on the course of the Baffin Current after its emergence from Lancaster Sound. A divergence occurs east of Bylot Island, with a northern branch flowing eastward to apparently contribute to the cyclonic circulation there and a southwesterly branch flowing southward parallel to the Bylot and Baffin island coasts. This divergence appears to be due to the tendency in high latitudes for currents to parallel the bottom contours, i.e., conservation of potential vorticity. (The bottom contours are indicated in Figure 25 by the 800 m

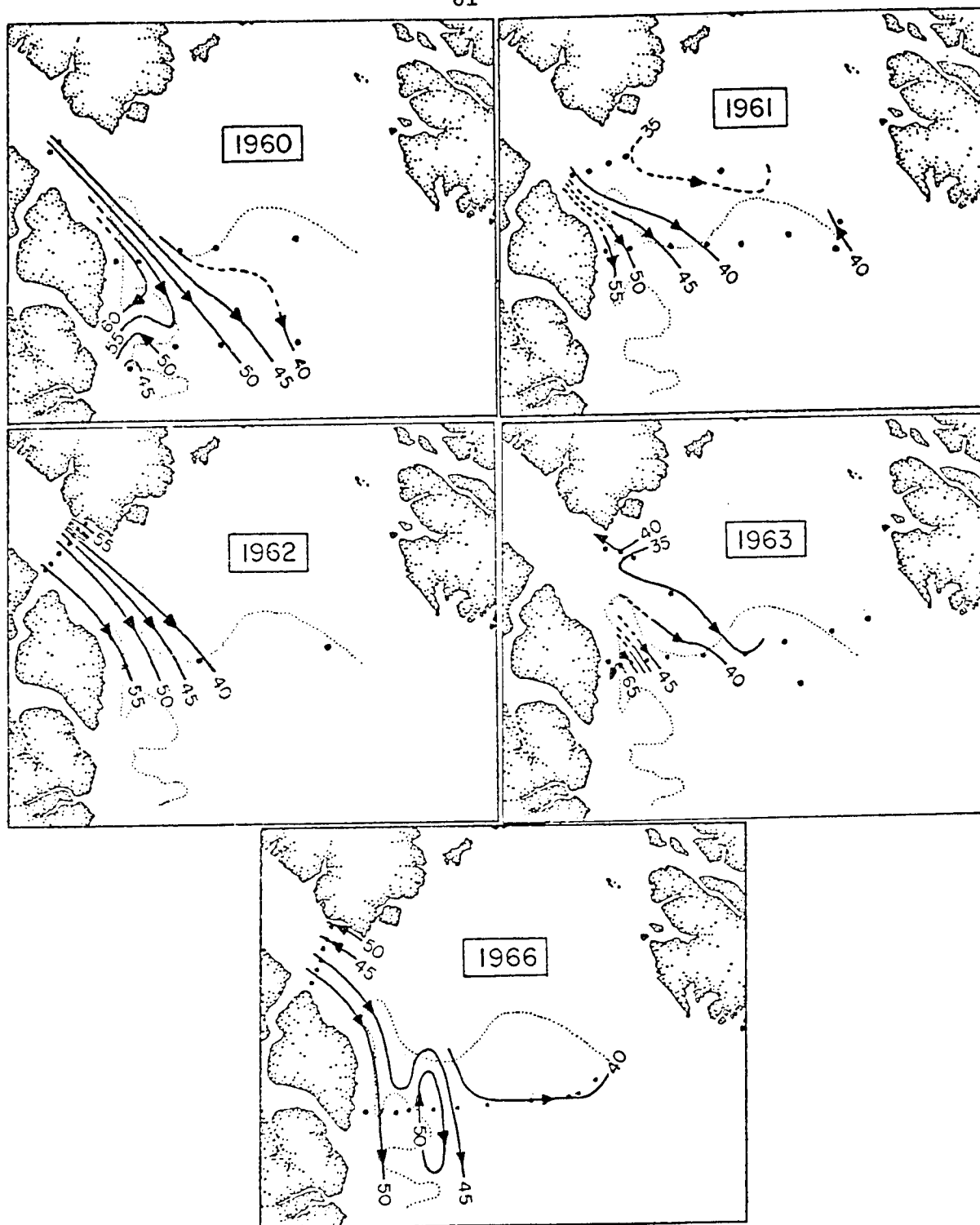


Figure 25. Dynamic topography of the surface relative to 500 db in northern Baffin Bay, with a contour interval of 5 dyn. cm. The dotted line indicates the 800 m isobath.

contour.) In 1960 an eddy occurred in the Baffin Current, east of Bylot Island, and may have been due to impingement of the current on the small

shoal area there. Presence in 1966 of an anticyclonic eddy slightly east of the shoal area suggested that an eddy similar to that observed in 1960 had formed and detached, moving eastward. The computed geostrophic currents extended deeper in 1966 than in other years, so bottom topographic influence may have been particularly large during that year and contributed to the exaggerated development of an eddy and resulting gyre. Similar eddies observed farther south in 1940 were referred to as countercurrents (Barnes 1941). The 1961-to-1963 data were insufficient to detect such eddies, although a near-shore countercurrent was apparent in 1963.

Data from northeastern Baffin Bay were insufficient, save for the 1969 data in Melville Bay discussed below, to allow construction of dynamic topographies there. The slopes of the isohalines (see, e.g., Fig. 6e), which can be assumed to represent the gross features of the internal mass distribution, suggest the presence in the northeastern region of a broad, sluggish northward flow in agreement with Figure 24. Geostrophic transport calculations (see below) also indicated a northward current which was slow relative to the southward Baffin Current.

The currents at 500 m computed relative to deeper reference levels were generally so small as to be on the order of magnitude of the measurement error. Qualitative conclusions regarding the flow at 500 m depth may be drawn, however, from the distribution of maximum temperature (Fig. 14). Due to the constancy of salinity and hence of density at the warm core, the plot showing horizontal distribution of T_{\max} approximates closely a plot of temperature on the $\sigma_t=27.6$ surface. Assuming no mixing, flow must occur along the isotherms on such an "isentropic diagram" (Montgomery 1938). Since the temperature-salinity relations indicated that mixing was, in fact,

occurring in northern and northwestern Baffin Bay at 500 m depth, flow was occurring across the isotherms. The eastward extending cold tongue off Bylot Island suggests the influence there of the Baffin Current. The apparent bottom topographic control on the current in this region also suggests that the current extends downward to the depth of the $\sigma_t=27.6$ surface (i.e., nearly to the bottom). The warmer westward tongue coincides approximately with the West Greenland Current in the region where it turns west in northern Baffin Bay, suggesting that the West Greenland Current extends to the depth of the $\sigma_t=27.6$ surface (400-500 m). The southward warm tongue in north central Baffin Bay coincides approximately with the southward turning portion of the Baffin Current, suggesting that this current, also, extends downward to about 500 m. The circulation at the warm core in Baffin Bay is therefore at least qualitatively similar to the baroclinic surface circulation.

Geostrophic transports through the indicated sections (Fig. 26) were computed for 1961-to-1963 and 1966. Data from other years were inadequate for transport calculations. A variable reference level based on the method of Defant (1961) was used; extrapolation of this reference level by the method of Helland-Hansen (1934) was necessary only in the easternmost portion of the 1966 section. Sections farther north were not used for transport calculations, since shallower depths precluded definition of a reliable reference level. Sections farther south were not used because extensive extrapolation of the reference levels onto the western Greenland banks would have been required. Heat transports were then computed using -1.85° C as a reference temperature. Heat transports below 50 m were computed in an attempt to isolate the effects of any short-term near surface

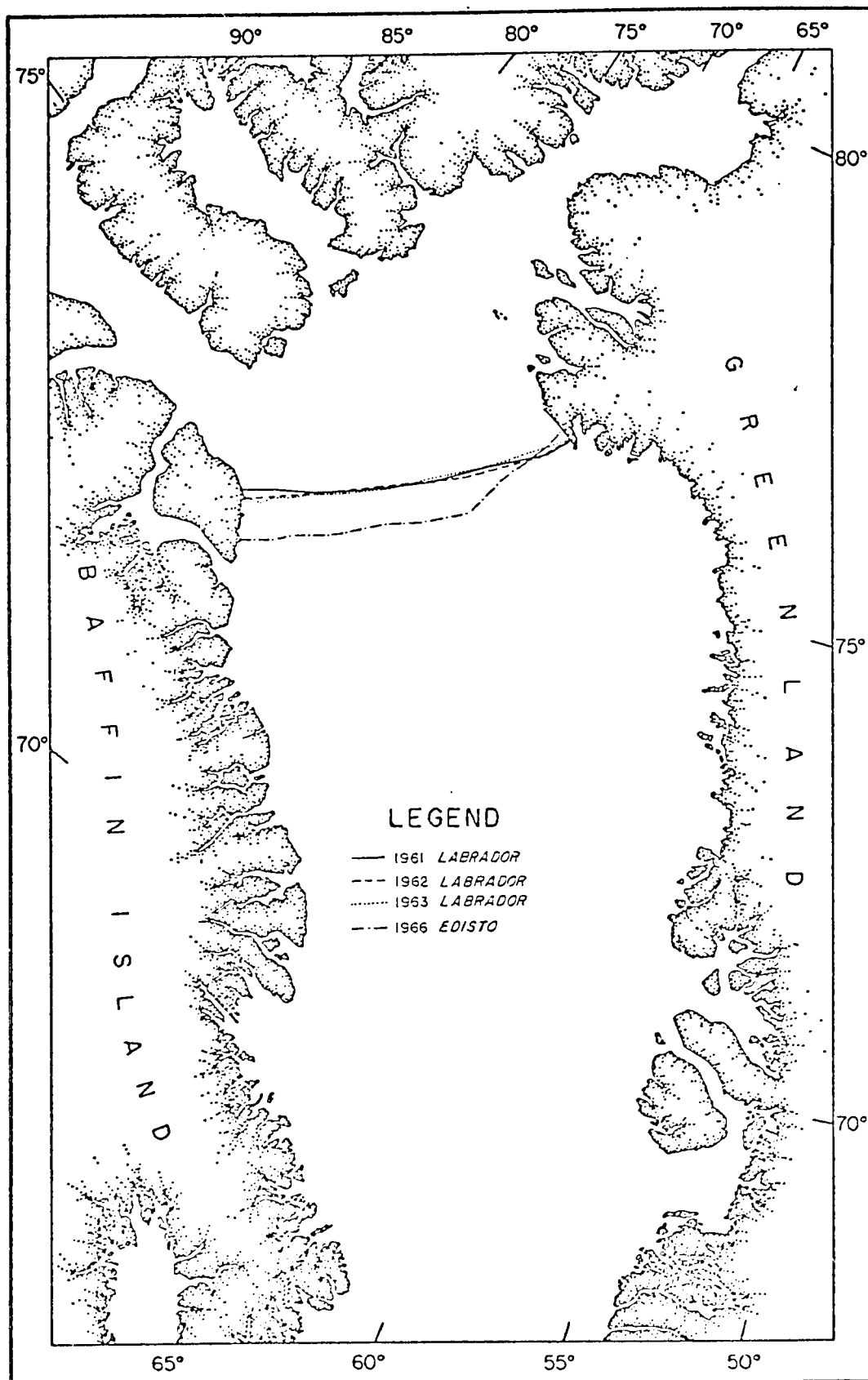


Figure 26. Geographical locations of the four sections used for computing the transports given in Table VI.

temperature variations. The resulting volume and heat transports are presented in Table VI.

Southward volume transport in the Baffin Current varied from 1.69 to $3.11 \times 10^6 \text{ m}^3/\text{sec}$, with a mean southward transport of $2.28 \times 10^6 \text{ m}^3/\text{sec}$. Northward transports off Cape York, in the northernmost part of the West Greenland Current, were smaller and varied from zero to $0.37 \times 10^6 \text{ m}^3/\text{sec}$, with a mean northward transport of $0.22 \times 10^6 \text{ m}^3/\text{sec}$. These values yield a net southward transport varying from 1.54 to $2.74 \times 10^6 \text{ m}^3/\text{sec}$, with a mean of $2.06 \times 10^6 \text{ m}^3/\text{sec}$.

The computed mean southward transport corroborates previous calculated transports. It is slightly higher than the southward transport of $1.51 \times 10^6 \text{ m}^3/\text{sec}$ estimated by Smith, Soule and Mosby (1937) using the 1928 GODTHAAB data, although their value is similar to the 1961 transport shown in Table VI. The mean transport is well above the $1.17 \times 10^6 \text{ m}^3/\text{sec}$ southward transport estimated by Vowinckel and Orvig (1962), based on calculations of the Arctic Ocean's water balance. Palfrey and Day (1968) computed a net southward transport through the 1966 section of $1.74 \times 10^6 \text{ m}^3/\text{sec}$, in good agreement with the value given for that section in Table VI. They also computed southward transports of 1.30 and $2.25 \times 10^6 \text{ m}^3/\text{sec}$ through 1966 sections located farther south in Baffin Bay. Collin (1965) arrived at a net southward transport of $1.5 \times 10^6 \text{ m}^3/\text{sec}$ using continuity arguments and computed transports through Smith, Jones and Lancaster sounds. The relatively low transport computed by Vowinckel and Orvig depended on the small difference between large numbers, and is therefore subject to greater possible error than the other values. Neglecting this value, a net southward geostrophic transport of $2.0 \pm 0.7 \times 10^6 \text{ m}^3/\text{sec}$ occurs through

TABLE VI

Computed volume transports T ($\times 10^{-6} \text{ m}^3/\text{sec}$) and heat transports Q ($\times 10^{-12} \text{ g-cal/sec}$) through sections shown on Figure 26

| Year | T (Baffin Current) | T (W. Greenland Current) | T (net) | Q (net) | Q (below 50 m) | T (below 50 m) | %Q below 50 m | %T below 50 m |
|------|--------------------|--------------------------|---------|---------|----------------|----------------|---------------|---------------|
| 1961 | -1.69* | 0.15 | -1.54 | -1.86 | -0.98 | -0.95 | 53 | 62 |
| 1962 | -2.28 | 0.00 | -2.28 | -2.34 | -1.56 | -1.67 | 67 | 73 |
| 1963 | -3.11 | 0.37 | -2.74 | -3.32 | -2.45 | -1.94 | 74 | 71 |
| 1966 | -2.04 | 0.34 | -1.70 | -2.95 | -1.81 | -1.34 | 62 | 79 |
| MEAN | -2.28 | 0.22 | -2.06 | -2.61 | -1.70 | -1.48 | 64 | 71 |

*Negative sign indicates southward transport.

northern Baffin Bay in the summer.

The total heat transports varied from 1.86 to 3.32×10^{12} g-cal/sec southward. A comparison of heat and volume transports above and below 50 m indicates that the percentage of heat transported below 50 m was generally proportional to the percentage of volume transport below that depth except in 1966. This suggests that temperature variations due to summer warming in the upper 50 m did not greatly affect the calculated heat transports except, possibly, during 1966.

A comparison of the net heat and volume transports reveals then, in order to attain a heat balance in northern Baffin Bay while neglecting surface heat exchanges, the mean temperature of the inflowing Arctic Ocean Water would have to have been -0.64° C in 1961, 0.82° C in 1962, -0.64° C in 1963 and -0.11° C in 1966. The temperatures required for a heat balance in 1961-to-1963 approximate the mean temperature of the inflowing Arctic Ocean Water. The slightly lower value for 1962 might reflect a slightly lower minimum temperature during that year (Table II). The temperature required for heat balance in 1966 was too high to be representative of the inflowing Arctic Ocean Water. The discrepancy might be explained in two ways:

- a. The 1966 data were obtained in August. Table VI indicates that an abnormally high percentage of the volume transport occurred below 50 m in 1966, while the percentage heat transport below 50 m was not correspondingly high. A net surface heat influx over an area of 10^{14} cm² (see section 4.1) might have contributed to the relatively high heat flow compared to volume transport; or

b. The Atlantic Water layer was abnormally thick in 1966 (Table V). It was also observed farther north than in earlier years (Fig. 15). It is possible, in view of the irregular nature of the flow in northern Baffin Bay, that a northward pulse of Atlantic Water occurred prior to the acquisition of the 1966 data. These data would then reflect southward flow of the Atlantic Water which had previously flowed northward. Such a northward pulse was also suggested by the presence, in Smith Sound, of water from the Baffin Bay cold core.

The latter phenomenon would have led to a higher heat content in the water column north of the transport section than in other years. Computations did not reflect this, possibly because the abnormally low temperature of the Arctic Water tended to cancel out the effect of the thicker Atlantic Water layer.

