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I. The Waters of the ICNAF Convention Area

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Introduction

At the request of the Hydrographic Subcommittee of the International Commission for the Northwest Atlantic Fisheries which met in Copenhagen on October 8, 1953, the authors provide in this report a short general review of certain features of the waters of the Western North Atlantic that make up the ICNAF Convention Area. As a matter of convenience the waters are discussed by sectors under each main heading. Hydrographic investigations are carried out by most of the signatory nations. Investigations which are repeated more regularly year by year are shown in Figure 1 which outlines the Convention Area.

General Submarine Physiography

The northwestern-north Atlantic as is discussed here has been described in general terms in the report of the Marion and General Greene Expedition to Davis Strait and the Labrador Sea (Smith, Soule and Mosby, 1937). It is described as that portion of the western Atlantic Ocean embraced by the normal drift of Arctic ice, and thus embraces the waters around and on the Grand Banks and northward, between North America and Greenland to the 70°N. parallel of latitude.

The bathymetric features of this area are shown in Figure 2. From 50°N. to 63°N. the bottom rises gradually to form, between Greenland and Labrador, the Labrador Basin, with depths extending to greater than 2200 fathoms. In the region of Davis Strait, there is an area with depths less than 400 fathoms, a ridge which separates the Labrador Basin from the Baffin Bay Basin where depths are greater than 1000 fathoms. The Labrador side of the Labrador Sea with its well defined continental shelf is in contrast with the narrow continental shelf on the Greenland side. As may be seen in Figure 2 the basins are readily outlined by the 600 fathoms isobath while the continental shelves are less than 200 fathoms. The intervening areas between the two contours, the continental slopes, may be steep, as along the west coast of Greenland, or broad as along the Labrador coast, particularly in its southern limits. To the north, however, the Greenland shelf broadens, and the 200 fathom *contour is found to extend 80 miles (148 km.)* off the coast. This broader shelf is divided into three principal shoals, Fylla, Little Hellefiske and Great Hellefiske Banks. Off northern Labrador, the shelf has a width of 70 miles (130 km.) which broadens to a width of 180 miles (333 km.) off Newfoundland, and in the vicinity of the Grand Banks becomes one of the broadest of continental shelves. The northeasterly extension of the 600

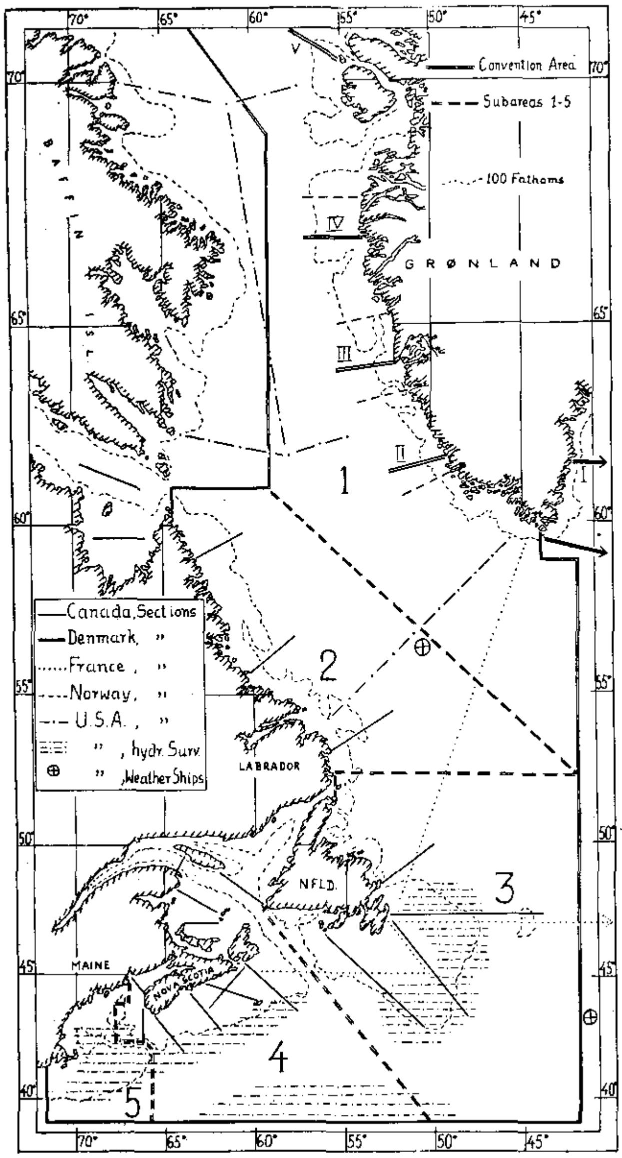


Fig. 1. Routine hydrographic work in the Convention Area. Most of the sections shown are repeated, at least, yearly. Of the sections passing the right frame of the figure the northernmost one (I) reaches Iceland, that from Cape Farvel and that from Newfoundland reach the English Channel.

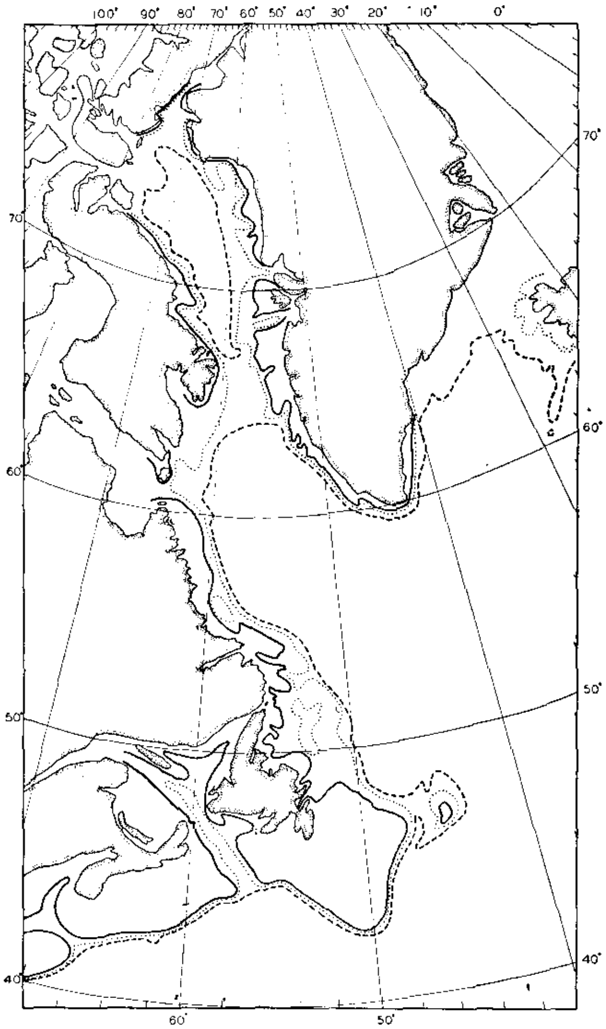


Fig. 2. Bathymetric map of Northwestern North Atlantic showing the 100 fathom ———, the 200 fathom ...., and 600 fathom - - - - - contours. (Simplified after Smith, Soule and Mosby, 1937).

fathoms isobath between the Greenland slope and Reykjanes Ridge creates an eastern appendage and a heart-shape form to the Labrador Basin. The waters in this northeasterly extension of the Labrador Basin are generally referred to as the Irminger Sea. Although this sector is not included in the area under discussion, the waters from the Irminger Current and the East Greenland Sea, which skirts it, contribute in a large measure to the waters in the Labrador Sea. West of the Grand Banks, the continental shelf is broad, forming the Scotian Shelf and Georges Bank. Features of the submarine physiography of the American Continent are the three great channels which cut deeply into the continental shelf and into the continent itself. These are, from north to south respectively, Hudson Strait, the Laurentian Channel and the Fundian Channel. These channels have a very significant influence on the oceanography of the shelf waters in that they bring waters of deep oceanic origin close to the shores of the continent. A more detailed description of the continental shelf from Labrador to Cape Cod is given by H. B. Hachey in the following chapter.

The Continental Shelf from Labrador to Cape Cod

It is a general fact that soundings from the shore of the continents towards the sea show that the depth increases slowly to a certain figure of about 100 fathoms, after which the increase in depth is more rapid. The location of this more rapid increase in depth is defined by oceanographers as the "edge" of the continental shelf, leading to the specific definition of the "continental shelf" as that part of the sea bottom between the shore and this "edge".

The edge of the continental shelf from Labrador to Cape Cod has been traditionally outlined by the location of the 100 fathom isobath. The details of this continental shelf, thus outlined, may be closely followed by plotting the isobaths of 25, 50, 75 and 100 fathoms, the results of which are illustrated in Figure 3.

It should be appreciated that the delimitation of the continental shelf is not as simple as indicated above, and particularly where matters of international law or rights are concerned. The general subject has been dealt with by Sverdrup et al. (1942), Shepard (1948), and Moulton (1952). In particular, reference is made to a detailed map by Veatch and Smith (Moulton, p. 13, 1952), which extends 530 miles along the east coast of the United States. As indicated by this map, the more rapid increase in depth, which might be used to locate the edge of the continental shelf, is near the 60-80 fathom depth. According to Shepard (1948) on a world-wide basis, the average depth at which the greatest change in slope occurs is 72 fathoms.

The submarine physiography of the continental shelf under consideration is illustrated in Figure 3. To the southeast of Newfoundland,

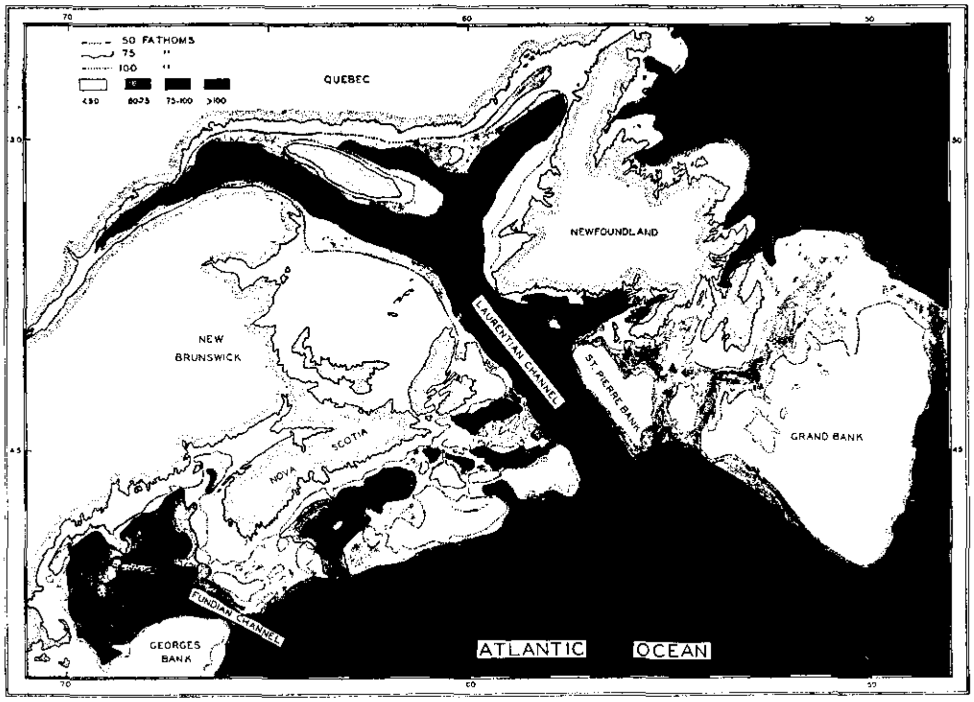


Fig. 3. Submarine physiography of the continental shelf from Labrador to Cape Cod.

the outer shelf consists of a mass of irregular banks which constitute the well known Grand Banks, chief of which are Grand Bank and St. Pierre Bank. Although the depths on these banks average about 30 fathoms, there are many places nearer the land with depths greater than 100 fathoms.

Separating the Grand Banks from the Scotian Shelf is the submerged Laurentian Channel with depths from 100 to 300 fathoms, and extending from the St. Lawrence river outwards into the Gulf of St. Lawrence and thence completely across the continental shelf. The Esquiman and Mingan Channels within the Gulf are branches of this Laurentian Channel extending respectively, northeastward toward Belle Isle Strait and northwestward above Anticosti Island.

The Scotian Shelf is an irregular shaped submarine plateau of irregular topography extending outwards from the coast to a distance of 100 to 150 miles. The more important elevations on this shelf are Sable Island, and Middle, Western, Roseway, LaHave, Emerald, and Sambro Banks, as well as the banks known as Banquereau, Canso, and Misaine. With the exception of several basins of limited extent whose depths are greater, the Scotian Shelf as a whole, is less than 100 fathoms below the sea surface. A large western area of the Scotian Shelf is of depths greater than 50 fathoms and less than 100 fathoms. Bounded as it is, on the north by the mainland of Nova Scotia, on the east by Canso Bank, Middle Ground, and Sable Island Bank, and on the west by Brown's Bank, and with its greater depths extending to the edge of the continental shelf to the southward, this submarine area has been termed the Scotian Gulf. Roseway, LaHave, Sambro, and Emerald Banks form elevations over this portion of the sea floor.

The Fundian Channel with depths greater than 100 fathoms separates the Scotian Shelf from Georges Bank and provides a comparatively deep submerged channel into the Gulf of Maine. Georges Bank is about 140 miles in length by 80 miles in width when based on the location of the 50 fathoms contour, and hence occupies an area of approximately 11,000 square miles. In the Gulf of Maine area the same type of bottom topography is found as farther north (Murray, 1947). On the inside is the Gulf of Maine with its troughs, basins and rises which resemble the Gulf of St. Lawrence The complexity of the relief of the floor of the Gulf of Maine has been illustrated by Murray (1947), basing his contours on the soundings of the U.S. Coast and Geodetic Survey. On the outer shelf, Georges Bank with its shoals is comparable with the Grand Banks, but not as wide. The shelf beyond the shoals of Georges Bank is comparatively smooth except for its irregular margin.

According to Shepard (p. 107, 1948) in this zone of irregular topography extending from Newfoundland to the Gulf of Maine, the bottom character follows a definite pattern. On the banks, sand bottom predominates, although gravel and stones are found in many localities. The inner deeps and the troughs are mud-covered but samples reveal a considerable number of stones mixed with mud. The inner zones along the shore have a rock or boulder bottom, and ridges rising above the inner deeps are also reported as being rocky.

Geologists tell us that in Preglacial Times eastern Canada extended to the edge of the continental shelf, 140 miles beyond the present southeastern coast of Nova Scotia, and Newfoundland was a part of the mainland. The old St. Lawrence River channel can be followed by soundings to the edge of the enlarged continent, where the shallow water ends, and the bottom descends towards the depths of the sea (Coleman, 1922). The fishing banks, extending from Newfoundland to Cape Cod are said to represent "a submerged upland of the Atlantic coastal plain", and the Gulf of Maine is the "drowned inner lowland" between the Banks and the oldlands of New England (Johnson, 1925). According to Johnson too, the "broad and shallow submerged platform bordering the Gaspé Peninsula and the shores of the St. Lawrence embayment—— appear to be normal subaerial features formed above sea level, then submerged and very slightly modified by marine agencies".

A submarine escarpment, sometimes divided into two or more branches and bordering one of the major fault fractures of North America, is discovered under the waters of the Gulf of Maine and traced to its connection with topographic features bordering the northwestern side of the Bay of Fundy. Students of continental faulting have been attempting to trace the faults of Cape Breton across the Cabot Strait to Newfoundland, and have suggested that the Laurentian Channel is in part of the nature of a "fault graben" (Gregory, 1929; Keith, 1930; Hodgson, 1930). Shepard (1931) made a complete study of the various charts of the Laurentian Channel, combined with a special investigation of Cabot Strait. From this study he concluded that the St. Lawrence trough probably was started by river erosion, possibly in part along fault lines. Later during the glacial period, tongues of ice moved down the valley causing great deepening and widening. The present form of the trough is believed to be due principally to this glaciation making it a submarine glacial trough.

The submarine physiography of the continental shelf from Labrador to Cape Cod is pertinent, even to considerable detail, to the study of oceanographical conditions and particularly in relation to fisheries. This continental shelf is in the area of confluence of the Labrador Current and the Gulf Stream where large scale mixing processes occur which provide a type of "slope water" that contributes largely to the characteristic waters that are in contact with the continental shelf and which penetrate toward the coast at the greater depths. The variations in the features of the sea floor determine the distribution of these "slope waters" and provide contrasting bottom water conditions within contiguous areas.

General Circulation

The circulation of waters of the upper layers in the northwest-north Atlantic as shown in Figure 4, consists of a northward flow along the Greenland slope, the West Greenland Current, a southward movement along the Canadian side, the Baffin Land and Labrador Currents, and, a northward set, the Atlantic Current in the southern part of the Labrador Sea. The waters over the Labrador Basin undergo a slow cyclonic movement. The West Greenland Current is composed of the East Greenland and the Irminger Currents which become re-energized on rounding Cape Farewell. The Labrador Current is composed of the Baffin Land Current and the West Greenland Current, and it too is re-energized on passing the mouth of Hudson Strait as waters from Hudson Bay and Foxe Channel are added to the current.

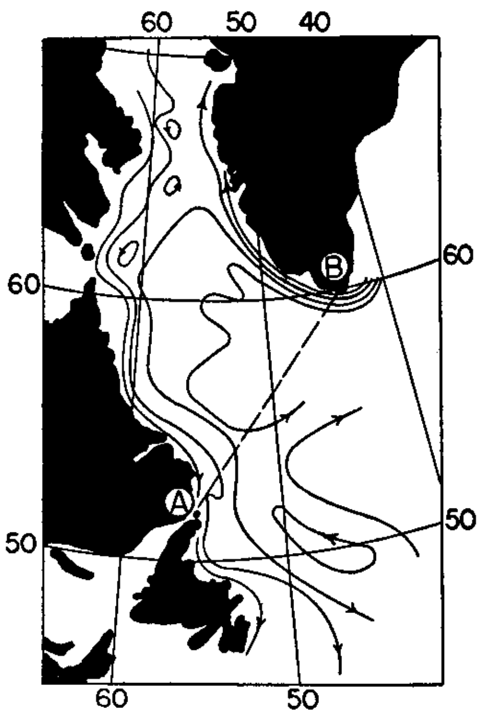


Fig. 4. The system of circulation of the upper water layers in the northwestern North Atlantic (after Smith, Soule and Mosby, 1937).