Sufficient data were acquired in Lancaster Sound during 1954, 1957, 1961 and 1962 to allow the construction of surface dynamic topographies relative to 500 db for each of those years (Fig. 27). A net easterly baroclinic flow was present during each year. A persistent meander-like feature, only poorly documented in 1954 but pronounced during the other years, consisted of a quasi-stationary northward loop in central Lancaster Sound. It appears to be related to the bottom topography represented by the 650 m isobath on Figure 27. The cyclonic curvature of the eastward flowing current core is of the proper magnitude to reflect the conservation of potential vorticity, although it is impossible to assess the importance here of frictional, non-linear and time-dependent effects. At the northernmost point in the northward loop the current core might be constrained, by a combination of the north shore of Lancaster Sound and the

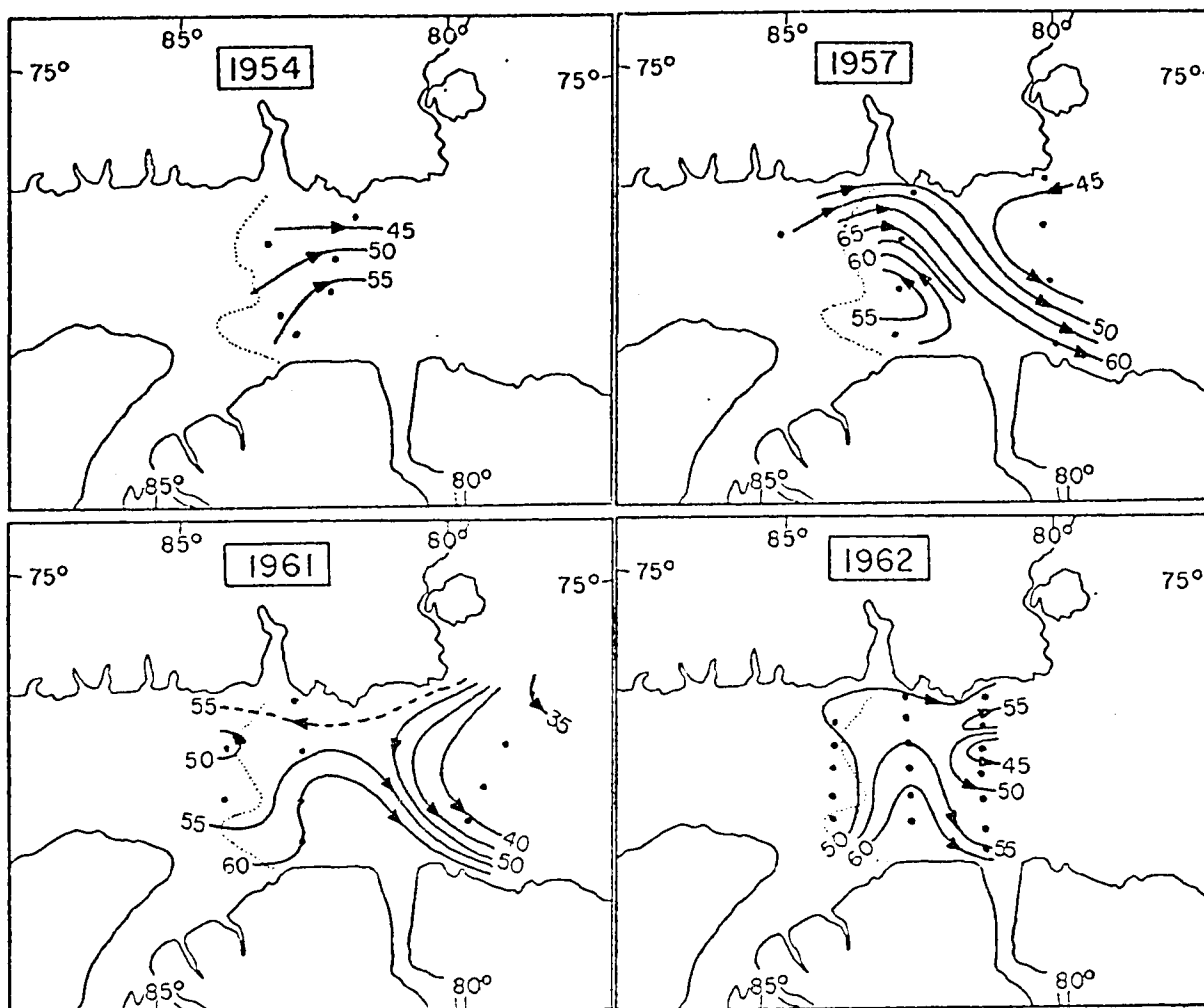


Figure 27. Dynamic topography of the surface relative to 500 db in Lancaster Sound for various years, with a contour interval of 5 dyn. cm. Dotted line is the 650 m isobath.

cyclonically circulating Baffin Bay Water to the east, to curve anticyclonically before exiting along southern Lancaster Sound. The cyclonic gyre in central Lancaster Sound in 1957 and the countercurrent along its shore in 1962 do not appear to be permanent features. Only the 1961 and 1962 data were sufficient for tracing of the Lancaster Sound current core to the Baffin Current (Fig. 25), but the overall circulation pattern suggests that the eastward currents observed in Lancaster Sound during 1954 and 1957 would ultimately enter the Baffin Current.

The cyclonic circulation in eastern Lancaster Sound appears to be a permanent feature and accounts for the presence there of Baffin Bay Water. The circulation of Baffin Bay Water westward into northern Lancaster Sound appears to be due to conservation of potential vorticity; the southward current impinges on Lancaster Sound from the relatively shallow shelf area off Devon Island (Fig. 2), and hence exhibits a tendency to turn west.

Kiilerich (1939) computed, using the 1928 GODTHAAB data, a net eastward transport through Lancaster Sound of $0.65 \times 10^6 \text{ m}^3/\text{sec}$ based on the 1000 db reference level. Collin (1963) computed an eastward transport of $1.0 \times 10^6 \text{ m}^3/\text{sec}$ relative to the 500 db level using the 1957 LABRADOR data. Palfrey and Day (1968), using 700 db as a reference level and two coincident 1966 EDISTO sections occupied over a 29 hour period, arrived at transports of $0.45 \times 10^6 \text{ m}^3/\text{sec}$ westward for the earlier occupation and $0.68 \times 10^6 \text{ m}^3/\text{sec}$ eastward for the later occupation. A recalculation, using the same data, yielded eastward transports of 0.3 and $0.6 \times 10^6 \text{ m}^3/\text{sec}$ for the earlier and later occupations and suggested that the westward transport obtained by Palfrey and Day was in error.

The relatively large variations in computed transport over a short period were suggested by Palfrey and Day to be due to tidal fluctuations in Lancaster Sound. While it seems improbable that the internal mass field would respond rapidly enough to reflect current fluctuations of tidal period, it is possible that internal waves of tidal or inertial period might occur which would strongly affect current computations. The time variations observed elsewhere in Baffin Bay suggest that such variations in computed currents may be real, but the data are insufficient for separation of real variations from artifacts due to internal waves or other sources.

Data from Lancaster Sound for other years did not allow detection of a reference level by Defant's or other methods. The internal mass distribution indicated the presence of horizontal pressure gradients at the bottom of the channel and therefore suggested that baroclinic currents extended to the bottom. Hence no further attempts were made to compute volume transport through Lancaster Sound. Direct current measurements would allow determination of the transports and their time variations, but such measurements have never been obtained.

Data from Jones Sound in 1962 and 1963 were sufficient for construction of surface dynamic topographies relative to 500 db (Fig. 28), though some extrapolation of this reference level onto shallow nearshore areas was necessary. A net westward baroclinic flow was occurring in 1962, while an eastward flow was occurring in 1963 along with a cyclonic circulation in easternmost Jones Sound. This cyclonic circulation accounts for the presence in Jones Sound of Baffin Bay Water. The westward flow in 1962 was reflected by the presence in central Jones Sound of abnormally low minimum temperatures evident in the temperature-salinity curves (Fig. 9a). An absence of Atlantic Water from Jones Sound suggests, however, that the westward flow in 1962 occurred primarily in the upper few hundred meters.

Kiilerich (1939) computed an eastward geostrophic transport through Jones Sound of $0.3 \times 10^6 \text{ m}^3/\text{sec}$, based on the 1928 GODTHAAB data. As for Lancaster Sound, he used a 1000 db reference level. Collin (1963) computed, using the 1957 LABRADOR data and a 500 db reference level, an eastward flow of $0.27 \times 10^6 \text{ m}^3/\text{sec}$. Computations by Palfrey and Day (1968), using 1966 EDISTO data from two sections occupied within a 12 hour period, yielded eastward geostrophic transports of $0.42 \times 10^6 \text{ m}^3/\text{sec}$ for the

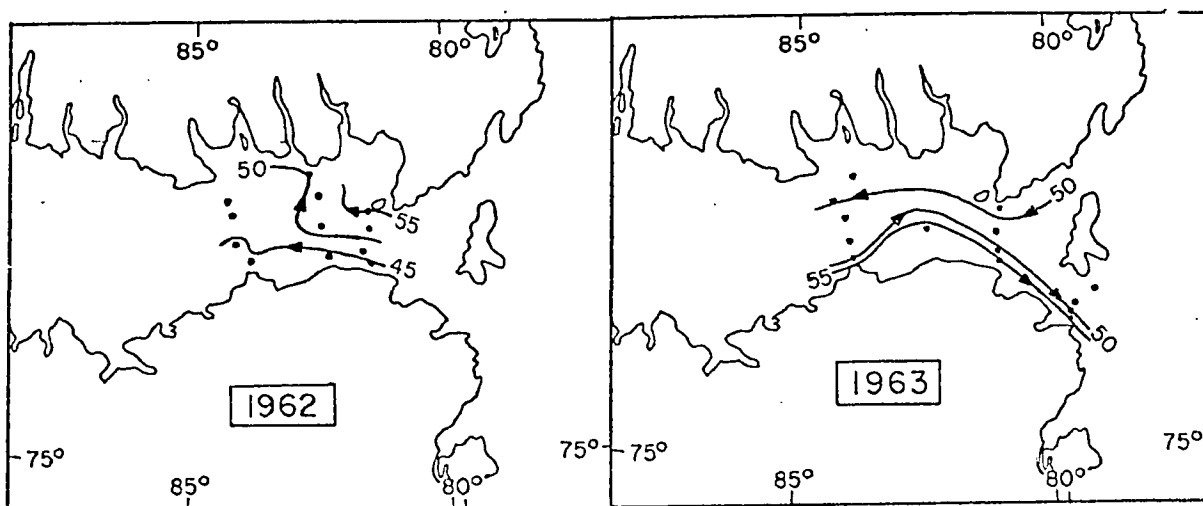


Figure 28. Dynamic topography of the surface relative to 500 db in Jones Sound for various years, with a contour interval of 5 dyn. cm.

earlier and $0.20 \times 10^6 \text{ m}^3/\text{sec}$ for the later section. Their reference level was 700 db, based on extrapolation of the reference level, determined by Defant's method, which was used in Baffin Bay.

Although the baroclinic flow through Jones Sound in 1962 was shown by the dynamic topography to be westward, it was impossible to obtain suitable reference levels for transport computations. As for Lancaster Sound, it is concluded that direct current measurements are needed to obtain better transport values and detect time variations. The net flow is eastward, but occasional westward flow appears possible.

Data sufficient for construction of a dynamic topography in the Smith Sound region were obtained only in 1968 from the CGC WESTWIND (Fig. 29). The relatively shallow, irregular bottom dictated choice of a 300 db reference level. The plot presented is from the first of two occupations of most of the stations; exceptions are those stations, which were only occupied once, in Smith Sound itself. No major changes in property distribution occurred between the two occupations, which were about a week apart,

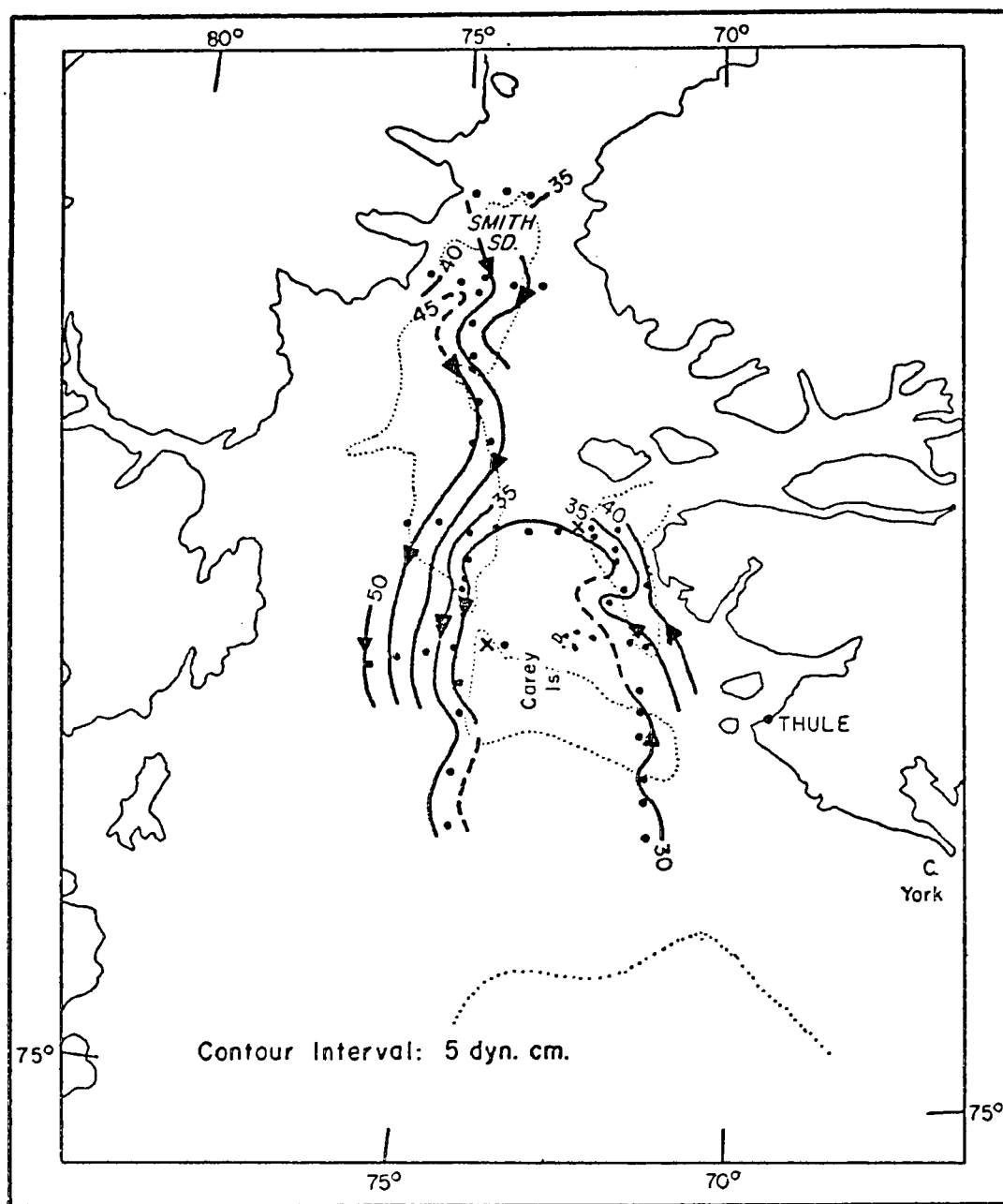


Figure 29. Dynamic topography of the surface relative to 300 db on 8-15 September 1968 in the Smith Sound region, with a contour interval of 5 dyn. cm. Dotted line represents the 500 m isobath, and locations of recording current meters are designated "x".

but minor changes were sufficient to preclude combining the data into a single plot. Data from the later occupation were inadequate for construction of a dynamic topography.

The major features of the baroclinic surface flow were a northward current east of the Carey Islands and a stronger southward current west of them. The flow through Smith Sound was southward. A poorly defined westward flow north of the Carey Islands formed a cyclonic circulation centered on them. The presence east of the Carey Islands of icebergs, which probably originated in eastern Baffin Bay, was noted from aboard the CGC WESTWIND in 1968. This suggested a northward surface flow there. Continuation of this flow to depths of 400-500 m was suggested by the presence of Baffin Bay Atlantic Water (see section 2.4) which must have come from the south. The southward baroclinic flow west of the Carey Islands was accompanied by water of the proper type to have come from Smith Sound.