West Greenland Current.

In the waters of West Greenland, currents of widely different origin and characteristics meet and mix, and the resultant oceanographic conditions are dependant upon which of the currents is dominant at any given time.

The main features of the current system around Greenland are shown in Figure 5. The East Greenland Current, which forms the main outlet from the Polar Sea is found near the East Greenland coast. Off south-east Greenland, this current is mainly restricted to the shelf area, while off the shelf and in part underneath the polar current, the warm Irminger Current of Atlantic origin is found. As they flow towards Cape Farewell the two mix intensively, and this mixing continues as they round Cape Farewell and proceed northwards along the West Greenland coast.

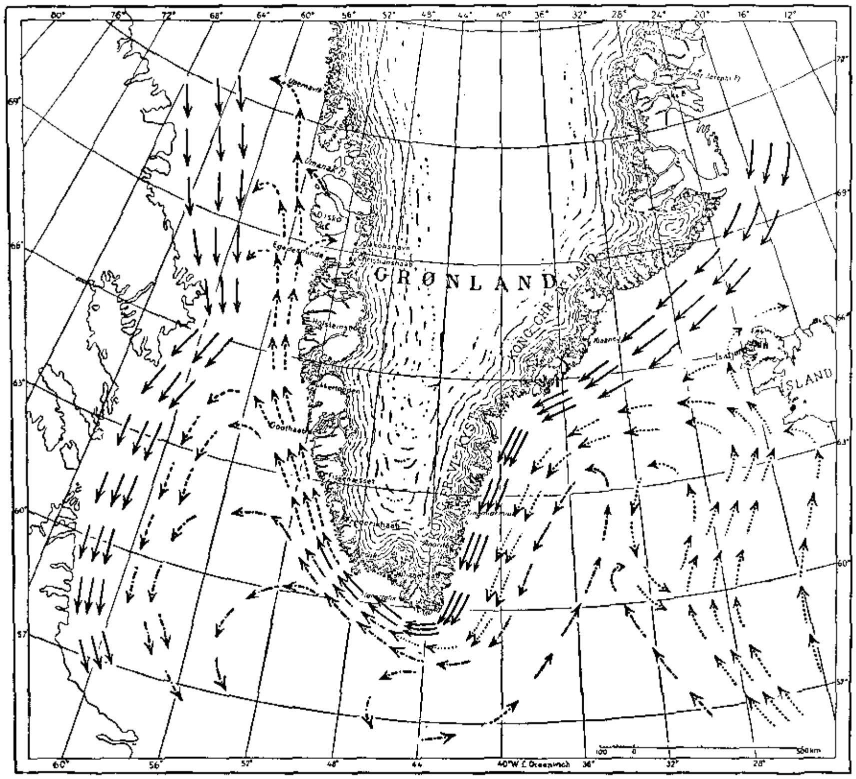
The mixing of the two water masses is the main cause of the very pronounced difference in the oceanographic conditions as found off East Greenland and West Greenland. While the East Greenland Current carries great masses of heavy polar ice, which block that coast for the greater part of the year, the West Greenland current appears, in most years, as temperate and nearly ice free from Frederikshåb to Holsteinsborg. In the polar water nearest the coast of East Greenland, practically no fishing takes place, while the warmer waters of West Greenland provide suitable conditions for a rich population of cod, which has made this region an important fishing area.

The Irminger Current because of its greater salinity has a tendency to sink beneath the Polar Current. This tendency is more pronounced off West Greenland where its core is found in progressively deeper layers as the current proceeds northwards.

As the West Greenland Current flows northward part of it branches off westward, especially in the latitude of Godthåb, while the remainder continues northward losing velocity. Its influence is, however, traceable as far north as Upernavik.

In the western part of Davis Strait and the Labrador Sea is found the Baffin Land Current which transports the cold water masses from

Baffin Bay and the Canadian Archipelago southwards. Off the southern part of Baffin Island, the current is joined by the West Greenland Current to form the Labrador Current.



Baffin Land Current and Labrador Current. West Greenland Current. Irminger Current. Subartic Mixed Currents.

Fig. 5. The sea current round Greenland. (Simplified after Kiilerich, 1939, Hermann and Thomson 1946).

The Water Transport of Currents in the Greenland Area.

From considerations of the water budgets of the Norwegian Sea and the Polar Sea, Sverdrup, Johnson and Fleming (1949) estimated that the water transport of the East Greenland Polar Current through Denmark Strait was of the order of $3.6 \times 10^6 m^3/sec$. This current is joined by the Irminger Current and the two currents together form the West Greenland Current. The following values of the water transport of the West Greenland Current are taken from the results of the Marion Expedition in July-September, 1928 (Smith, Soule and Mosby, 1937).

Table I.

Section	Volume of Flow
Off Cape Farewell	$3.2 \times 10^6 m^3/sec$
,, Invigtut	7.4
,, Fiskenaesset	6.6
,, Godthaab	5.3
,, Holsteinsborg	1.3
,, Egedesminde	1.3
,, Disko Island	0.9

Disregarding the value for the Cape Farewell section which is obviously in disagreement with the other sections, we find that the West Greenland Current off southwest Greenland is about twice the volume of the East Greenland Polar Current off East Greenland, which indicates that the water transport of the East Greenland Polar Current and the water transport of the warm Irminger Current are of about the same magnitude. These values are based on observations from a single season. The mean value of seven sections off Cape Farewell and of four off Ivigtut given by Smith, Soule and Mosby, (1937) gives a water transport of the West Greenland Current of $6.0 \times 10^6 m^3/sec$.

The most pronounced feature indicated in Table I is the very strong decrease in the volume flow between Godthåb and Holsteinsborg. The reason for this is that the main part of the West Greenland Current at this locality bends westward and follows the slope of the ridge between

Greenland and Baffin Island until it joins the Baffin Land Current.

Labrador Current.

The Labrador Current has been described (Iselin 1927) as a cold water stream which flows southward over the continental shelf inside of the comparatively motionless homogeneous mass of North Atlantic water. As compared with the main body of water in the Labrador Sea the current is characterized by its low salinity and temperature. As a result of the investigations of the "Marion" and "General Greene" expeditions (Smith, Soule and Mosby, 1937), the Labrador Current is found to have its origin in the vicinity of Cumberland Sound where the West Greenland and Baffin Land Currents join.

The Labrador Current may be divided into two streams, an inshore and an offshore one. The inshore stream contains the greater volume of cold water and is confined to the continental shelf. This stream enters Hudson Strait on the Baffin Land side, and flows as far as Big Island before it recurves southward to mix with the waters flowing out of Hudson Bay, and flows out past Cape Chidley (Smith, Soule and Mosby, 1937). The offshore stream, which contains waters that are characteristic of the warmer West Greenland Current, tends shoreward near the mouth of Hudson Strait, but does not enter, and continues to flow southward over the continental slope (Smith, Soule and Mosby, 1937). Continuing down the coast, the Labrador Current follows an easy sinuous course which exhibits two major bends, the one between Cape Harrigan and Cape Harrison, Labrador, and the other between Cape Bauld and Funk Island, Newfoundland (Smith, Soule and Mosby, 1937).

The characteristic low temperatures of the Labrador Current persist as the water flows southward, the great stability of the water layers preventing the penetration of solar heat by convection. On reaching the vicinity of Belle Isle Strait, water from the inshore stream is carried at times, through Belle Isle Strait into the Gulf of St. Lawrence. Water moving outward along the southern side of Belle Isle Strait enters into the southward flow of the Labrador Current. Continuing southward of Belle Isle Strait, the Labrador Current meets the northern face of the Grand Banks in the latitude of St. John's, and is split, the slope branch continuing down the edge of the Grand Banks while an inshore branch follows the gully past Cape Race (Smith, Soule and Mosby, 1937). A velocity diagram of the Labrador Current from Smith, Soule and Mosby (1937), is shown in Figure 6 and the axis of maximum flow is indicated by the velocities in miles per day.

The Gulf Stream System.

The Gulf Stream System, which with the Labrador Current, constitutes the main feature of the circulation system of the waters of the Western North Atlantic, may be recognized off the Atlantic seaboard by its comparatively high temperature. Off Halifax, the northern edge of the Stream may be within 230 miles (425 km.) of the coast and as far away as 420 miles (780 km.). Its meanderings, as shown by changing position of its northern edge, have an indirect effect on the waters between it and the coast due to the adjustments in the water masses that are associated with these shifting positions. Gulf Stream waters, breaking away from the system, and carrying with it many forms of marine life associated with more tropical waters, are incorporated in the water masses close to the coasts, and thus directly effect the characteristics of these waters. It is, however, in the area of the confluence of the Labrador Current and the Gulf Stream in the southwestern area of the Grand Banks and to the westward, that large scale mixing processes occur, which provide a type of "slope water" that contributes largely to the characteristic waters that are in contact with the continental shelf, and penetrate at greater depths even into the Gulf of St. Lawrence. These "slope waters" have a profound influence on the waters on the Scotian Shelf, both indirectly through gradual mixing processes, and directly when incursions of these slope waters take place.

Gulf of St. Lawrence.

The circulation of the waters, in the upper layers, in the Gulf of St. Lawrence are in general cyclonic (Bjerkan 1919). Waters that enter the Gulf past Cape Ray, Newfoundland, are deflected to the right and flow northeastward along the west coast of Newfoundland. Along the north

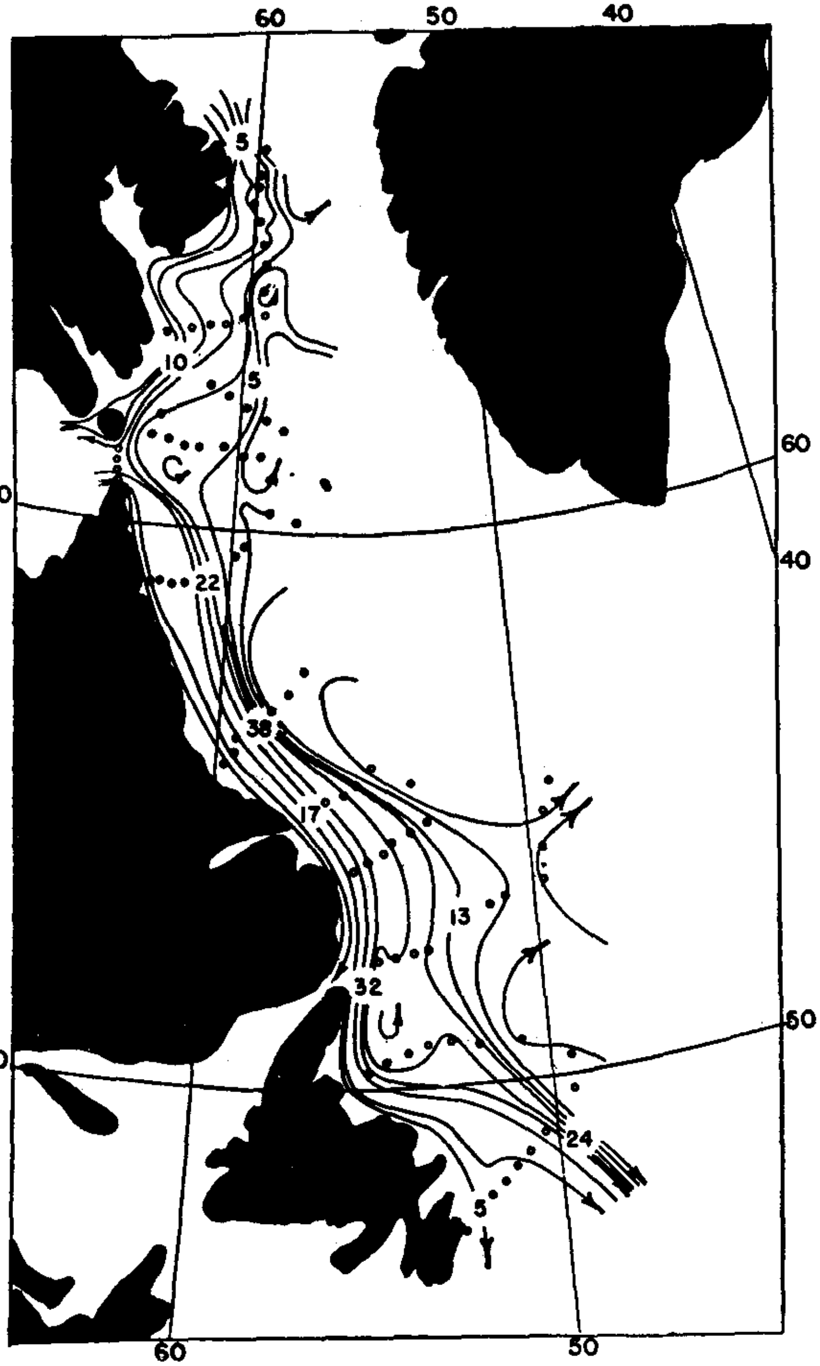


Fig. 6. Labrador current July 22 - Sept. 17, 1928. The velocities shown in miles per day indicate the axis of maximum flow.

shore of the Gulf there is a westward drift of a water type that may have had its origin north of the Strait of Belle Isle, while part of the northeast flow is deflected by Mekattina Bank, lying across the head of the Esquiman Channel. Investigations have shown the circulation of the waters in the Strait of Belle Isle to consist of (a) a progressive inward movement of Labrador coastal water on the north side, (b) a progressive outward movement of Gulf of St. Lawrence water on the south side, (c) a dominant outward flow of Gulf water (Huntsman, Bailey and Hachey, 1954,) and (d) a dominant inflow of Labrador coastal water (Dawson, 1907). The westward flow through Belle Isle Strait rounds Cape Whittle to enter Jacques Cartier Passage. Part of the movement in Jacques Cartier Passage appears to be closely allied to tidal movements, so that there is a net westward movement of water past the western end of Anticosti Island, while at the eastern end there is an eastward movement possibly caused by the shallowing water in the Passage. This current recurves and flows west along the south coast of Anticosti Island where it enters the general circulation of the estuary of the St. Lawrence River.

In the Gaspé Passage between Anticosti Island and the Gaspé Peninsula, there is a preponderance of movement to the east in the form of the Gaspé Current. This current is very strong and its effects can be seen and felt for many miles from the coast. It keeps well to the Gaspé Peninsula but loses a considerable amount of its velocity on passing into the shallows of the southern portion of the Gulf. The waters in this southern Gulf area form a general eastward flow which works toward Cabot Strait, where the main efflux of the Gulf of St. Lawrence takes place. Dawson (1913) described the efflux of the Gulf of St. Lawrence in Cabot Strait as the Cape Breton Current, and as being a constant current flowing to the southeast. The waters borne by the Cape Breton Current on passing through Cabot Strait round Cape Breton Island and flow along the Scotian Shelf.

Scotian Shelf.

Water movements over the Scotian Shelf follow a general southwest direction. Eddies are frequently found, and are considered to be due to bottom configuration. There are frequent incursions of slope water which considerably alter oceanographic conditions from time to time (Hachey 1953). In general, the waters from the Gulf of St. Lawrence flood the eastern portion of the shelf but on proceeding westward are confined to an inshore band. In the western portion, the waters are strongly influenced by offshore waters. The mixture of the two enters the Gulf of Maine and Bay of Fundy circulation after passing Cape Sable at the southern tip of Nova Scotia.

Gulf of Maine and Bay of Fundy.

The physical oceanography of the Gulf of Maine has been extensively dealt with by Bigelow (1928). He described the circulatory movements (Figure 7) as follows: In the eastern side of the Gulf, the tendency is northward along Nova Scotia into the Bay of Fundy on its southern side, northward towards New Brunswick, and out of the Bay along the south side of Grand Manan with a counterflow into the Bay via Grand Manan Channel. Recent observations disagree with earlier ideas regarding residual movements in Grand Manan Channel. McLellan (1951) concluded from hydrographic observations that there is no conclusive evidence of residual currents in Grand Manan Channel, while MacGregor and McLellan (1952) observed net southward movements from Geomagnetic Elektrokinetograph (Von Arx, 1950) measurements.

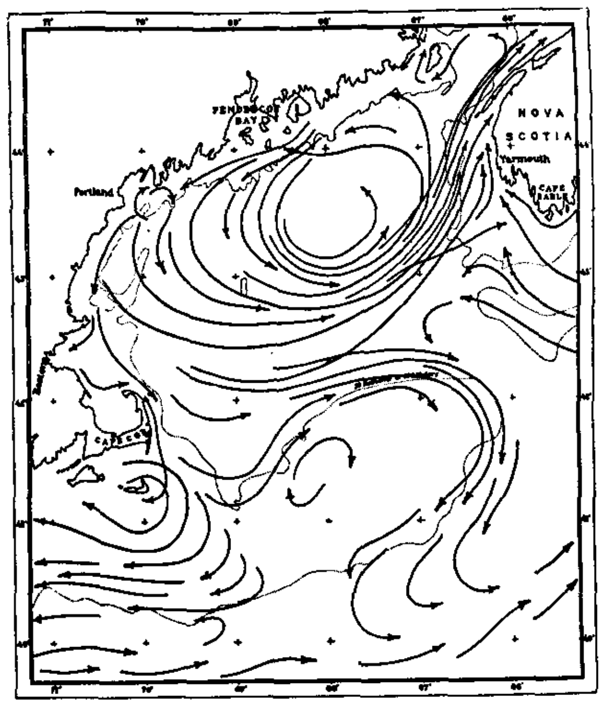


Fig. 7. Schematic representation of the dominant non-tidal circulation of the Gulf of Maine, July to August. (After Bigelow, 1928).