Currents were measured, in 1968, at 100 m and 300 m depths north and west of the Carey Islands (Muench in press)(Fig. 29). The measurements indicated tidal and inertial currents; and currents which varied over time scales of several days from zero to about 10 cm/sec in magnitude but did not vary appreciably in magnitude between 100 and 300 m depth at a given time. These latter currents were apparently barotropic. The measured currents were variable in direction, but primarily southwesterly at the location north of the Carey Islands. West of the Carey Islands they were southerly.

The baroclinic currents, computed relative to 400 db, were small (<2 cm/sec) at the 100 m level and nonexistent at 300 m at both locations. Most baroclinicity and related current activity occurred in the upper 100 m. Below 100 m time varying, arotropic currents dominated.

While data from other years were insufficient for construction of dynamic topographies in the southern Smith Sound region, they were sufficient to suggest on the basis of dynamics that a cyclonic baroclinic circulation in southern Smith Sound coupled with a southward flow through

Smith Sound was occurring also in 1928, 1962, 1963 and 1966. The presence of Atlantic Water well northward into southeastern Smith Sound in 1963 and 1966 (Fig. 15) also suggested a northward flow similar to that occurring in 1968. The 1961 data suggested a sluggish southward flow throughout southern Smith Sound and gave no evidence of a cyclonic circulation there. A lack of Baffin Bay Atlantic Water in southeastern Smith Sound in 1961 tends to substantiate this flow pattern. The northward flow in 1962 appeared weak relative to that in 1963, 1966 and 1968, which might explain the absence of Atlantic Water from southeastern Smith Sound in 1962.

The irregularity of the bottom topography, as represented by the 550 m isobath (Fig. 29) made it difficult to determine the extent of bathymetric control on the currents in the southern Smith Sound region. A tendency for the southward current in northern Smith Sound to turn east may have been due to the course of the current into deeper water. The horizontal meander-like pattern might then have been consequent to the initial eastward displacement in northern Smith Sound. More detailed data are needed, however, to determine the possibly time-dependent nature of this pattern.

The 1969 WESTWIND data allowed construction of surface dynamic topographies relative to 400 db in the southern Smith Sound-Melville Bay region (Fig. 30). On the earlier occupation, the major baroclinic flow feature was a relatively intense westerly current hugging the coast south of Cape York. A westerly surface current was suggested, also, by observations in 1969 of a high concentration of icebergs south of Cape York; these probably originated to the south along the western Greenland coast. The dynamic topography of the 200 relative to the 400 db surface was similar to the topography shown in Figure 30, suggesting that the baroclinic currents

persisted to 200 m depth. The tongue-like westward extension of warm water (Fig. 21) also suggested occurrence there of a westward current at several hundred meters depth.

The eastern portion of the 34 dynamic centimeter isoline (Fig. 30, upper) appeared to define an anticyclonic meander. A southward surface flow coinciding with the southerly branch of this meander was also suggested by a "plume" of icebergs observed extending southward from the vicinity of Cape York in 1969. The southward branch of such a meander might have contributed to the historical belief (section 1.1) in a southerly current originating off Cape York. Its presence at 300-400 m depth is suggested by a southwestward bulge in the 1.75°C isotherm on the $\sigma_t=27.5$ surface (Fig. 21), although this is not clearly defined.

The 32 dynamic centimeter isoline (Fig. 30, upper) suggests a cyclonic circulation extending towards the Carey Islands as in 1968, although the 1969 data were insufficient for delineation of the feature.

The data acquired on 27 September-2 October (Fig. 30, lower) allowed tracing of the westward current, observed earlier south of Cape York, northward to east of the Carey Islands where it had become northerly as in 1968. The 32 dynamic centimeter isoline indicates a complex configuration for the cyclonic eddy noted on the earlier date. The accuracy of the 1969 temperature and salinity measurements was only $\pm 0.05^{\circ}\text{C}$ and $\pm 0.05\text{ }^{\circ}/\text{oo}$, however, therefore the details in the pattern of the 32 dynamic centimeter isoline may or may not be real. A southerly flow west of the Carey Islands, was accompanied by lower water temperatures, suggesting that a portion of the water there was from Smith Sound. The suggestion of a countercurrent in the extreme western portion of the southerly current was defined by a single station; since such a feature has not been previously observed it

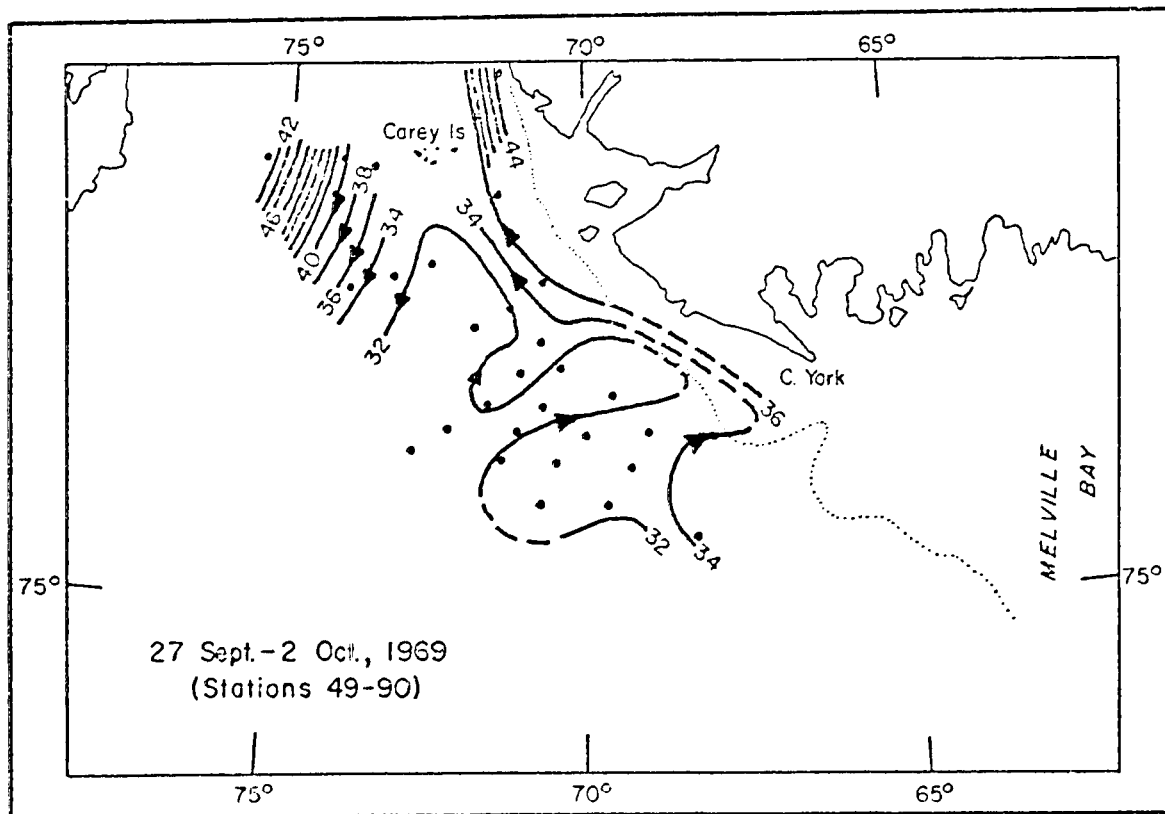
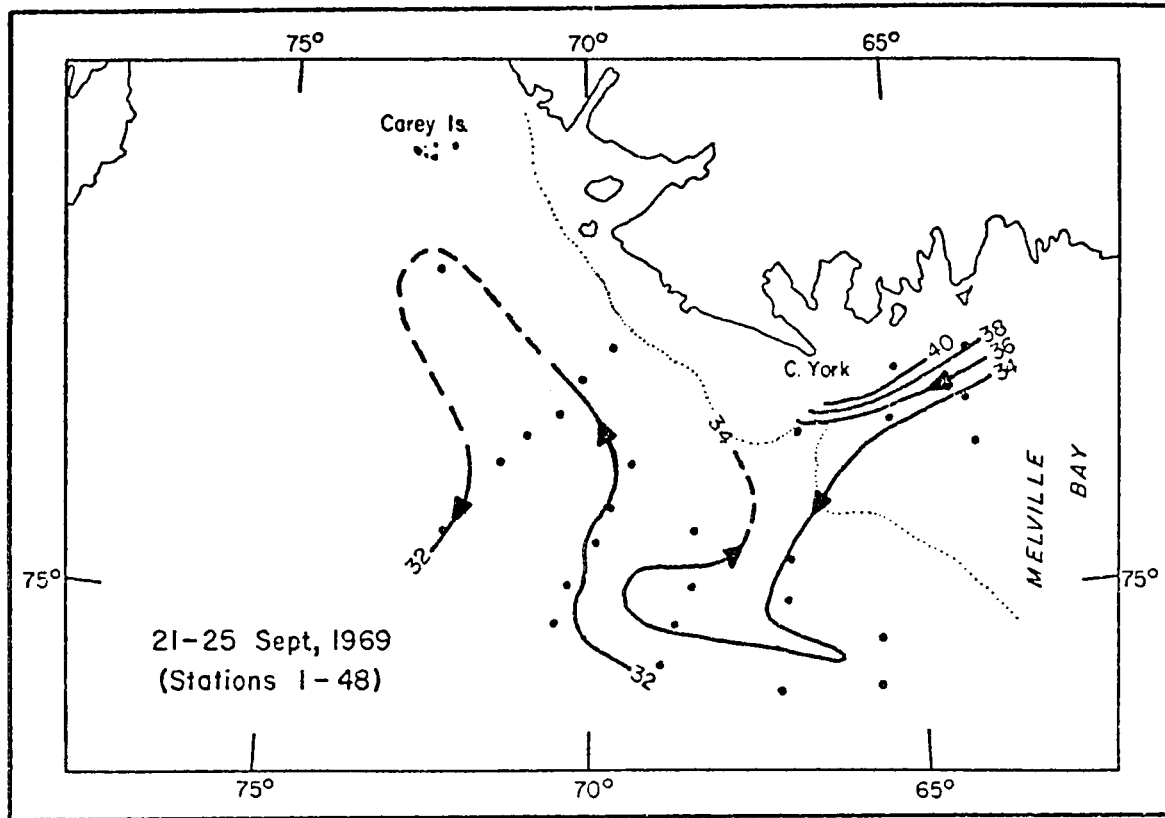


Figure 30. Dynamic topography of the surface relative to 400 db in the southern Smith Sound-Melville Bay region, with a contour interval of 2 dyn. cm. Dotted line indicates the 250 m isobath.

should be viewed with caution.

Time changes between the earlier and later occupations of the stations were sufficient, in cases where these stations overlapped, to preclude combination of the data. Time changes during the carrying out of each leg of the cruise were probably sufficient so that many of the depicted details in Figure 30 should be regarded with caution. The westerly flow south of Cape York, which was observed in 1928, 1964 and 1966 (being clearly absent only in 1962), and the southward flow west of the Carey Islands, which was evident from all data from that area, are felt to be persistent features.

The apparent vertical extent of the currents in the Cape York region to great depths would be expected to result in bottom topographic influence. The 34 and 36 dynamic centimeter isolines (Fig. 30 upper) south of Cape York delineate a zone of divergence. This might be due to a tendency for the northern portion of the westward current to parallel the coast, while the southern portion parallels a 250 m isobath and turns southward. Such a divergence may have contributed to upwelling which led, in turn, to the relatively high surface salinities noted in the area (section 2.4). More quantitative analysis of the phenomenon is precluded by the inaccuracy of the 1969 data and by the time changes in the system.

Geostrophic transports through Smith Sound have been computed by various writers (Kiilerich 1939; Bailey 1956; Collin 1963; Collin 1965) and the results compiled and enlarged upon by Muench (1966), who concluded that a net southerly transport of about $0.2 \times 10^6 \text{ m}^3/\text{sec}$ occurred. This transport was suspected to be, however, subject to large variations even including northward flow.

Flow variability in Smith Sound and Nares Strait has been relatively well documented. Nutt (1966) deduced a high variability, including flow

reversal, from ice movements. Current measurements made during August 1963 in Smith Sound (Palfrey and Day 1968) indicated a weak (5-10 cm/sec) net southward flow with semidiurnal tidal components at times as great or greater than the mean flow. Current measurements made in western Kane Basin in May 1969 (Tidmarsh et al. 1969) indicated a similar pattern. Reversals of the current leading to northward flow were observed, but the data were insufficient to detect any periodicity in these other than their coincidence with northward tidal currents. Palfrey and Day (1968), using the 1966 EDISTO data, computed geostrophic currents through Kane Basin and southern Smith Sound; their transports varied from zero to $1.26 \times 10^6 \text{ m}^3/\text{sec}$ southward, and they suggested a pulsating flow on the basis of these transports. Such computed currents might have been strongly affected, however, by lack of a suitable dynamic reference level, time dependant effects and presence of internal waves. Current measurements would allow delineation of the flow through Smith Sound.

Deficiencies in the geostrophic method of computing accurate volume transports are pointed out by the lack of volume continuity implied by the existing transports for Smith, Jones and Lancaster sounds and northern Baffin Bay. Observed time changes suggest that neglect of the time varying terms, integral to the geostrophic approximation, is unjustified. Barotropic transports, neglected in geostrophic calculation, may be significant. According to Table VI a net inward flow via the three sounds of about $2 \times 10^6 \text{ m}^3/\text{sec}$ is required to satisfy volume continuity. The computed transports for Smith, Jones and Lancaster sounds yield, however, net inflows varying from $0.2 \times 10^6 \text{ m}^3/\text{sec}$ to $1.6 \times 10^6 \text{ m}^3/\text{sec}$, with a probable mean inward transport of about $1.0 \times 10^6 \text{ m}^3/\text{sec}$. Direct current measurements

are needed, as it is unlikely that further geostrophic computations would improve this agreement.

Tides are of relatively little importance to the mean circulation in northern Baffin Bay. They may, however, cause internal waves which would affect geostrophic computations. As noted above, semidiurnal tidal currents have been identified in the Kane Basin and southern Smith Sound regions. A theoretical study of the tides in Baffin Bay by Godin (1966) predicted the occurrence of semidiurnal and diurnal tides, but his model was inadequate in the peripheral regions and sounds where currents have been measured. The observed tidal currents are generally of greater magnitude than the observed mean currents, but the current measurements are from areas where all evidence suggests relatively sluggish mean currents.

In summary, recent data corroborates the cyclonic circulation pattern, in northern Baffin Bay delineated by the 1928 and 1940 data. Northward transport in the West Greenland Current in northern Baffin Bay is far smaller than southward transport in the Baffin Current, volume continuity being satisfied by a net inflow of Arctic Ocean Water via Smith, Jones and Lancaster sounds. Considerable complexity is present in the northernmost portions of the Baffin and West Greenland currents, and the entire system is characterized by non-steady-state behaviour.

3.2 Driving forces for the circulation.

Primary features of the observed circulation in the northern Baffin Bay region are: (a) The cyclonic circulation within northern Baffin Bay itself; and (b) The net southward transport through the northern Baffin Bay system. Possible driving forces for the cyclonic circulation interior

to Baffin Bay will now be considered.

The atmospheric pressure distributions over the northern Baffin Bay region during the late summer, when most of the oceanographic data were acquired, suggest the presence of a cyclonic wind curl resulting from geostrophic winds (Fig. 31a, upper; Fig. 31b, lower). If it is assumed (Aagaard 1969) that the geostrophic wind speed and direction approximately equal the surface wind speed and direction, and that the wind stress τ_w exerted on the water surface by the wind is expressed by

$$\tau_w = \rho_a C w^2$$

where: $C =$ a transfer coefficient $= 2 \times 10^{-3}$;

$\rho_a =$ the air density; and

$w =$ the wind speed,

the resultant wind stress curl can be estimated from the atmospheric surface pressure distribution. If lateral stresses, time dependency and non-linear terms in the equations of motion are neglected, the Sverdrup transport M_y due to wind stress curl may be expressed by

$$M_y = -\text{curl } \tau_w / \beta$$

where: $\beta = df/dy$, $f =$ Coriolis parameter and $y =$ north-south coordinate.

Computations of the Sverdrup transport using the wind stress curl for the period July-September, as estimated from Figures 31 and 32, yield a northward transport, integrated across the width of Baffin Bay, of about $10^7 \text{ m}^3/\text{sec}$. The northward baroclinic transports in the West Greenland Current in northern Baffin Bay varied from zero to $0.37 \times 10^6 \text{ m}^3/\text{sec}$; it is

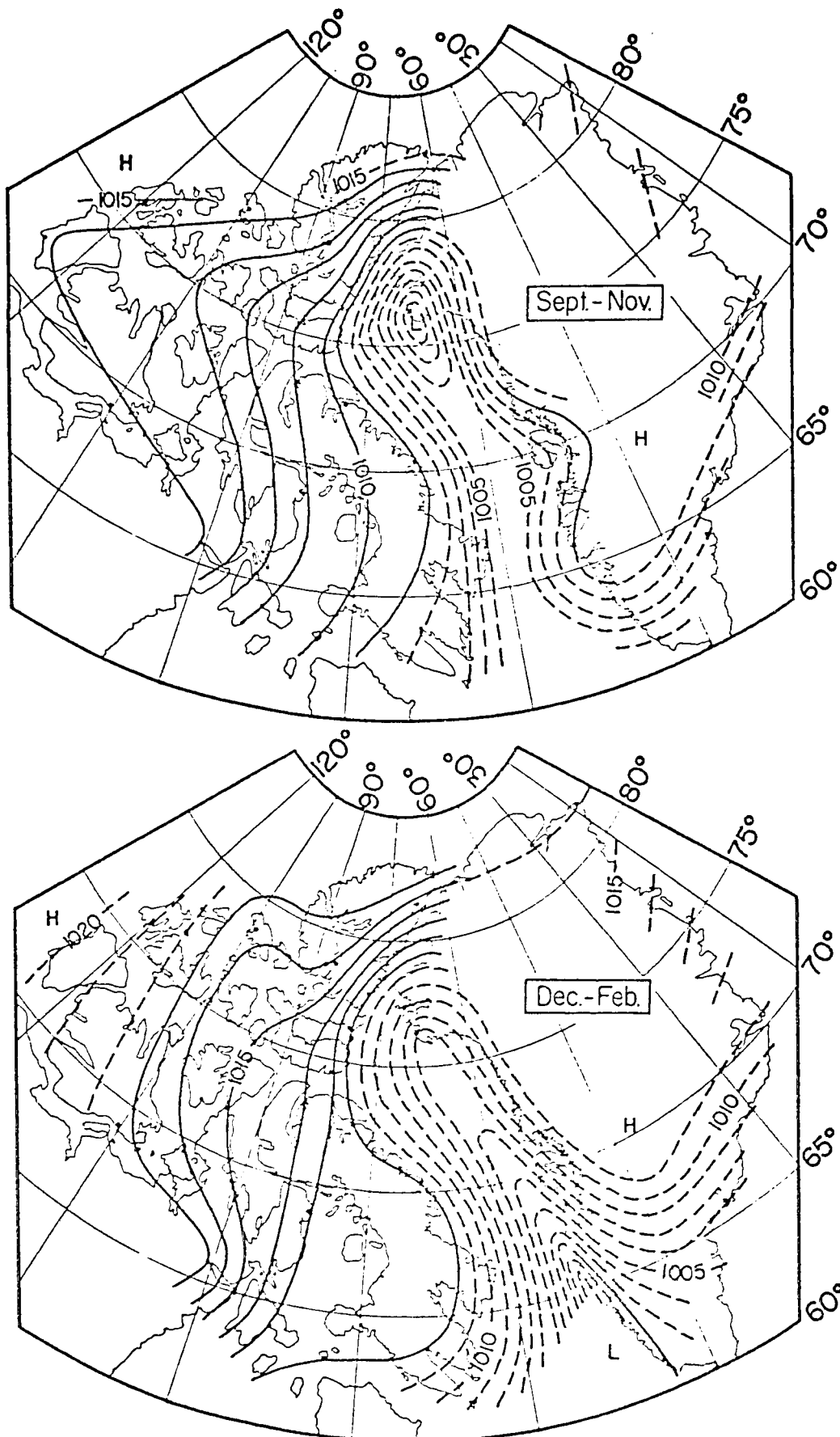


Figure 31a. Quarterly mean atmospheric surface isobars; contour interval 1 mb.

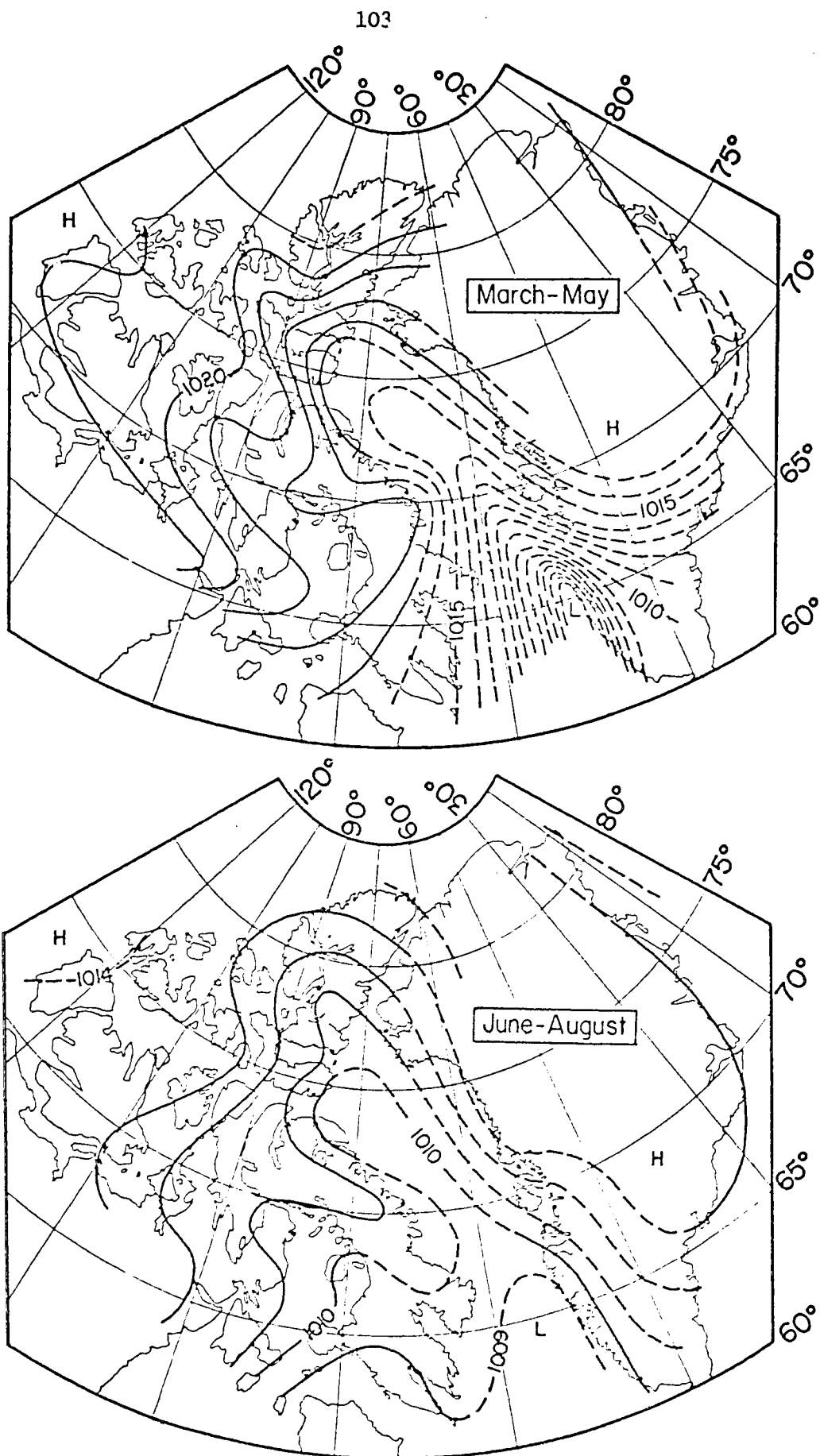


Figure 31b. Quarterly mean atmospheric surface isobars; contour interval 1 mb.

apparent therefore that the integrated Sverdrup transport is not a realistic figure. The discrepancy is suspected to lie in the neglect of frictional, time-dependent and non-linear terms in the equations of motion; each of these terms might be expected to be significant in a relatively confined region such as Baffin Bay, although data are insufficient to allow assignation of numerical values to them.

Considering the tendency of the atmospheric isobars to parallel the course of the West Greenland Current, it is possible that an Ekman current system is established by the southerly winds and contributes to the West Greenland Current. Application of the above assumptions regarding stress transfer through the air-sea interface to the June-August atmospheric isobars yielded a northward deep current (i.e. geostrophic current between the surface and bottom Ekman layers) of about 10 cm/sec. This is of the same order as the baroclinic current speeds in the West Greenland Current. A rigorous comparison is impossible, however, since the deep current is essentially barotropic while only the baroclinic currents were determined from oceanographic data from the West Greenland Current. A barotropic deep current would be undetectable from temperature and salinity data, and would be superposed on the baroclinic currents. None of the data indicate the presence along the western Greenland coast of coastal downwelling which would be expected in such a current system. It is concluded that, although Ekman coastal current systems are a possible contributing factor to the West Greenland Current, the available data do not fully support their presence.

The wind stress appears, on the basis of the atmospheric pressure distribution, to be higher in the winter than in the summer by about a factor

of two, with a consequent quadrupling of the wind stress curl. Insufficient knowledge is available, however, concerning stress transfer through pack ice to allow a quantitative estimate of wind-driven transports during the winter months. Both the mobility of the pack ice and the parameters governing air-sea momentum exchange through ice are uncertain.

During 1928, 1964, 1966 and 1969 a surface wedge of relatively low salinity water up to 50 m thick was observed extending eastward 15-30 nautical miles from the western Greenland coast (see, e.g., Fig. 6d). This wedge was particularly pronounced along the periphery of Melville Bay and south of Cape York. The subsurface slope of the isohalines and hence the isopycnals, due to the presence of the surface wedge of low salinity and hence low density water, appeared to be related to the baroclinic current flowing northward and westward in the Melville Bay region and off Cape York in 1928, 1964, 1966 and 1969 (see, e.g., Figs. 24 and 30). In 1962 the low salinity wedge was absent, while data from other years were insufficient to determine whether or not the wedge was present. The 1968 data were too far north to detect such a wedge in the Cape York region, but the presence east and north of the Carey Islands of a low salinity surface layer (indicated on Figure 19 as a warm surface layer between stations 36 and 40) suggests that the wedge may have been present farther south and spread out to form a layer after rounding Cape York.

In 1964 the entire West Greenland Current underlay the low salinity wedge, while in 1962 both wedge and current were absent. In 1966 northward flow beneath the wedge comprised approximately 15% of the total northward flow in the West Greenland Current.

Approximate computations using the computed geostrophic current velocities for the low salinity wedges and the precipitation figures from Bader

(1961) suggest that sufficient runoff would be available each summer, from the western Greenland near-coastal ablation zone, to maintain a wedge similar to that observed for 2 to 3 months. It is possible that the near-shore wedge, due to admixture of runoff from western Greenland, provides a baroclinic driving mechanism for the northward near-shore current commonly observed along the western Greenland coast. Since this current appears to be a primary means for advection of warm Atlantic Water northward toward Smith Sound, some correlation between the presence of the wedge and northward penetration of Atlantic Water (Fig. 15) might be expected. In 1966 and 1969, when a pronounced wedge was present, the Atlantic Water appeared well northward. In 1968 the layer appeared to have dissipated and spread out, so that it was impossible to determine whether or not a well defined wedge had been present farther south, while in 1969 the wedge was still well defined east of the Carey Islands and Atlantic Water extended well north. In 1928, however, presence of a strong near-shore current and wedge south of Cape York did not result in an observed extension of the Atlantic Water toward Smith Sound. Possibly the wedge has a general tendency to spread out into a layer in the vicinity of Cape York, leading to a decay of the underlying current core.

On the basis of the extreme cases, i.e. in 1962 when neither the wedge nor the West Greenland Current appeared and in 1964 when the wedge appeared to cause most of the flow in the West Greenland Current, it appears that the runoff from western Greenland can exert considerable influence on the northward flow of water in northern Baffin Bay.

The dynamics of the Baffin Current are probably somewhat different. Wind-driven currents may be significant; for example one similar to that

computed for the West Greenland Current with summer speeds on the order of 10 cm/sec. The concentrated nature of the Baffin Current suggests that westward intensification processes may be significant. The dynamics of northern Baffin Bay are complicated, however, by large inflows from the north, east and west and by complex bathymetry. Computations suggest that the intensification farther south of the Baffin Current is of the proper order to conserve absolute vorticity. The relatively low salinity near-surface layer of Arctic Ocean Water which originates from Smith, Jones and Lancaster sounds is a prominent feature in northwestern Baffin Bay. This is either a result of or contributes to the observed depression of the isohalines (hence the isopycnals) in the western portion of the Bay (see, e.e., Fig. 6e) and may play a part in the baroclinic pressure field associated with the Baffin Current.

Explanation of the net southward flow through northern Baffin Bay requires consideration of oceanographic and meteorological factors outside the immediate northern Baffin Bay region. The wind field deduced from the atmospheric surface isobars (Figs. 31a and 31b) offers no suggestion that wind stress is a factor in the net southward transport. The year-round prevailing winds in the Canadian Islands would be expected, conversely, to drive a westward rather than an eastward transport through Jones and Lancaster sounds. It is improbable that a southward transport is wind-driven through Nares Strait in the winter, because this strait is covered by fast ice which would be expected to markedly inhibit transfer of wind stress to the water. In summer, when the ice in Nares Strait breaks up, a southeasterly wind would be expected to drive a northward current there, in contradiction to the observed southward flow.

The southward transport through the system must be due, therefore, to an internal driving mechanism. Since the 250 m deep sill in Kane Basin represents the deepest channel between the Arctic Ocean and Baffin Bay, consider the dynamic depth anomalies of the upper surfaces of the Canadian and Eurasian Basins of the Arctic Ocean and of Baffin Bay relative to 250 db (Table VII).

TABLE VII

Dynamic depth anomalies with respect to the
250 db level

| Location | D (dyn.m.) | Year(s) | Platform | Stations |
|-------------------------------|------------|---------|----------|--|
| Arctic Ocean (Canadian Basin) | 0.57 | 1965 | T-3 | 1, 4, 5, 13, 30, 35 |
| | | 1966 | T-3 | 6, 8, 12, 15, 22, 50, 54, 61, 63, 68, 72, 76, 84, 90 |
| | | 1967 | T-3 | 5, 7, 10, 15, 20 |
| Arctic Ocean (Eurasian Basin) | 0.27 | 1964 | ARLIS-II | 1, 15, 35, 55, 75, 95, 119, 138, 157, 171, 191, 211, 221, 241, 259 |
| Northern central Baffin Bay | 0.25 | 1961 | LABRADOR | 83, 84, 86, 87, 92 |
| | | 1962 | LABRADOR | 53, 58 |
| | | 1963 | LABRADOR | 90, 91, 92, 99 |
| | | 1966 | EDISTO | 64, 65, 66, 67, 80 |

If the 250 level is accepted as a reference level, the sea surface is seen to be higher both in the Eurasian and Canadian Basins than in northern Baffin Bay. The surface level difference in the case of the Eurasian Basin is small relative to that in the case of the Canadian Basin. If these

sea level differences are real, i.e. if the 250 db surface is level, they would cause a southward flow by providing a horizontal pressure gradient directed into Baffin Bay. The small pressure gradient across the length of Nares Strait might be overridden by opposing atmospheric pressure gradients and lead to the irregular flow, with reversals in direction, observed. It is possible to draw correlations, in the single case where sufficient data are available, between flow reversals and atmospheric pressure variations across Nares Strait.

From late July to late September 1963 an account was kept of the movements in Nares Strait of ice island WH-5. These movements were assumed to indicate motions in the surface water layers (Nutt 1966). Coincident with the movements of WH-5 in Nares Strait, atmospheric pressures were obtained from Thule and from Alert (at the northern end of Nares Strait) which allowed deduction of atmospheric pressure variations are plotted as functions of time, for purposes of comparison, in Figure 32.

Generally good correlation was obtained between the occurrence of northerly surface flow and northward directed pressure pulses. On 14-23 July, 10-13 August and 20-23 August, an approximate agreement between northerly flow and northward pressure pulses was evident. Southerly pressure pulses on 23-29 July and 23-29 August coincided roughly with southerly flow. It was impossible to determine, due to lack of observations, whether the northerly pressure pulses on 30 July-2 August and 20-23 August resulted in northerly flow of WH-5. Little net motion occurred during the periods 14-19 August and 23 August-21 September; lack of detailed observation of WH-5 during these periods precludes correlation with the pressure variations.