General results of drift bottle experiments show a southwestward movement along the coast of Maine to the vicinity of Cape Elizabeth. Bigelow (1928) states further that off Cape Elizabeth the general drift is southerly. In the region of Massachusetts Bay, two drifts were noted, one anticlockwise around its coast and the other southerly across its mouth and down along Cape Cod. The drift is out to the eastward from Nantucket Sound, and generally southerly and eastward past Nantucket Sound. A general clockwise movement is suggested around Georges Bank.

General Circulation as shown through the Distribution of Ice.

The distribution, characteristics, and various problems of sea ice and icebergs have been discussed extensively in a number of supplements to the Pilot Chart of the North Atlantic, issued by the United States Hydrographic Office. In the supplement of May, 1952, ocean currents and the seasonal distribution of ice are discussed under the title: "Artic Ice and Its Drift into the North Atlantic Ocean".

In the Greenland area, three types of ice are encountered, namely: Storis, West Ice and Winter Ice. Storis refers to the sea ice which has drifted southward along the east coast of Greenland. It is formed in the Arctic basin, often attaining forms and thickness which even the most severe uniform freezing could not produce. It is the result of repeated freezing and rafting of various flows, combined with snowfall. The Storis encountered in south Greenland must be assumed to be mostly 2 and 3 years old. It moves southward along the east coast of Greenland (Figure 8), and in season rounds Cape Farewell to move northward along the west coast of Greenland. This ice clearly outlines the influence of the East Greenland Current.

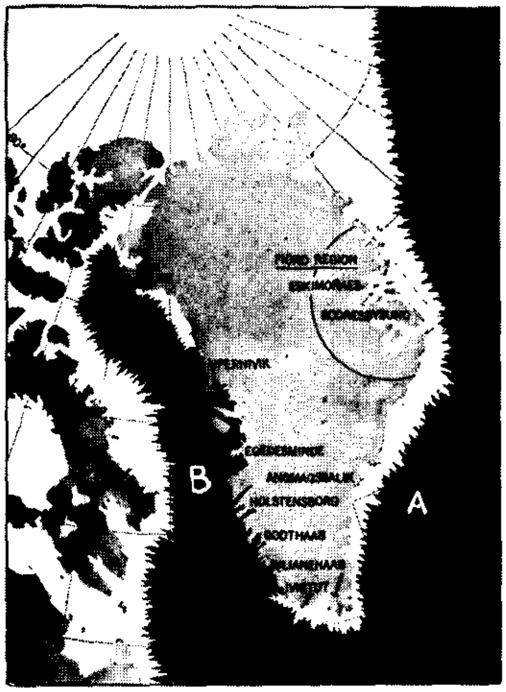


Fig. 8. Distribution of Storis (A) and West Ice (B). After U.S. Hydrographic Office Suppl. to Pilot Chart of North Atlantic Ocean, May, 1952.

The extension of Storis varies much with the seasons and from year to year. Off Southwest Greenland, the first Storis often appears at Cape Farewell in late January, but only in small scattered amounts. It arrives in greater quantities in March, and during the months of April to June it has its greatest extension along the west coast. In severe ice years, the Storis reaches north of Godthåb, but in most years, the northern limit is found between Frederikshåb and Fiskenæsset. From July onwards the ice decreases in quantity and from the middle of August to the end of the year all of the west coast of Greenland is generally free from Storis. These remarks refer to conditions as found in recent years. The distribution of ice, however, has been greatly affected by the climatic changes which have taken place during the last fifty years.

In Figure 9, the average limits of the Storis in April and in August are shown for the two periods 1898-1913 and 1919-1942. The limits shown for the first period are those given by Speerschneider (1917), and the limits for the second period are the 50% frequency curves given by the Deutsches Hydrographisches Institut (1950). Although the limits were calculated in somewhat different ways, it is probable that they give a fairly correct picture of the regression of ice during this century.

West Ice resembles Storis but reaches Baffin Bay through the various channels in the Canadian Archipelago. In Greenland, west ice is generally confined to the northwestern portion but favourable wind conditions may move it to the Greenland part of the Davis Strait sector.

Winter Ice is that which is formed in the harbours and fiords during the coldest part of the

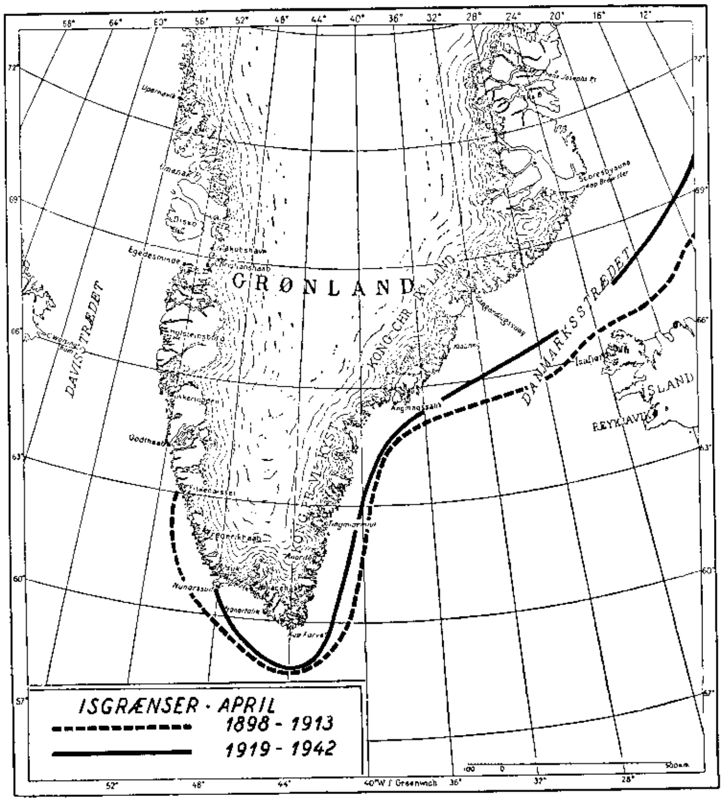


Fig. 9A. Average limits of Storis for April in Greenland waters.

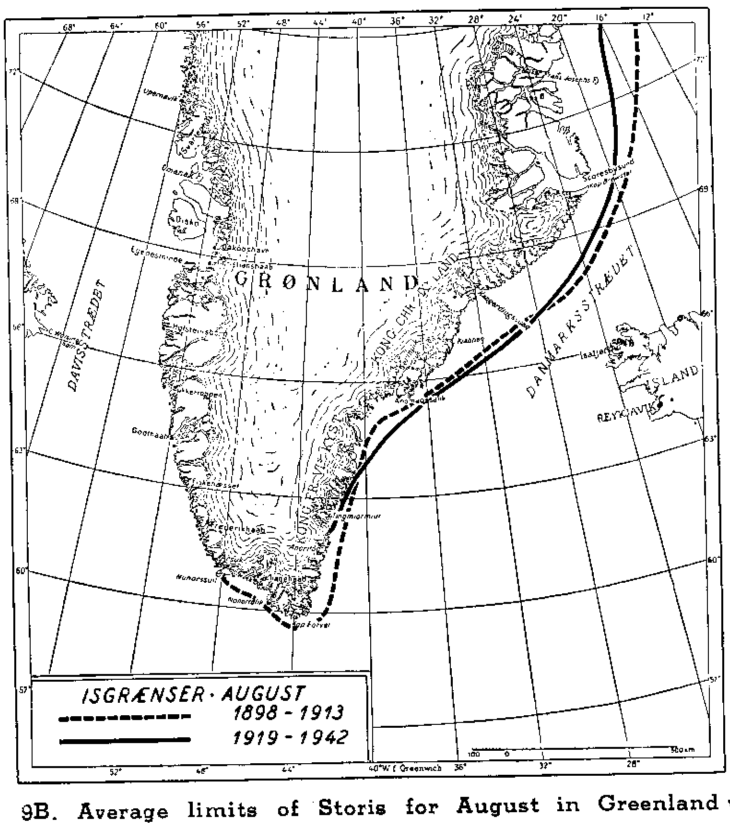


Fig. 9B. Average limits of Storis for August in Greenland waters.

year and may attain a thickness up to 3 feet. It is most prevalent in East Greenland and north of Davis Strait. Winter Ice plays an important role in sealing up icebergs and preventing their dispersal.

In discussing the Arctic Ice in relation to the Labrador Current, it is apparent that the two chief sources of ice encountered in the western Atlantic are Davis Strait and Foxe Channel. The pack ice from Baffin Bay reaches Hudson Strait in late October or early November and is there joined by heavy flows from Foxe Channel. The combined streams move south along the coast arriving off Belle Isle Strait in December. Heavy flows of true arctic type often extend 100 miles east of the entrance of Belle Isle Strait, and much of this ice enters the Strait. From January until after March, currents and winds govern the distribution of ice often as far south as latitude 46°N. Large quantities of it may drift along the eastern edge of the Grand Banks, while some of it extends along the east coast of Newfoundland, around Cape Race and thence southward and southwestwardly over the neighbouring banks. Large quantities of ice which enter the Gulf of St. Lawrence through Belle Isle Strait leave it on the Cape Breton side of Cabot Strait and then spread southward and southwestward toward Sable Island. The distribution of field ice on the Atlantic Coast of North America (after Huntsman (1930) and the U.S. Hydrographic Office Supplement to the Pilot Chart of the North Atlantic Ocean, 1948) is shown in Figure 10.