Flow reversals in Nares Strait due to the observed pressure variations

are feasible; the southward surface pressure gradient due to the sea surface slope is about 2×10^3 dynes/cm², while the atmospheric variations are on

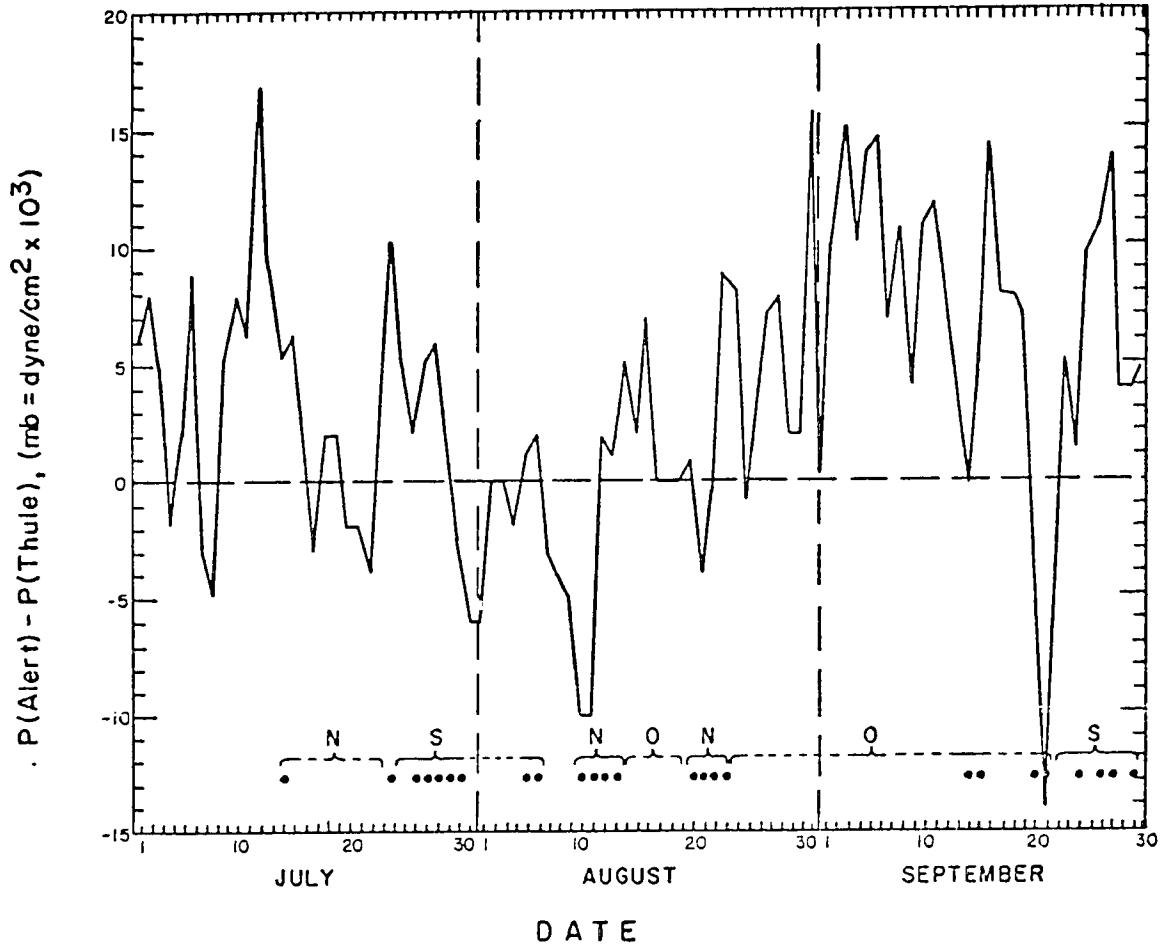


Figure 32. Mean daily atmospheric pressure differential between Alert (at north end of Nares Strait) and Thule (at south end of Nares Strait) during July-September 1963. Black dots with lettered brackets indicate duration and direction of flow of surface water deduced from motion of ice island WH-5 (N-northward flow, S-southward flow, O-no net flow.).

the order of 10 to 15×10^3 dynes/cm² and are more than sufficient to affect the surface flow direction. The relatively large sea level difference across Lancaster Sound would be expected, conversely, to drive a consistently eastward flow, as observed. Jones Sound might be expected to exhibit a flow behavior intermediate between Smith and Lancaster sounds.

The baroclinic pressure gradients relative to 250 db are maximum at the surface and decrease to zero at 250 db (approximately 250 m). Flow due to atmospheric pressure variations is barotropic in nature, however, and would extend undiminished to the bottom of the channel. Atmospheric pressure differences would therefore be expected to have a relatively large influence on the net transport in the deeper layers of Nares Strait. Since the deeper layers north of Kane Basin are a possible source of water of the proper type to contribute to Baffin Bay Deep and Bottom water, atmospheric pressure differentials might be expected to exercise some control over the southward flow of these waters.

Inasmuch as insufficient data are available to determine whether the 250 db surface is level, the merits of the above hypothesis must rest on its agreement with observed phenomena. A levelling survey northward along Nares Strait and westward along Lancaster Sound would be necessary to finally resolve the problem of sea surface slope.

In summary, the observed features of the circulation of northern Baffin Bay may be explained by a combination of baroclinic and barotropic driving mechanisms. The cyclonic circulation appears to be driven by a combination of wind stress and the near-surface low salinity wedges due to melt water; in northwestern Baffin Bay the relatively low density Arctic Ocean Water also plays some role in the baroclinic field associated with the circulation. The net southward flow through the system appears to be driven primarily by a difference in sea level between the Arctic Ocean and Baffin Bay which is due to differences in internal mass distribution. The observed irregularity of flow in Nares Strait may be explained in terms of atmospheric pressure variations overriding the southward pressure gradients due to a relatively small sea surface slope.

IV. THE NORTH WATER

4.1 A description of the phenomenon

The North Water is a large polynya located in northern Baffin Bay (Figs. 1, 3). It is noteworthy in that it persists throughout the winter despite the fact that atmospheric conditions would appear to dictate formation of 1.5 to 2m ice cover similar to that found on the surrounding water.

The existence of the North Water has been documented since the 17th century (see section 1.1). Eskimos in the Thule region took it for granted and used it as a winter hunting ground, which suggests its presence prior to discovery by European explorers. R. M. Koerner (personal communication) has suggested, on the basis of snow accumulation on the Devon Island Ice Cap, that open water has occurred in northern Baffin Bay during the winter for the past several hundred or possibly several thousand years. The North Water is, then, an essentially permanent feature of the eastern Arctic.

Available information, in the form of air reconnaissance and satellite observations, concerning the geographical extent of the North Water have been compiled by Moira Dunbar (1970) (Fig. 33). Outstanding features were suggested by her to be the relatively stable northern boundary, the stable ice edge along Ellesmere Island and the extensions into Jones and Lancaster sounds. Observed year-to-year variations in the southern boundary were generally of the same magnitude as monthly variations. All but the extreme northern portions have been observed to contain varying concentrations, as high as 9/10, of young ice. Ice age and concentration both increase from north to south, making it difficult to locate unambiguously the southern boundary. The northernmost portion of the North Water, in northern Smith

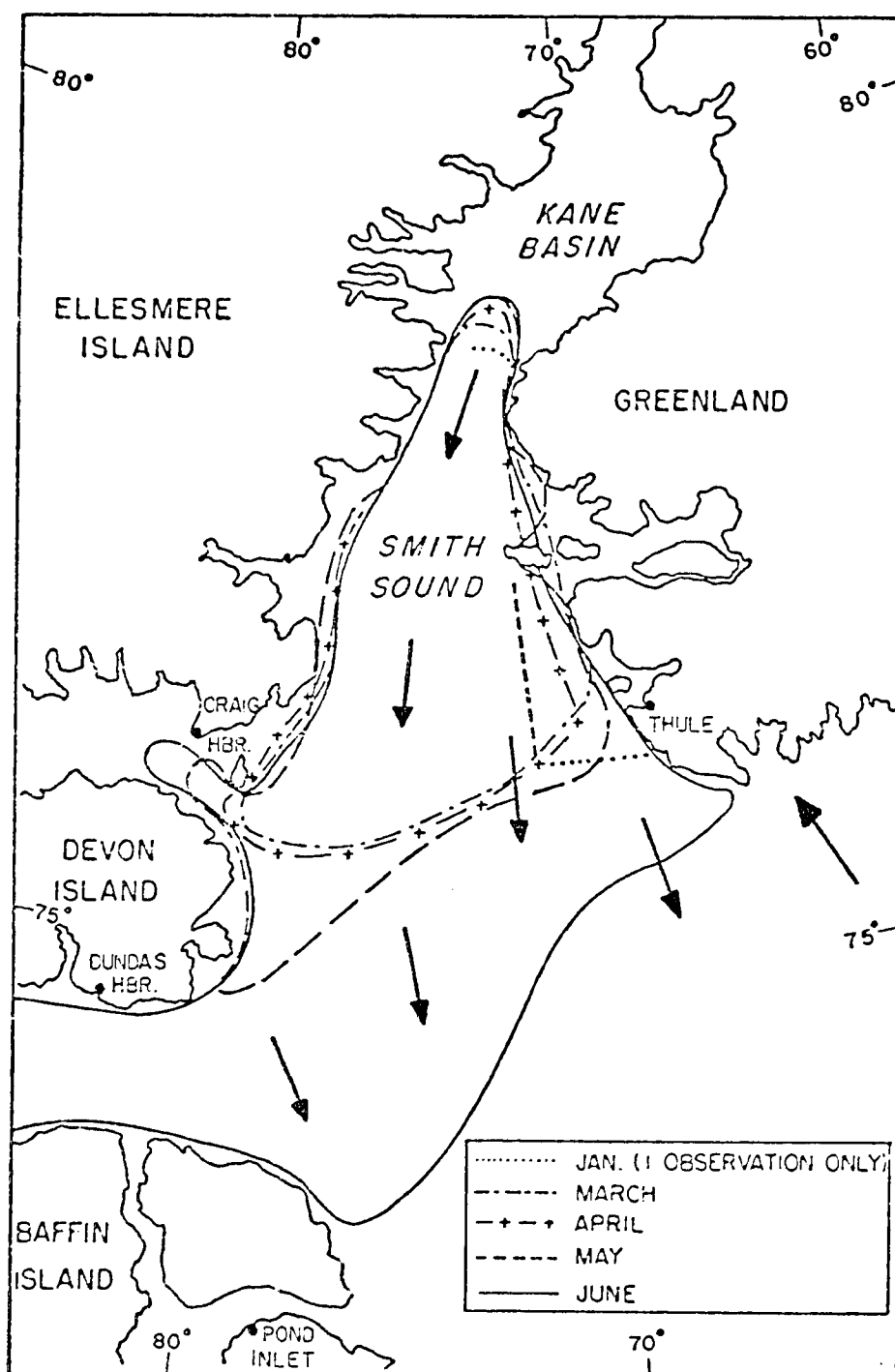


Figure 33. Approximate monthly mean extent of the North Water (from Dunbar 1970), indicating locations of meteorological stations. Arrows indicate mean directions of prevailing winter (September-February) winds. Heavy dotted line is reference for computation of Ekman upwelling (see text).

Sound, exhibited the most frequent open water and lightest ice cover.

Insufficient data are available to establish the date on which the North Water first becomes a distinguishable feature, although this is suspected to occur in October with the advent of winter ice formation. Winter observations are few, due primarily to darkness. The extent of the North Water from the time of its formation through March is therefore, save for the single observation in January, uncertain. During the late spring break-up in northern Baffin Bay, the open water enlarges into Jones and Lancaster sounds. During some years, eastward extension into Melville Bay has also been observed. The North Water loses its identity in July-August when the ice plug in northern Smith Sound disintegrates, at which time northern Baffin Bay becomes part of the drift route for south-flowing ice from Nares Strait.

4.2 Some possible causes

Previously proposed theories concerning the formation and maintenance of the North Water (section 1.1) were generally speculative, qualitative and based on few, if any, data. Recent data allow the more quantitative investigation of some possible mechanisms.

In order to prevent winter ice formation, the heat flowing upward through the air-sea interface must be replenished from below. Using atmospheric data from the four stations indicated on Figure 33 and making certain assumptions about surface oceanographic and atmospheric conditions, the monthly mean heat fluxes through an open water surface in the North Water region have been computed (Appendix B). The results (Table VIII) agree well with those of Badgley (1966) obtained from measurements over an open

TABLE VIII

Computed* monthly mean surface heat fluxes through an open water interface in the North Water region ($\text{g-cal/cm}^2\text{-sec} \times 10^3$)

| Month | $Q_i + Q_b$ | Q_e | Q_h | NET Q |
|-----------|-------------|-------|-------|-------|
| January | 2.0 | 3.1 | 9.5 | 14.6 |
| February | 2.0 | 3.0 | 10.0 | 15.0 |
| March | 1.1 | 2.9 | 8.3 | 12.3 |
| April | -1.1 | 2.3 | 5.2 | 6.4 |
| May | -2.7 | 0.8 | 0.9 | -1.0 |
| June | -3.7 | 0.5 | 0.0 | -3.2 |
| July | -3.0 | 0.2 | -0.8 | -3.6 |
| August | -1.7 | 0.3 | -0.2 | -1.6 |
| September | 0.2 | 1.2 | 0.7 | 2.1 |
| October | 1.0 | 1.9 | 2.3 | 5.2 |
| November | 1.6 | 2.5 | 5.3 | 9.4 |
| December | 2.0 | 3.0 | 8.9 | 13.9 |

*See Appendix B for computations.

lead in Arctic pack ice. Prominent features of the heat fluxes are the relatively large (up to $15 \times 10^{-3} \text{ g-cal/cm}^2\text{-sec}$) and constant upward fluxes during December-March; these are primarily a consequence of low air temperatures resulting in a large sensible heat flux Q_h . During April-May a reversal from net upward to net downward heat flux occurs, then in August-September the change back to a net upward flux. The sensible heat flux Q_h is dominant during November-April, while the radiative flux $Q_i + Q_b$ dominates in May-August.

Possible mechanisms by which heat might be supplied from below, in order to prevent ice formation during the winter (September-April) when the surface heat flow is upward, are convection, diffusion and upwelling. Winter convection in northern Baffin Bay, occurring concurrently with formation of 1.5 to 2 m ice, has been noted (section 2.1.3) to extend downward to about

150 m, the bottom of the cold core. Because temperatures are reduced to the freezing point during this process, insufficient heat is present in the upper 150 m to prevent ice formation. The relatively warm intermediate depth Atlantic Water layer might, however, provide heat sufficient to deter ice formation. While data acquired prior to 1969 suggest that thermohaline convection in the North Water area would be expected to extend only to the bottom of the cold core, and would therefore be ineffectual as an upward heat transfer mechanism to prevent ice formation, the 1969 data revealed a local area of reduced vertical stability south of Cape York; ice formation there would have been sufficient to create downward convection to the Atlantic Water, and heat would have been transported to the surface to inhibit further ice formation. The divergence leading to this region of reduced stability is contingent, however, upon the presence of a long-shore current south of Cape York which has been shown above to be driven by a melt water wedge originating from the western Greenland coast. It seems unlikely that either the fresh water wedge or its consequent region of divergence would persist far into the winter. Winter vertical convection in this region, as effective in transporting heat upward from the Atlantic Water and preventing ice formation, must therefore be considered unlikely.

Vertical diffusion may be ruled out as an effective method of upward heat transfer from the Atlantic Water layer. Assumption of a vertical eddy conductivity of $10 \text{ cal/cm-sec-}^{\circ}\text{C}$ and use of the temperature profiles (Fig. 4a) yields an upward diffusive heat flux from the Atlantic into the Arctic Water of about $1.5 \times 10^{-3} \text{ g-cal/cm}^2\text{-sec}$. This flux is an order of magnitude less than that required to maintain the area ice-free.

The winter geostrophic winds as deduced from the atmospheric isobars in the North Water area (Figs. 31, 33) suggest Ekman offshore upwelling as a

possible mechanism for transport of heat from the Atlantic Water to the surface. Such upwelling is not evident in the late summer, but the winter wind pattern over northern Baffin Bay is not generally fully developed until late September-early October and may not have been detected in August-September. If the North Water were strictly an open water area, the assumptions made in section 3.2 regarding stress transfer through an open water surface may be used and yield a west-southwesterly Ekman transport of about 0.6×10^4 g/cm-sec. Occurrence of this transport across the heavy dotted line (indicated on Figure 33), which is about 100 km long and extends from the southern boundary of the North Water to the northern limit of observed summer occurrence of Atlantic Water, yields a total offshore transport of 0.06×10^6 m³/sec. This transport might be compensated for by upwelling in the Cape York coastal region. Assuming that the upwelling water is Atlantic Water having a temperature of about 1° C, approximately 0.2×10^{12} g-cal/sec (relative to -1.85 °C) would be supplied by the upwelling.

Figure 33 allows an estimate of the open water surface area in the North Water, and therefore an estimate of the total heat (based on Table VIII), required to maintain this open water. The total indicated area is about 4×10^{14} cm². An undetermined amount of this total area is covered by pack ice which will reduce the upward heat flow. Using the thermal conductivity for pack ice given in Neumann and Pierson (1966) and assuming 1 m thick ice, the maximum winter heat flux upward through the ice is 1.5×10^{-3} g-cal/cm²-sec. This is an order of magnitude smaller than the upward heat flux through an open water surface. To the accuracy of the calculations made here, the upward heat flux through ice-covered areas will be significant, relative to that through open water areas, only in cases where relatively high ice concentrations occur. The presence of ice suggests,

moreover, that upward heat flux is insufficient to prevent ice formation and hence negligible for our purposes. Assume that the North Water is 5/10 covered by pack ice. The effective open water area is then $2 \times 10^{14} \text{ cm}^2$, the computed upward February heat flow is $15 \times 10^{-3} \text{ g-cal/cm}^2\text{-sec}$ (Table VIII) and the upward heat flux over the entire open water area is about $3 \times 10^{12} \text{ g-cal/sec}$.

The heat required to prevent ice formation may now be compared with the amount made available by upwelling. If the required heat were derived from water having a temperature of 1° C , then about $0.9 \times 10^6 \text{ m}^3/\text{sec}$ of this water (based on heat content using -1.85° C as a base temperature) would be required to supply heat from below. This quantity is an order of magnitude greater than the $0.06 \times 10^6 \text{ m}^3/\text{sec}$ which could reasonably be supplied by coastal upwelling. It is probable, moreover, that the mean temperature of water supplied from below would be less than 1° C ; a proportionately greater volume of water would then have to be supplied to provide the same amount of heat. An additional factor is the probable presence of pack ice in the region where upwelling was computed. Although little is known of stress transfer through pack ice, it is probable that Ekman transports and their associated upwelling would be markedly decreased by the presence of pack ice. It is concluded that offshore upwelling due to winds is insufficient to maintain the North Water.

It has been assumed, in the above discussion, that sufficient heat was available in the Atlantic Water layer to effectively prevent ice formation in the North Water. Recent oceanographic data permit examination of this assumption. Temperatures in the Atlantic Water were higher during 1968, in the North Water area, than during other years with the possible exception of 1969. Station 36 (Fig. 19) indicated a maximum heat content in the water

column, during 1968, of about 2×10^5 g-cal (relative to -1.85° C) in a 1 cm^2 column. Assuming upward heat flows as given in Table VIII, this would be sufficient to maintain an ice-free surface for 3-4 months but not throughout the winter. This relatively high heat content was observed, moreover, only in a restricted area east of the Carey Islands; the heat content underlying the rest of the North Water area was considerably less and would have been less effective in preventing ice formation. Comparison with other years (Fig. 15) suggests that far less heat was generally contained in the water than during 1968, and therefore that ice formation could have been prevented by heat from this source only for periods shorter than 3-4 months.

Another possible heat source is the continual northward advection of Atlantic Water into the North Water area by the West Greenland Current. The total computed baroclinic transport in the West Greenland Current was seen to vary from zero to $0.3 \times 10^6 \text{ m}^3/\text{sec}$ northward (section 3.1). Nearly 50% of this transport occurred in the upper 50 m; velocities at 500 m, the depth of the warm core, were only 1-2% of the surface velocities. The actual northward baroclinic transport of Atlantic Water was therefore considerably less than the total northward baroclinic transport in the West Greenland Current. Since a significant portion of the West Greenland Current appeared to be driven by fresh water runoff (and ensuing low salinity wedges) from terrestrial sources, the northward baroclinic transport in winter might be less than that in summer due to the winter absence of runoff. A winter tendency of the intensified winds to create a northward barotropic transport would be counteracted to some uncertain extent by the presence of pack ice. Since the volume of water at 1° C necessary to supply heat sufficient to maintain the North Water was computed above to be $0.9 \times 10^6 \text{ m}^3/\text{sec}$, it is apparent that insufficient Atlantic Water is advected northward by the West

Greenland Current to serve this purpose. There is no evidence to suggest a winter increase in northward flow sufficient to supply the necessary heat for maintenance of the North Water.

Bailey (1957) suggested that a surface pool of warm water east of Jones Sound persisted throughout the winter and might be responsible for open water there (section 2.1.3). If such a layer were 50 m thick and had a mean temperature of 5°C it would be effective in preventing ice formation only until mid or late November, using heat fluxes from Table VIII. The layer is generally cooler and thinner, however, and ice formation could not be deterred in this way even until November. It was shown, moreover, that the warm surface water does not remain at the surface in northern Baffin Bay but is advected southward as a subsurface layer. The warm pool is therefore judged unimportant in maintaining the North Water, although its presence might delay ice formation in the area east of Devon Island for several weeks relative to surrounding areas.

In view of these heat budget considerations, it is concluded that mechanical ice removal processes must be important in maintaining the North Water. In particular, northerly winds and southward currents might sweep the ice southward as it is formed. The key to mechanical processes of ice removal appears to lie in the arch-shaped ice dam regularly observed in northern Smith Sound; this dam prevents ice from entering the North Water from the north. The reason for location of the dam is uncertain, since its formation has never been observed. It is probable that remnants of the preceding winter's pack ice in Kane Basin adhere to each other at the onset of the September-October freeze; the resulting larger floes might then become lodged in the constriction represented by the northern opening of Smith Sound. Essentially this same explanation was offered by Simpson (1958),

although he did not substantiate his arguments with heat budgets.

Following formation of the ice dam in Smith Sound, northerly winds would concentrate the loose pack ice in the southern portion of the North Water, as observed. Since the direction of ice drift is approximately 45° to the right of the wind direction in the northern hemisphere, maximum ice concentration would be expected off Jones Sound. The region along the Greenland coast between northern Smith Sound and Thule would contain a relatively low ice concentration. Southerly winds in Melville Bay would tend to concentrate ice there rather than creating open water, contrary to Kiillerich (1933) who did not have access to sufficient atmospheric data to determine the winter wind field. The apparently occasional but poorly documented occurrence of open water in Jones and Lancaster sounds during the winter is also probably due to winds and currents; no warm water occurs in Jones Sound, and the warm water in Lancaster Sound is too deep to contribute heat to the surface.

The reason for the locations of the western, southern and eastern North Water boundaries is unclear. The extent of the shore-fast ice along the western and eastern boundaries would be expected to reflect a balance between the mechanical forces tending to remove the ice and the structural strength, as effective in preventing such removal, of the ice itself. Maintenance of the eastern boundary might be aided by the numerous icebergs there, driven by 10-20 cm/sec tidal currents, breaking up young pack ice as it is formed; the resulting pieces could then be readily swept southwestward by the wind. In order to prevent an accumulation of ice along the southern boundary, continual advection of ice across the boundary must occur. The location of the boundary would reflect a balance between the rate of ice input from the north and the rate at which it is removed southward. Since

little is known about ice movement in northern Baffin Bay during the winter, no estimates can be made of the rates of southward advection. Ice potential computations, which allow an estimate of the rate of ice formation in the North Water, would be affected by the rate at which the newly formed ice is broken up and removed; since nothing is known of the rate of breakup and renewal, nothing can be said concerning the production. The reasons for the location of the southern boundary must remain uncertain pending winter observations.

The spring-summer increase of the North Water area, primarily to the southwest and into Lancaster Sound, may be approximately explained utilizing the computed downward heat flux (Fig. 33). The date on which the surface area begins to increase coincides approximately with the date on which the surface heat flux reverses from net upward to net downward. The net downward flux would be expected to create a thin (on the order of a few meters) surface layer which would be warmer than the underlying water and also less saline due to meltwater admixture. This surface layer would therefore be less dense than the underlying water and would tend to retain its identity. Heat added to the surface layer in the open water through the surface might then be effective, as the surface layer is advected to the southwest by the southward currents, in melting the ice from below. The occasional tendency for extension of open water into Melville Bay, south of Cape York, suggests that surface currents there may occasionally flow eastward. In addition, meltwater puddles formed atop the pack ice would lower its albedo and hasten absorption of solar radiation and consequent melting from above.

The data suggest that the above hypothesis for enlargement of the North Water in spring is feasible. If 5/10 ice cover 1 m thick is assumed on the North Water in early May, the heat influx is of the proper order, although

small, to account for the observed decrease in ice cover. It is impossible to estimate the effect of melt water puddles and consequent absorption of solar radiation on the ice melt; this influence would be, however, in the

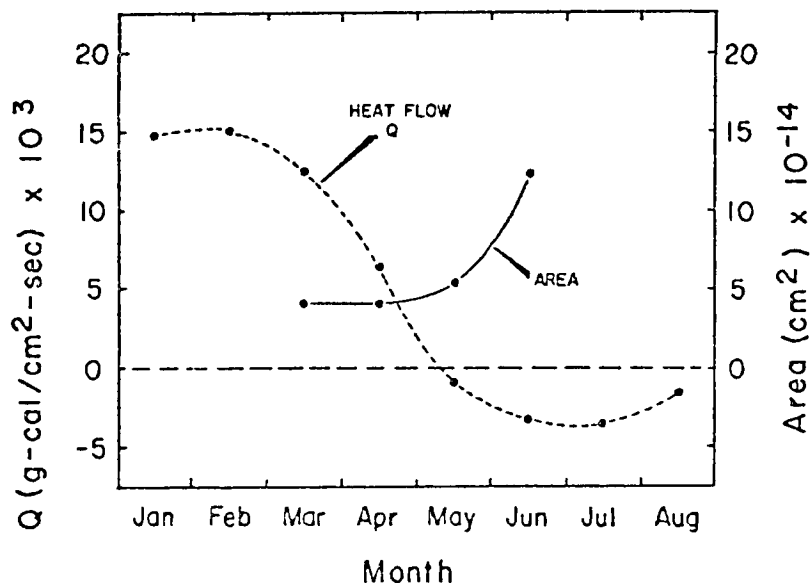


Figure 34. Monthly variations in computed heat flux through an open water surface in the North Water area (from Table VIII) and of the North Water surface area as obtained from Figure 33.

direction to correct the apparent too-large decrease in ice cover. Surface heat influx is more than sufficient to account for the May-June decrease in ice cover. It is concluded that the spring increase in open water area results from net surface heat influx, the pattern of ice disintegration being a reflection of the surface currents.

4.3 Summary

All data suggest that the North Water is maintained primarily by mechanical removal of ice by winds and currents. Transfer of heat to the surface from the Atlantic Water via thermohaline convection is impossible, while diffusion and upwelling are insufficient as mechanisms of upward heat transfer

for prevention of ice formation. Oceanographic data suggest that insufficient heat is present in the area to maintain open water, and that insufficient heat is transported northward into the area by currents.

The observed spring-summer increase of the North Water area may be accounted for approximately by the computed net downward heat flux coupled with advection of warmed water, by currents, to beneath the ice. Absorption of solar radiation by the ice also plays a role.

V. SUMMARY AND CONCLUSIONS

1. The waters of Baffin Bay may be divided, on the basis of their vertical temperature distribution, into three layers. The characteristics of this structure have remained unchanged from 1928 to 1966. The layers are:

a. An upper layer of relatively cold (0°C to -1.8°C), low salinity (<31 ‰ to about 34.4 ‰) Arctic Water. This extends from the surface down to 100-150 m in northeastern Baffin Bay, to 200-300 m in northwestern Baffin Bay and Lancaster Sound, and to the bottom in Smith and Jones sounds. Minimum temperatures in a prominent cold core occur in northeastern and north central Baffin Bay; the core is a consequence of winter thermohaline convection within Baffin Bay. A shallower and less pronounced core in northwestern Baffin Bay, Smith, Jones and Lancaster sounds is a remnant of the winter convective layer in the Arctic Ocean.

Arctic Water is water of Atlantic Ocean origin which either entered Baffin Bay via Davis Strait and was modified by cooling and admixture of runoff within northern Baffin Bay, or was modified by cooling and freshening within the Arctic Ocean prior to entering Baffin Bay from the north.

b. An intermediate layer of warmer (0°C to $>2^{\circ}\text{C}$), more saline (34.2 ‰ to 34.5 ‰) Atlantic Water. This extends from the bottom of the Arctic Water layer down to 1200-1300 m in northern Baffin Bay proper, down to 600-700 m in Lancaster Sound, and is absent from Smith and Jones sounds. Maximum temperatures occur within a

pronounced warm core in this layer

The Atlantic Water originates from the Atlantic Ocean via Davis Strait, from whence it is advected northward by the West Greenland Current while losing heat due to mixing with Arctic Water. Relatively shallow sill depths between Baffin Bay and the Arctic Ocean prevent a southward flow of Arctic Ocean Atlantic Water.

c. A deep layer of relatively cold (0°C to -0.4°C) Deep Water, including water below 1800 m of extremely uniform temperature (-0.40°C) and salinity ($34.48\text{ }^{\circ}/\text{oo}$); the Bottom Water. The Deep Water is characterized by a temperature decrease and nearly uniform salinity as the depth increases to 1800 m. Horizontal temperature and salinity variations, if any, are undetectable.

The Deep Water is identical to water occurring at 200-250 m depth in the Eurasian Basin of the Arctic Ocean. Water of proper type to contribute to Deep Water only above about 1400 m depth occurs commonly in Nares Strait, however. Deep Water below 1400 m, including Bottom Water, is suspected to originate from the Arctic Ocean as irregular pulses; however, these have never been observed. It is improbable that winter thermohaline convection within Baffin Bay contributes to the Deep Water.

2. The southern Smith Sound, Jones and Lancaster sound regions are sites of mixing between inflowing Arctic Ocean Water and Baffin Bay Arctic Water of the same density. The major proportion of the Arctic Ocean Water is less dense than Baffin Bay Water, however, and therefore flows over it. The resulting near-surface layer of Arctic Water of

Arctic Ocean origin occurs primarily in northwestern Baffin Bay, where it is advected southward by the Baffin Current. Its presence produces a prominent downward slope, to the west, of the isohalines.

Mixing between Baffin Bay Atlantic Water and inflowing Arctic Ocean Water occurs primarily in the southern Smith Sound region. Of the three channels entering northern Baffin Bay, only Smith Sound has sufficiently deep access to the Arctic Ocean to allow southward flow of Arctic Ocean Water dense enough to occur at the same depth as and therefore mix with Baffin Bay Atlantic Water. This mixing, coupled with irregular flow through Nares Strait, contributes to an irregular temperature distribution in the southern Smith Sound region.

3. The net computed (baroclinic) transport through northern Baffin Bay is $2.0 \pm 0.7 \times 10^6 \text{ m}^3/\text{sec}$ southward. Approximately twice as much inflow into Baffin Bay occurs through Lancaster Sound as through Smith or Jones sounds. Surface flow reversals have been observed in Smith Sound, while reversals of baroclinic flow have occurred in both Smith and Jones sounds.

It was impossible to obtain a volume transport balance by geostrophic calculations. This was felt to be due primarily to time variations coupled with the non-synopticity of oceanographic data; the neglect of barotropic currents inherent in the dynamic method may also have been a contributing factor.

The net southward transport through northern Baffin Bay appears to be driven by a higher surface elevation in the Arctic Ocean than in Baffin Bay; this difference is due to differences in water structure between the upper 250 m layers of the Arctic Ocean and Baffin

Bay. Atmospheric pressure disturbances may be sufficient to cause flow reversals in Nares Strait.

4. The baroclinic flow pattern within northern Baffin Bay consists of the northward West Greenland Current in its eastern part; the southward Baffin Current in its western part; and a cyclonic gyre east of Devon Island. Cyclonic circulation of Baffin Bay Water occurs in southern Smith Sound and in eastern Jones and Lancaster sounds. The West Greenland Current transports northward only about 10% of the volume transported southward by the Baffin Current, and is not always present during the late summer.

The cyclonic circulation within northern Baffin Bay is driven by a combination of winds and baroclinic fields resulting from low-salinity near-shore wedges due to meltwater admixture. Both the baroclinic field due to the presence of less dense Arctic Ocean Water and westward intensification processes appear to be important in the dynamics of the Baffin Current.

Bottom topography exerts some control on the currents in Lancaster Sound and off Cape York. This control, coupled with the temperature distribution in the Atlantic Water, suggests that the currents extend nearly to the bottom there. Year-to-year variations, including formation of eddies and gyres, may also occur in Lancaster Sound and off Cape York.