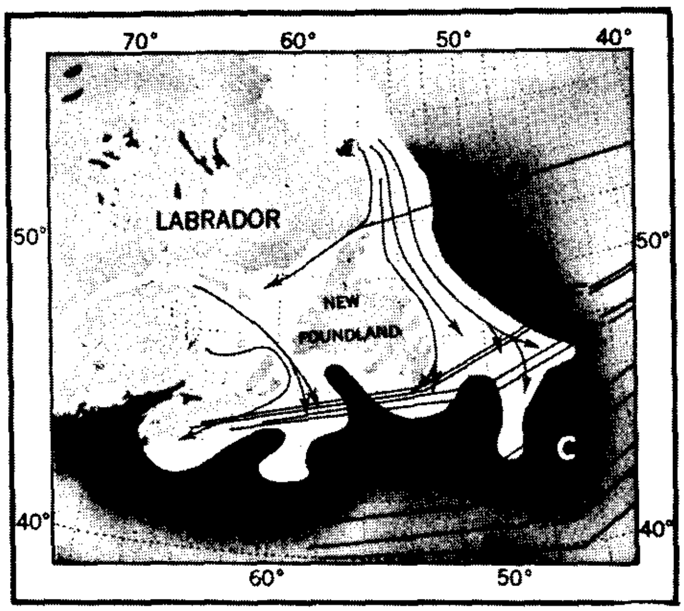


Fig. 10. The distribution of Field Ice south of Newfoundland. The lettered lines represent steamer tracks. (After Huntsman, 1930, and U.S. Hydr. Off. Suppl. to Pilot Chart of N. Atlantic Ocean, May, 1952).

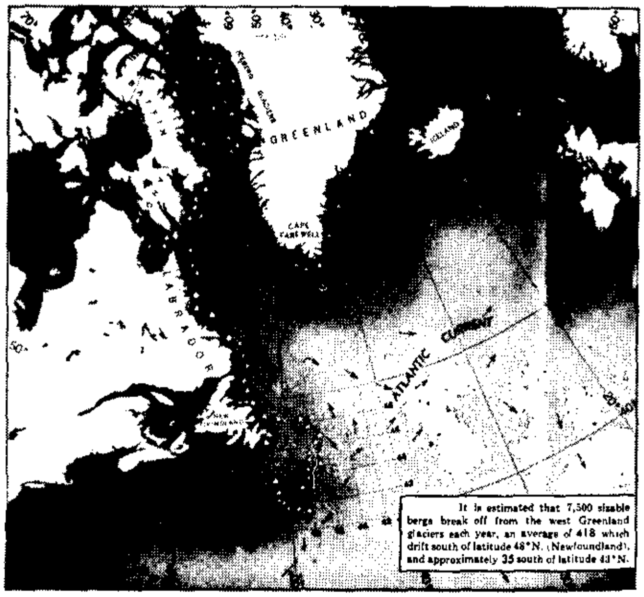


Fig. 11. Drift of icebergs from their source into the North Atlantic. (After U.S. Hydr. Off. Suppl. to Pilot Chart of the N. Atlantic Ocean, May, 1952).

In addition to pack ice, which is entrained by the various currents and shifted about by the winds, there are the icebergs which because of their greater draft are less effected by the winds and more by the currents. Figure 11 shows the drift of icebergs from their source into the North Atlantic. The principal source of icebergs in the North Atlantic is the west coast of Greenland, although a few are produced from glaciers on Ellesmere Island. Many icebergs from East Greenland are transported to Cape Farewell, where scattered bergs are met as much as 100 to 200 miles south of this promontory. The main stream of bergs drifts northward as far as Holsteinsborg. North of Holsteinsborg is the northern iceberg area which is the main source of icebergs encountered near Newfoundland. These bergs are carried by the northward movement of that part of West Greenland Current which enters Baffin Bay. They then enter the southerly drift along the east coast of Baffin Land and enter the eastern ends of Frobisher Bay and Hudson Strait, following the same general pattern as the field ice from Baffin Bay. They are found as far west as Big Island in Hudson Strait and everywhere in Ungava Bay. Icebergs which follow the inshore part of the Labrador Current are, in general, caught up in the various breaks in the coast line, while those which are carried in the offshore branch of the Labrador Current, reach the tail of the Grand Banks.

The Labrador Current and the ice follow a definite course from the Arctic to the tail of the Grand Banks. The effect of wind upon the drift of icebergs being small, their subsequent movements depend largely upon the complex and variable current patterns which exist along the boundary between the Labrador and the Atlantic Currents. During the latter part of March and the first part of April, the flooding Labrador Current holds closely to the eastern slope of the Grand Banks and sometimes curls around the tail and extends for a considerable distance northwestward along the southwestern slope. As the volume of the discharge increases, the mixing zone between the two currents moves further offshore to the southwest, south, and southeast of the tail, and bergs fan out along the edge of the Gulf Stream System. During the summer however, the Labrador Current dwindles in volume in this region, and bergs remain near the banks or are turned northeastward by the currents before reaching the tail of the Grand Banks. During this cycle in the flow of the Labrador Current, the volume of flow of the Atlantic Current is also changing, and the resulting changes in relative strength of the two currents produce complicated changes in the location of their common boundary and in the courses, drift rates, and life expectancy of bergs reaching this vicinity (Supplement to Pilot Chart of North Atlantic Ocean, May, 1952).

Horizontal Distribution of Temperature and Salinity in the Summer.

The charts presented (Figures 12-15) in the following discussion of temperature and salinity are based on observations made over a network of stations used by the Atlantic Oceanographic Group and the Newfoundland Fisheries Research Station of the Fisheries Research Board of Canada. These networks are illustrated in Figure 1. The 1950 post seasonal cruise of the International Ice Patrol from Labrador to Greenland was used to give data for the Labrador Sea and West Greenland Current. A considerable amount of additional data collected by various United States and Canadian agencies were used in a supplementary manner. All data used were collected during the period between the last week of July and the first week of September, 1950. These diagrams form part of a series presented by Bailey and Hunt (1954).

Most of the major ocean currents in the area are recognizable from the surface temperature distribution (Figure 12). South of Greenland, the northward flowing West Greenland Current has surface temperatures ranging from 2°C. inshore to 8°C. offshore. The main body of the Labrador Sea exhibits temperatures above 8°C. while the southward flowing Labrador Current shows the influence of decreasing latitudes and mixing with warmer waters from offshore as its temperatures increase from 3°C. in the north to 7°C. at the tail of the Grand Banks. In the triangle between the Grand Banks and the Scotian Shelf, the slope water is seen to increase from a narrow band off Cape Cod to a wide one south of Cape Ray. Temperatures in the slope water are higher than 17°C. but less than 25°C. In the longitude of 60°W. an eddy from the northern edge of the Gulf Stream may be seen intruding into the slope water. Surface temperatures at the northern edge of the Gulf Stream are generally greater than 25°C., and traces of it just reach into the diagram.

Over the shallow portions along the coast and over the banks, surface temperatures show only small variations from one area to another, ranging in general from 13° to 16°C. Along the south coast of Newfoundland where temperatures are less than 15°C., there is the suggestion of a movement from the east coast of Newfoundland into the Gulf of St. Lawrence. Increasing temperatures along the Gaspé coast illustrate the warming of the Gaspé Current as it loses velocity and enters the shallow area in the southwestern Gulf. Along the southeast coast of Nova Scotia, the lowered temperatures off Halifax illustrate the upwelling of colder sub-surface waters. In the area south of Cape Sable, temperatures as low as 10°C. are observed which appear to be the result of strong tidal mixing over the shallows in this area. In the Bay of Fundy area, tidal mixing is extremely important. Temperatures there, during this period, are somewhat higher than on the Nova Scotia coast.