5. Seasonal variations in the water structure include winter development of a cold, uniform near-surface layer due to freezing which becomes isolated during the summer by a warmed surface layer, leaving a cold core beneath. Summer meltwater addition in Nares Strait and west of

Jones and Lancaster sounds causes freshening, relative to winter salinities, down to deeper than 100 m.

Seasonal circulation variations are uncertain. An approximate doubling of wind stress, with quadrupling of the wind stress curl, occurs during the winter; lack of knowledge of vertical stress transfer through pack ice precludes, however, a quantitative estimate of the winter wind-driven currents. The absence, during the winter, of near-shore low salinity wedges due to meltwater admixture might result in diminution or absence of the West Greenland Current. No appreciable changes would be expected in the net southward flow through the system, because the same processes affecting density in the upper 250 m occur both in Baffin Bay and in the Arctic Ocean.

6. The North Water is a large, semi-permanent polynya which occurs regularly throughout the winter in the northern Baffin Bay-Smith Sound region. The extent of this open water has been roughly documented only for the March-June period; after June it loses its identity due to the eastern Arctic melt. Year-to-year variations in the open water area are as large as monthly variations during a given year, which contribute to difficulties in determining its exact extent.

Computations using summer oceanographic data suggest that insufficient heat is available in the Atlantic Water layer in northern Baffin Bay to prevent ice formation in the North Water. A probable winter decrease in the transport of the West Greenland Current would further decrease the available heat. Mechanical processes of ice removal must therefore be primarily responsible for the formation and maintenance of the North Water.

While overall summer conditions have remained constant, the northern Baffin Bay region is characterized by a high degree of time variability in its circulation and hydrographic structure details. It is therefore possible at this time to depict these features on only a gross scale.

Further summer oceanographic exploration in northern Baffin Bay will yield little new information. Combined winter oceanographic and meteorological sampling would, however, increase overall understanding of the circulation and hydrography. A winter oceanographic sampling program might include a fall general survey, a detailed winter hydrographic sampling program and a second general survey following the winter program. The general surveys would determine whether the year was an "average" one in terms of hydrography and circulation. A comparison between the first and second general surveys would allow detection of any net changes which may have occurred during the winter. The winter survey would concentrate on:

- a. Extent of the Atlantic Water;
- b. Details of vertical convective processes;
- c. Ice conditions;
- d. Surface water temperature and salinity;
- e. Formation of the boundaries of the North Water; and
- f. Small scale meteorological observation concurrent with oceanographic observations.

The exact winter sampling pattern would be determined by the results of the first general survey and by the data as it is obtained during the winter survey itself. Flexibility would therefore be maintained and allow surveying of any pertinent features.

Recording current meters would operate in Smith and Lancaster sounds throughout the winter. These would allow transport estimates and detection of time variations as reflected in the currents.

Implementation of the above program would allow testing of the hypotheses concerning seasonal changes and formation of the North Water. Oceanographic knowledge of the Arctic regions in general, interconnected by the channels through the Canadian Islands, would also be increased.

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APPENDIX A
THE OCEANOGRAPHIC DATA

TABLE A-1

Information on oceanographic data from the northern Baffin Bay region and the Lincoln Sea

| <u>Year</u> | <u>Platform</u> | <u>Period</u> | <u>Source</u> | <u>CRN</u> ¹ |
|-------------|-------------------|-------------------|---------------------------------|-------------------------|
| 1916 | () ² | July-November | Knudsen 1923 | |
| 1928 | GODTHAAB | July-August | Riis-Carstensen 1931 | |
| 1940 | NORTHLAND | September | Barnes 1941 | |
| 1954 | LABRADOR | July-August | CODC ³ | 189 |
| 1956 | LABRADOR | September-October | CODC | 219 |
| 1957 | LABRADOR | September | CODC | 244 |
| 1959 | WESTWIND | September | NODC ⁴ | 638 |
| 1960 | WESTWIND | August | NODC | 670 |
| 1960 | LABRADOR | September | GODC | 340 |
| 1961 | EDISTO | September | NODC | 688 |
| 1961 | LABRADOR | September | CODC | 341 |
| 1962 | ATKA | September | NODC | 966 |
| 1962 | LABRADOR | October | CODC | 362 |
| 1963 | EVERGREEN | July-August | Franceschetti <u>et al</u> 1964 | 174 |
| 1963 | LABRADOR | September-October | CODC | 005 |
| 1964 | LABRADOR | September-October | CODC | 020 |
| 1965 | LABRADOR | September-October | CODC | 001 |
| 1966 | EDISTO | July-August | Palfrey and Day 1968 | 807 |
| 1967 | () ⁵ | June | Seibert 1968 | |
| 1968 | WESTWIND | September | Muench in press | |
| 1969 | KB-2 ⁶ | May | NODC | |
| 1969 | WESTWIND | September-October | NODC | |

Locations of the above data are indicated on Figures A-1 through A-4.

- (1) CRN designates cruise reference number used by listed source.
- (2) Platform unspecified in reference.
- (3) Canadian Oceanographic Data Centre.
- (4) National Oceanographic Data Center.
- (5) Data were acquired using pack ice as a platform.
- (6) KB-2 was the ice station from which data were acquired.

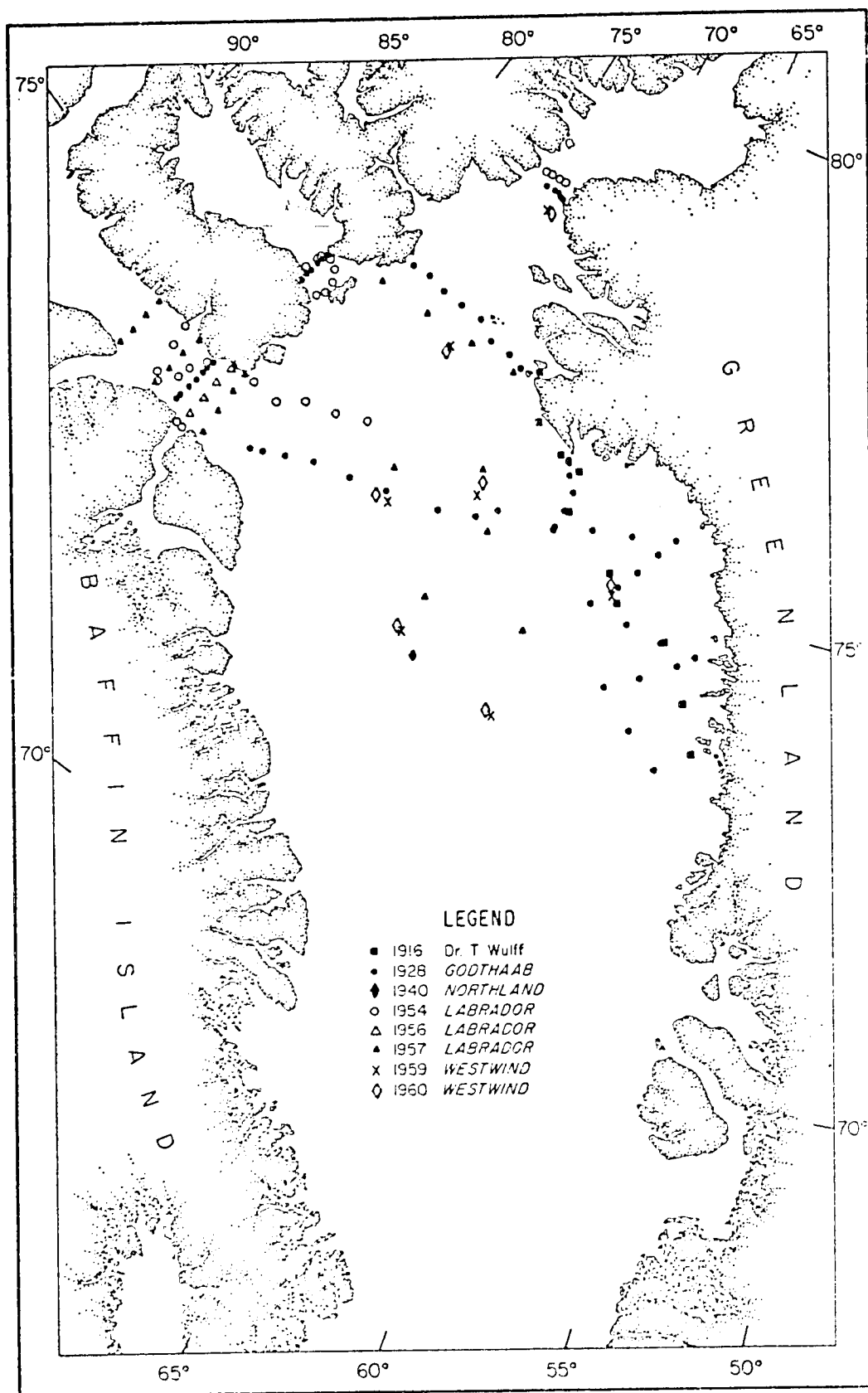


Figure A-1. Locations of oceanographic stations occupied from 1916 to 1960.

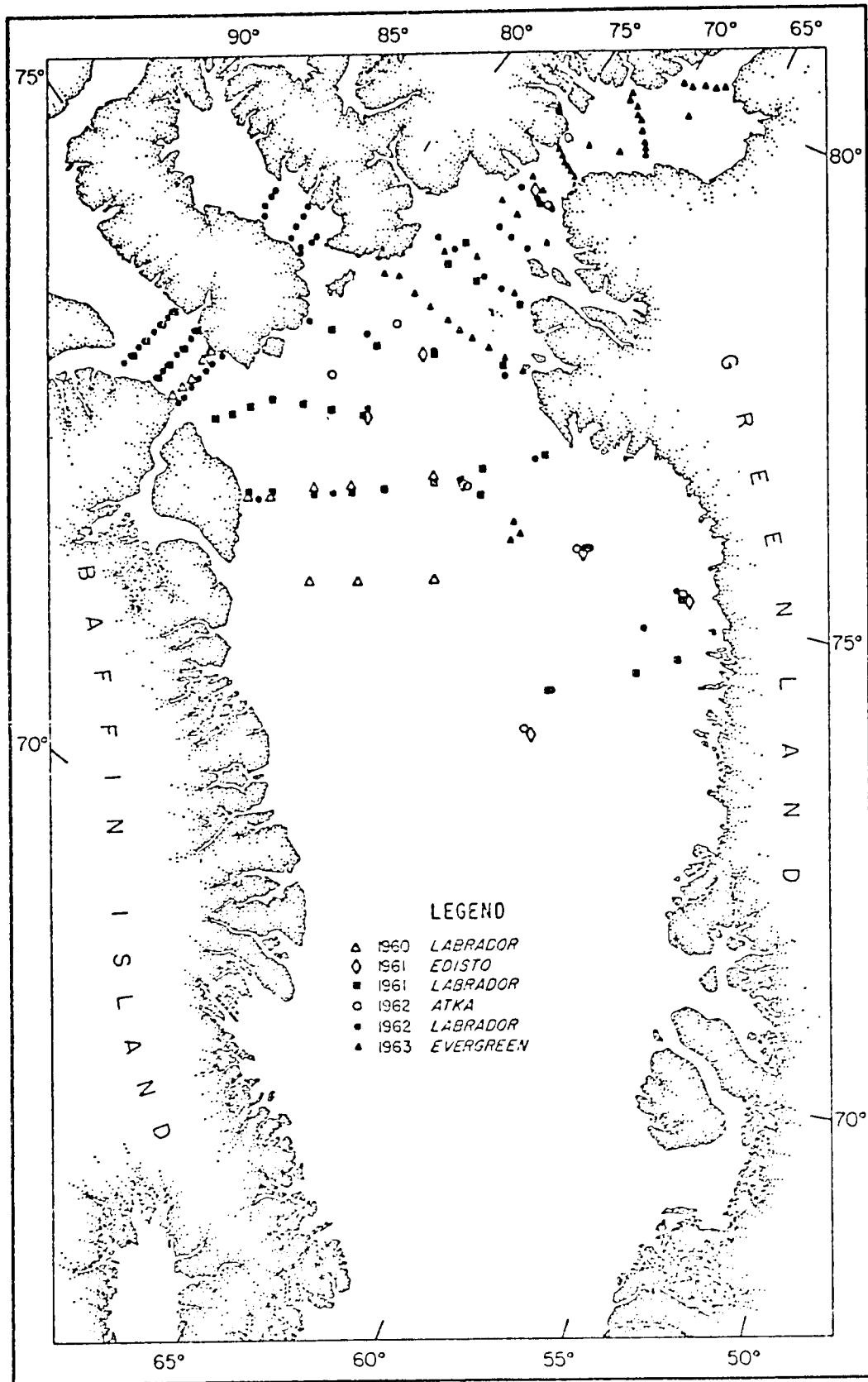


Figure A-2. Locations of oceanographic stations occupied from 1960 to 1963.

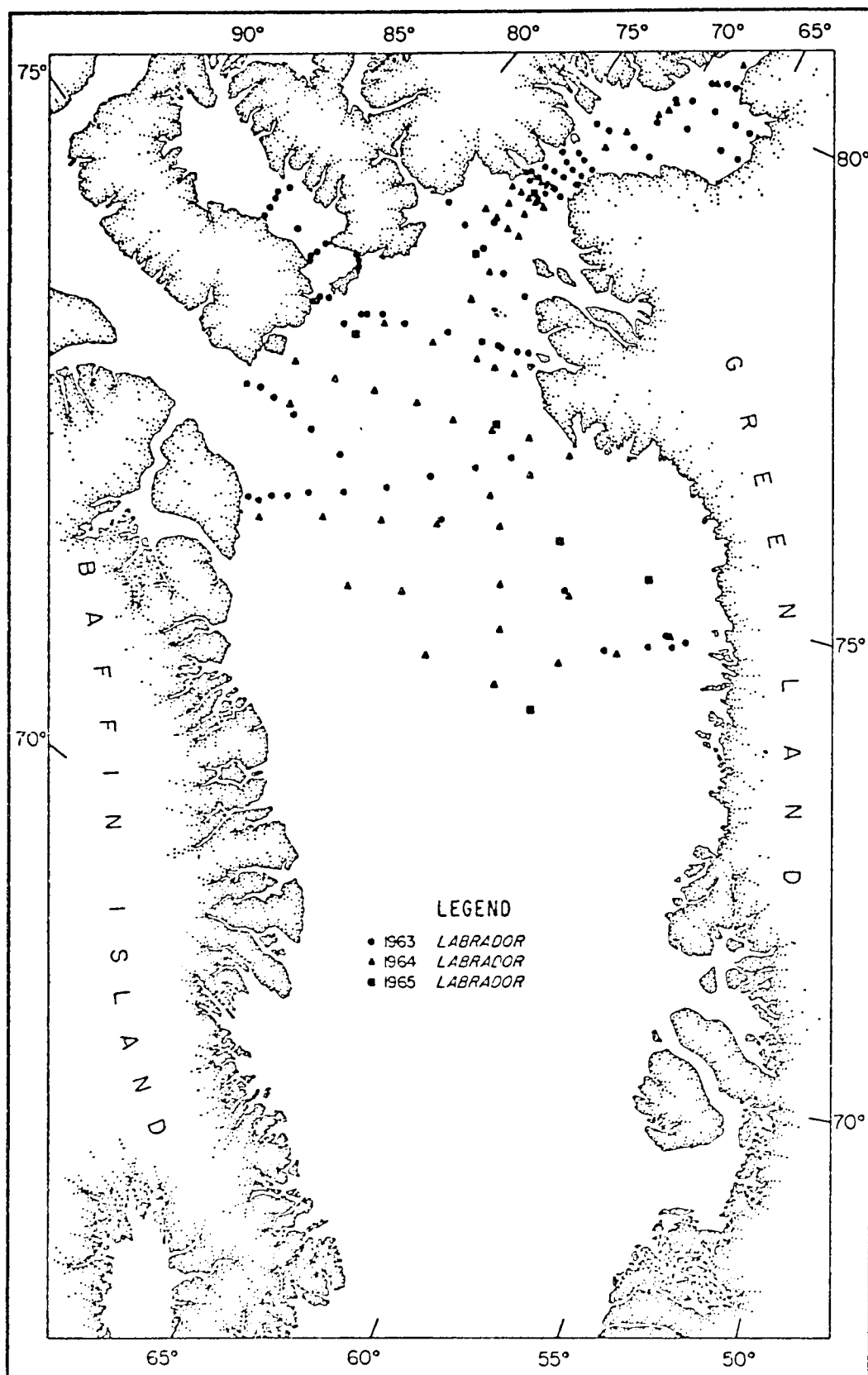


Figure A-3. Locations of oceanographic stations occupied from 1963 to 1965.