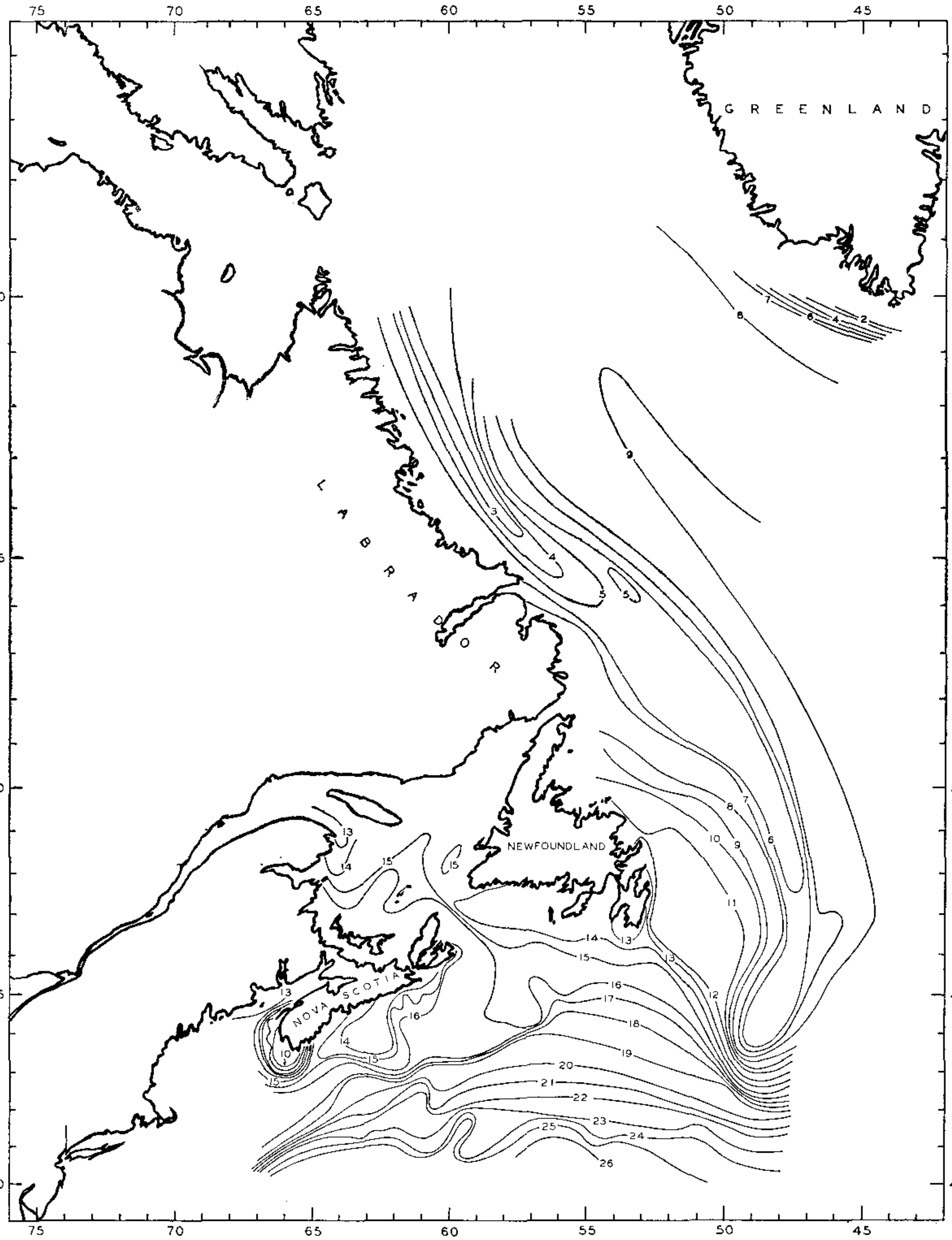


Fig. 12. Surface temperature distribution, August, 1950.

At 100 metres (Figure 13) the nature of the several currents is better revealed by their temperatures than at surface where solar heating and wind stirring tend to obliterate the contrasts. The influence of the bottom topography is most pronounced because the main banks are of depths less than 100 metres. At this level, the nature of the West Greenland Current is revealed. The

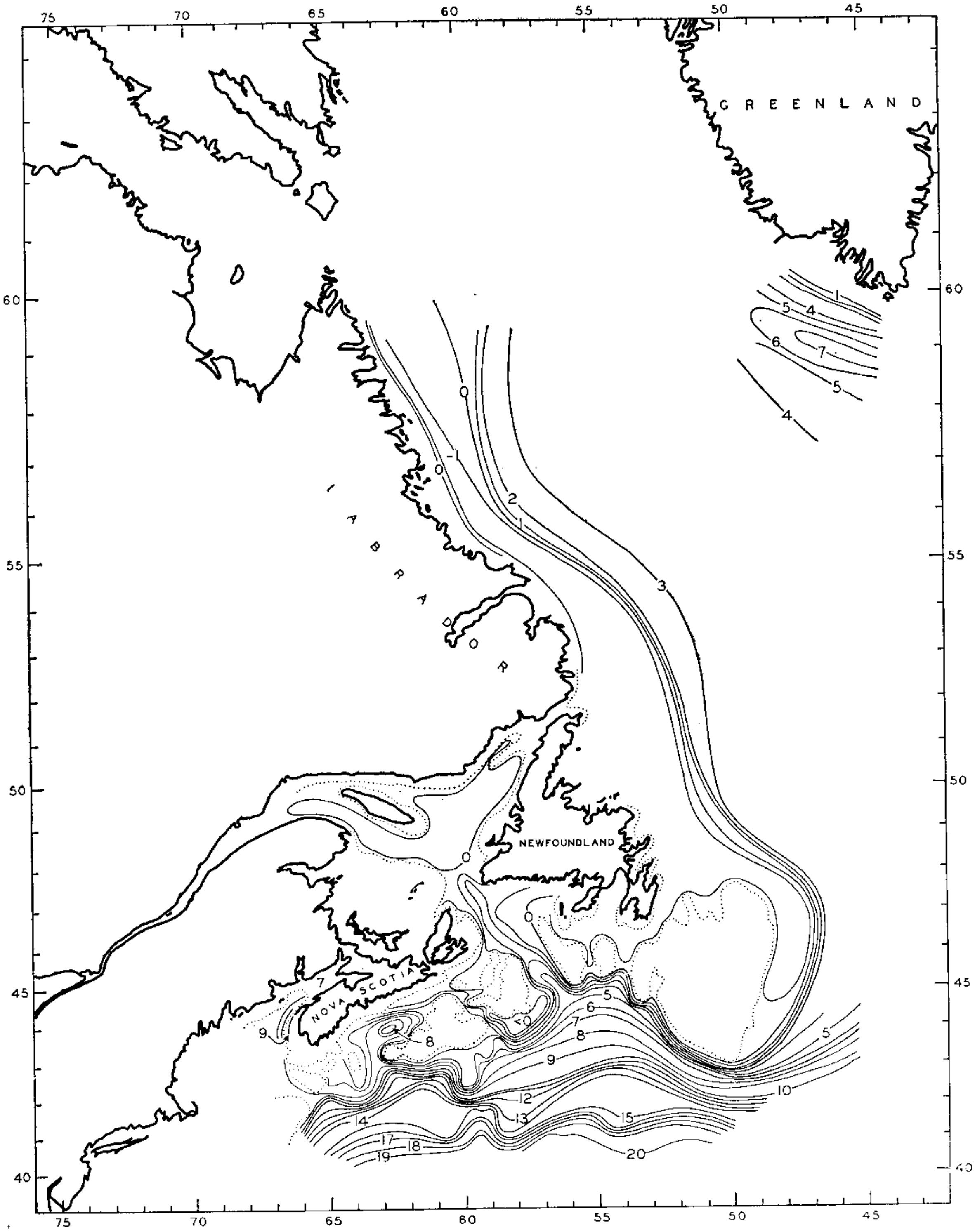


Fig. 13. Temperature distribution at 100 metres, August, 1950.

water from the East Greenland Current is inshore while a tongue of Irminger water with temperatures greater than 7°C. flows northwestward offshore. The main body of Labrador Sea water has temperatures between 3° and 4°C. This level appears to cut through the top of the large mass of nearly homogeneous water found in the Labrador Sea. In the Labrador Current, tem-