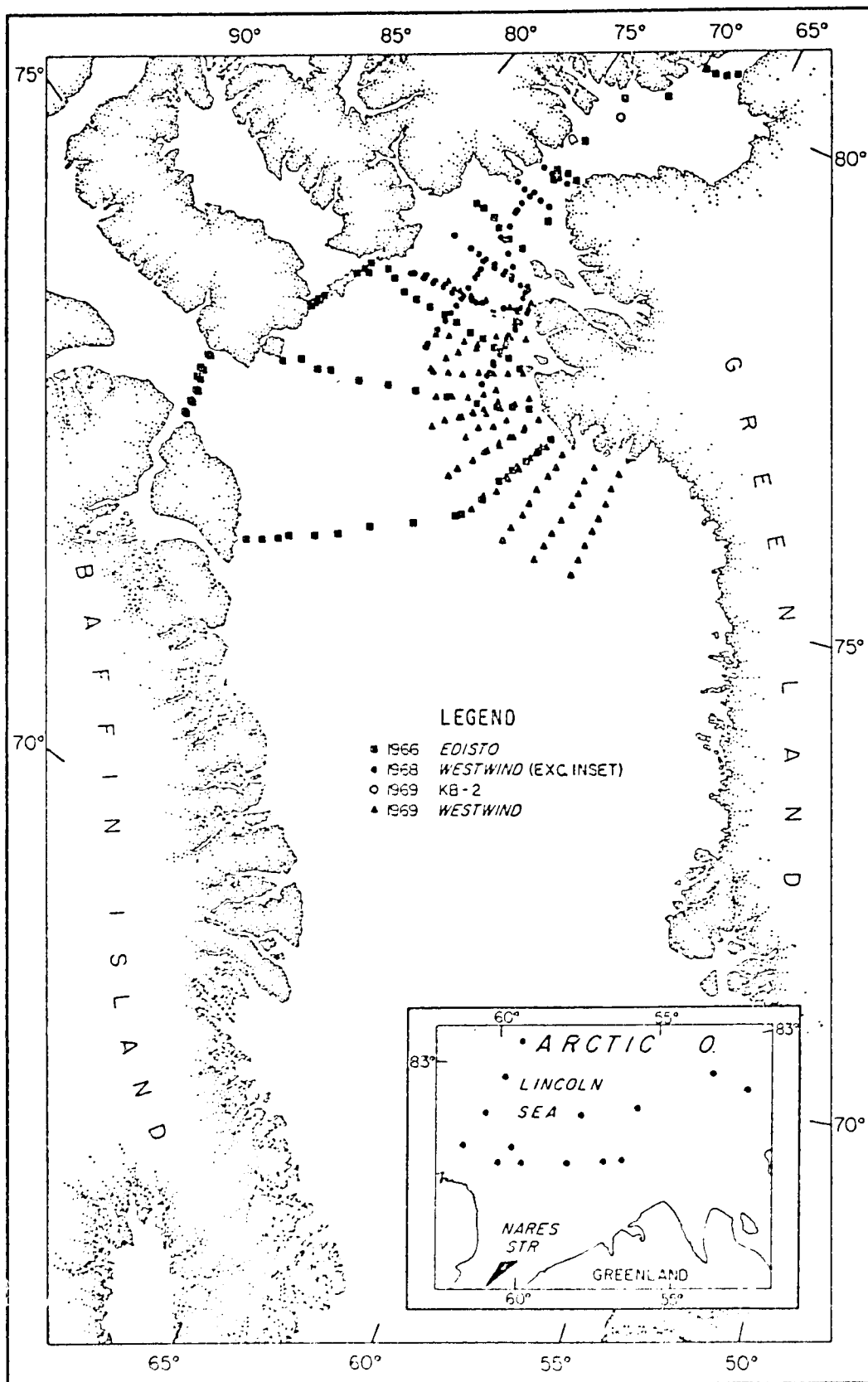


Figure A-4. Locations of oceanographic stations occupied from 1966 to 1969 and (inset) stations occupied in the Lincoln Sea in 1967.

TABLE A-2

Information on oceanographic data from outside the
northern Baffin Bay region

| <u>Year</u> | <u>Platform</u> | <u>Period</u> | <u>Location*</u> | <u>Source</u> | <u>CRN</u> |
|-------------|-----------------|------------------|------------------|----------------------------|------------|
| 1961 | LABRADOR | August-September | M'Clure Strait | CODC | 341 |
| 1962 | LABRADOR | October | Davis Strait | CODC | 362 |
| 1964 | LABRADOR | October | Davis Strait | CODC | 020 |
| 1964 | ARLIS-II | February-Sept. | Eurasian Basin | Tripp and Kusunoki 1967 | AR-2 |
| 1965 | LABRADOR | October | Davis Strait | CODC | 001 |
| 1965 | T-3 | July | Canadian Basin | Tripp 1966 | WO-1 |
| 1965 | T-3 | Oct., Dec. | Canadian Basin | Tripp 1966 | WO-2 |
| 1966 | T-3 | April | Canadian Basin | NODC | WO-3 |
| 1966 | T-3 | June, Sept. | Canadian Basin | NODC | WO-4 |
| 1967 | T-3 | February | Canadian Basin | NODC | WO-6 |

*Approximate locations of the data used from these regions are
indicated on Figure 7 of the text.

TABLE A-3

Information on current data from northern Baffin
Bay

| <u>Year</u> | <u>Platform</u> | <u>Period</u> | <u>Source</u> | <u>Record Length (days)</u> |
|-------------|-----------------|---------------|----------------------|-----------------------------|
| 1963 | EVERGREEN | July | Palfrey and Day 1968 | 3 |
| 1968 | WENDWIND | September | Univ. of Wash. | 16 |
| 1969 | KB-2 | May | Univ. of Wash. | 16 |

Locations of current measurements are shown on Figure A-5.

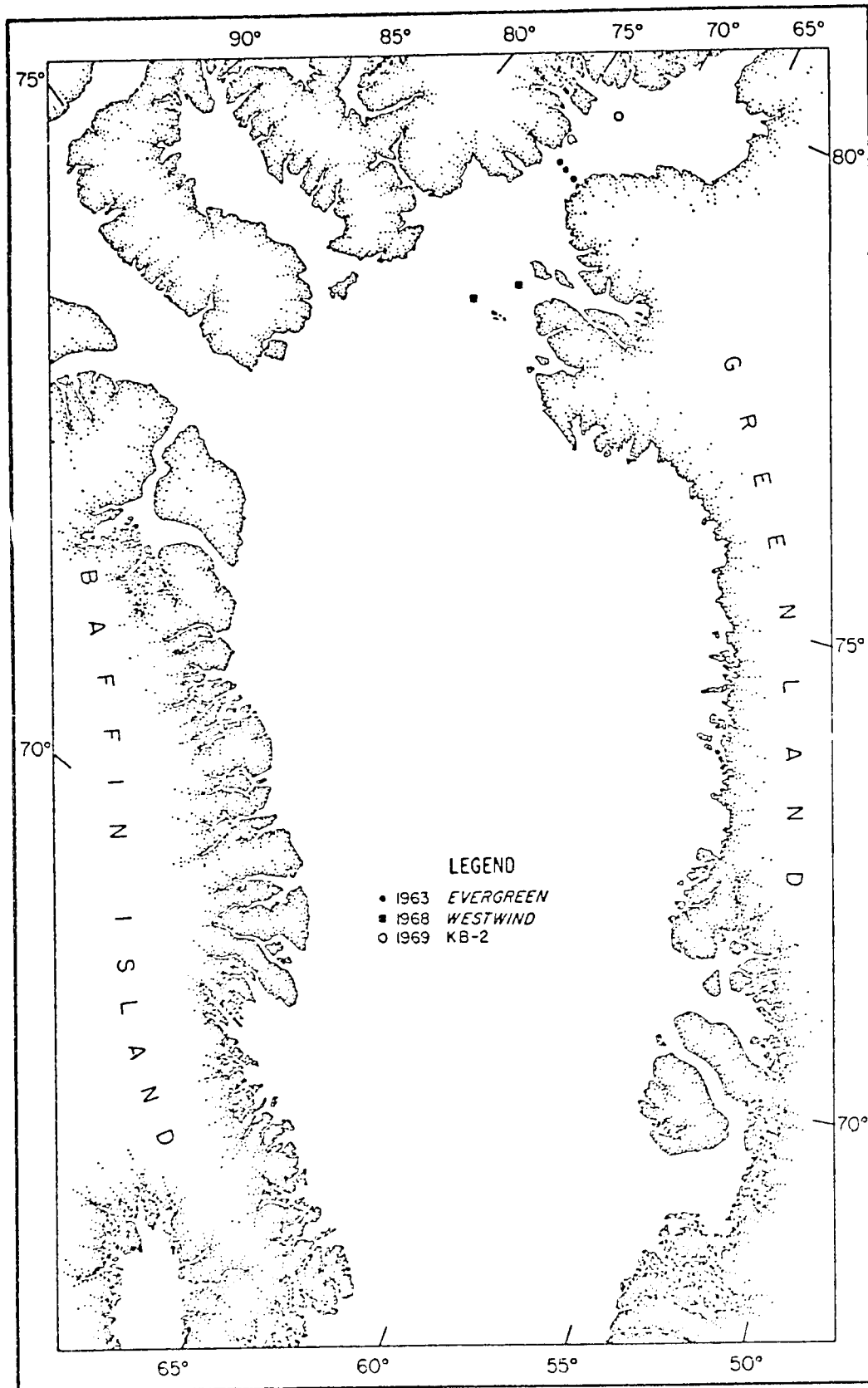


Figure A-5. Locations of current measurements.

APPENDIX B

Computation of monthly mean heat fluxes through an open water surface
in the North Water region

The heat budget of northern Baffin Bay is of prime interest in considering feasible mechanisms for preventing ice formation, and thus maintaining the North Water. This heat budget may be expressed by the equation

$$Q_h + Q_e + Q_b + Q_i + Q_\theta + Q_a = 0$$

where:

Q_h = sensible heat exchange through the surface;

Q_e = latent heat exchange through the surface;

Q_b = net back radiation from the surface;

Q_i = net incoming short-wave radiation to the surface;

Q_θ = heat absorbed to change water temperature; and

Q_a = heat advected in or out from below by water currents.

Of these, Q_θ will be assumed negligible, i.e., a steady-state will be assumed. Q_a has been estimated, in the text, from summer oceanographic observations.

To compute the remaining terms, it was necessary to consider the atmospheric data. Monthly mean air temperatures and vapor pressures, at ground level, and cloud cover were available from the four stations indicated on Figure 34 (National Weather Records Center, Asheville). Mean atmospheric lapse rates during winter (December-March) and summer (June-August) were available from Thule (Putnins et al. 1960). Intermediate values were assumed for the intervening periods.

Modification of the near-surface atmospheric layer due to an upward

flux of heat and moisture would be expected to occur during the winter months when air temperature and vapor pressure are lower than those of the water. A model describing such modification has been developed by Businger (1954). The winter wind pattern suggests that monthly mean temperatures and humidities at Craig Harbour and Thule are representative of upstream boundary conditions for an air modification model. Lower rather than higher temperatures at Dundas Harbour and Pond Inlet suggest that the winds do, in fact, parallel the Baffin Island coast, as indicated on Figure 3 (in text), or may even have an easterly component.

The winter lapse rates at Thule were essentially zero, suggesting vertical mixing of the air masses during their descent over rough topography from the ice cap. Similarly rough topography upwind from Craig Harbour suggests that the lapse rate there was similar, therefore the Thule lapse rate was assumed representative of upstream conditions.

The roughness parameter over open water, also necessary for computation of the air mass modification, was chosen as 0.01 cm based on Roll (1965).

The northern North Water boundary was assumed to be the upstream boundary. Modified air temperatures and vapor pressures were computed, using Businger's (1954) model, at 1 m elevation and 50 km intervals downstream to a line along the latitude of Thule. These values were averaged to obtain means (Table B-1), for the entire area, which were used in the below computations. Modified mean temperatures were generally about 20% higher, during the winter, than the unmodified values, while the modified vapor pressures were 2-3 times larger than the unmodified values.

Due to the approximate nature of the data, the methods presented by Sverdrup, Johnson and Fleming (1942) were considered adequate for computation of Q_e and Q_h . Q_e was computed using the equation

$$Q_e \text{ (g-cal/24 hrs)} = 0.0143 (e_w - e_a) W_a L,$$

where:

e_w = vapor pressure of the water (mb);

e_a = vapor pressure of the air (mb);

W_a = mean wind speed (m/sec); and

L = latent heat of vaporization = 576 cal/g.

Walmsley's (1966) value of W_a for northern Baffin Bay, computed on the basis of daily atmospheric pressure distributions, was used rather than the speed based on the seasonal mean geostrophic winds. This was necessary due to high wind variability in the region.

Sensible heat exchange Q_h was computed using the Bowen ratio, which yields the equation

$$Q_h \text{ (g-cal/24 hrs)} = 0.65 \frac{T_w - T_a}{e_w - e_a} Q_e,$$

where:

T_w = water temperature at surface ($^{\circ}$ C); and

T_a = air temperature at surface ($^{\circ}$ C).

Table B-1 indicates the values of the above parameters used for each month, while Table VIII (in text) indicates resulting values of Q_e and Q_h .

The incoming solar radiation Q_i absorbed by the water may be computed using the equation

$$Q_i \text{ (g-cal/cm}^2\text{-min)} = 0.027 (1 - 0.071 \bar{C}) \bar{h} \left(1 - \frac{r}{100}\right),$$

where:

\bar{C} = mean cloud cover (tenths);

\bar{h} = mean solar altitude (degrees); and

r = reflectivity of the water surface (%)

The values of \bar{C} used were obtained by averaging monthly mean values from the four stations surrounding the North Water (Fig. 3, in text) and are given in Table B-1. Values of r were obtained from Table 26 (Sverdrup et al. 1942), and \bar{h} was determined using the Nautical Almanac.

The net back radiation Q_b was computed from the equation

$$Q_b \text{ (g-cal/cm}^2\text{-min)} = 0.94 (Q_{beff} (1 - 0.083 \bar{C})),$$

where Q_{beff} = the effective back radiation to a clear sky from open water; a function of sea surface temperature and relative humidity, it was determined from Figure 25 in Sverdrup, Johnson and Fleming (1942). The resulting values of net radiative transfer $Q_i + Q_b$ are given in Table VIII (text).

The foregoing computations were admittedly approximate. T_a and e_a should have been measured at a specified distance above the water surface rather than at varying elevations at land stations, as was the case. The values of \bar{C} tell nothing of the cloud character, so Q_i and Q_b may be in error by some undetermined amount. Surface water temperatures were assumed to be at the freezing point during the winter, but could only be estimated for May-July. Water vapor pressures for May-July, which depend on the temperatures, are also uncertain. Because of the relatively small air-water temperature differences during May-July, erroneous surface water temperatures would lead to relatively large errors in Q_h and Q_e ; Q_h and Q_e are therefore less reliable for May-July than for other months.

TABLE B-1

Parameters used in computing surface heat flux in the
North Water region

| Month | T_w ($^{\circ}\text{C}$) | T_a ($^{\circ}\text{C}$) | e_w (mb) | e_a (mb) | W_a (m/sec) | \bar{C} (tenths) |
|-----------|------------------------------|------------------------------|------------|------------|---------------|--------------------|
| January | -2 | -22 | 5.25 | 0.95 | 7.6 | 4.3 |
| February | -2 | -24 | 5.25 | 0.91 | 7.4 | 4.3 |
| March | -2 | -21 | 5.25 | 0.94 | 7.0 | 3.9 |
| April | -2 | -15 | 5.25 | 1.48 | 6.5 | 4.5 |
| May | -1 | -5 | 5.57 | 3.47 | 4.7 | 6.4 |
| June | 2 | 2 | 6.92 | 5.56 | 3.6 | 6.2 |
| July | 3 | 6 | 7.43 | 6.91 | 4.2 | 6.8 |
| August | 3 | 4 | 7.43 | 6.61 | 3.7 | 6.5 |
| September | 0 | -2 | 5.99 | 3.69 | 5.5 | 7.2 |
| October | -2 | -7 | 5.25 | 2.62 | 7.0 | 6.8 |
| November | -2 | -14 | 5.25 | 1.47 | 7.1 | 6.0 |
| December | -2 | -21 | 5.25 | 1.10 | 7.6 | 4.3 |