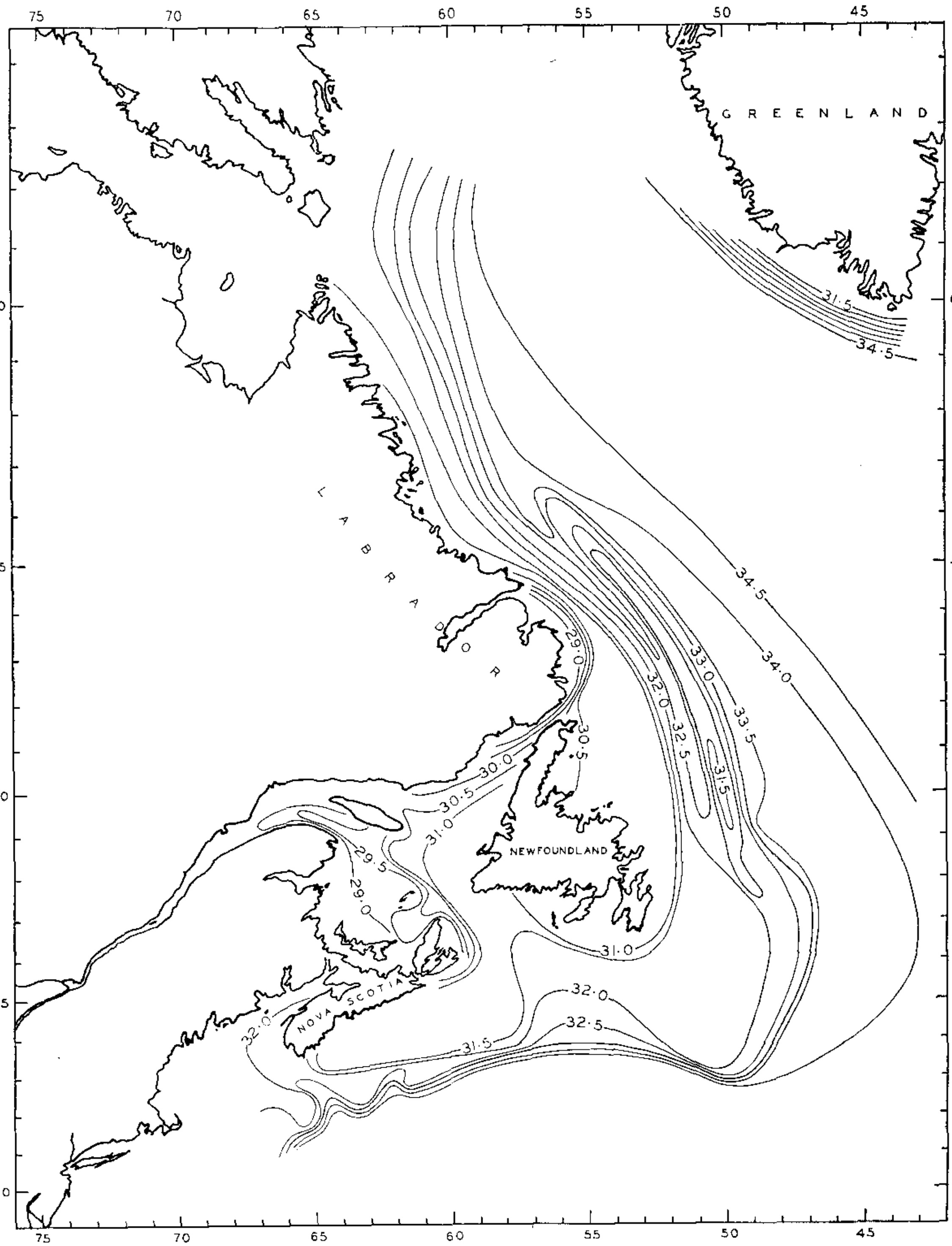


Fig. 14. Surface salinity distribution, August, 1950.

peratures are less than -1°C. A major feature of the chart is the extent of this large mass of Arctic water which extends from the Arctic Sea to the edges of the Grand Banks at latitude 45°N., half way to the equator. This Arctic water does not reach the Gulf of St. Lawrence nor the Scotian Shelf. In the Gulf of St. Lawrence, temperatures at 100 metres are all below 0°C. This water

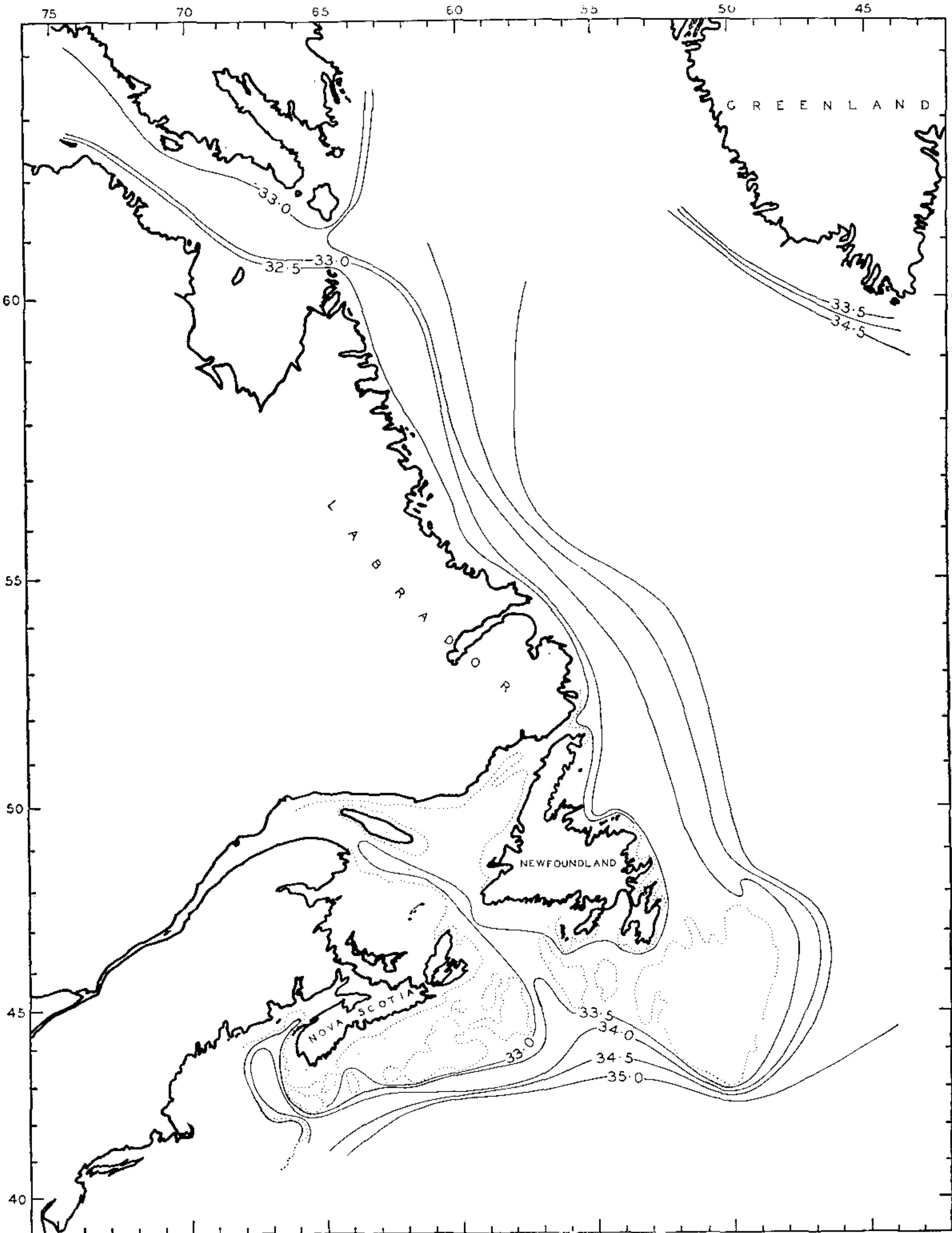


Fig. 15. Salinity distribution at 100 metres, August. 1950.

probably originated, in part, outside of the Gulf at an earlier time, and in part, was formed inside the Gulf through winter chilling. Lauzier and Bailey (1952) estimate that the amount of locally formed cold water (less than 0°C.) to be between one-third and one-half of the total volume of cold water found in the Gulf. Outside of the Gulf of St. Lawrence in the Laurentian Channel, the outline is noted of the upper portion of the wedge of slope water that reaches into the Gulf of St. Lawrence at lowered depths. This wedge reaches well into the estuary of the St. Lawrence river. At the eastern edge of the Scotian Shelf, a small isolated body of Arctic water is in evidence. The exact origin of this water is not certain but it may have come through the channels between the banks south of Newfoundland, or have worked westward from the tail of the Grand Banks after a flooding in the early spring, as suggested by Bjerkan (1919).

South of Nova Scotia in the deep channels that penetrate the shelf, there is a small body of relatively warm water which appears to have been left from an incursion of slope water that took place at an earlier date. The undulating boundary between the slope water and the shelf probably causes this phenomenon to occur at fairly frequent intervals when the slope water lies close to the edge of the shelf, as has been observed at times. Hachey (1953) describes such a case that occurred during the winter of 1949.

The distributions of salinity are based almost wholly on observations made by the Atlantic Oceanographic Group and the Newfoundland Fisheries Research Station. One section across the Labrador Current from South Wolf Island, Labrador, to Cape Farewell, Greenland, made by the International Ice Patrol, was used, as well as, occasional observations made by the Canadian Hydrographic Service.

Most of the waters sampled were coastal with salinities generally less than 33.0.$^{\circ}/_{oo}$ Some slope water with salinities as high as 34.5$^{\circ}/_{oo}$ was observed. There were no observations of Atlantic water in southern sectors which has salinites higher than 35.0$^{\circ}/_{oo}$. In the Labrador Sea salinities were greater than 34.5$^{\circ}/_{oo}$ but less than 35.0.$^{\circ}/_{oo}$.

The salinity distribution at the surface (Figure 14) shows that the waters in the Labrador Current are composed of many tongues and eddies. This is particularly true in the offing of Belle Isle. Smith, Soule and Mosby, (1937), show an eddy in the same area and suggest that this region is unmistakably associated with currents coming from farther south in the Atlantic.

The efflux from Hamilton Inlet appears to have some influence on the surface waters off southern Labrador and in the vicinity of the Strait of Belle Isle. The lowered salinities in the southwestern Gulf show the effect of the St. Lawrence river discharge. In Cabot Strait, the outflow is traceable to the Scotian Shelf, and an inflow of surface water is evident along the west coast of Newfoundland.

At the 100 metre level (Figure 15), salinities were everywhere above 33.0$^{\circ}/_{oo}$ except in Ungava Bay. In Hudson Strait there were no observations in 1950 but the isohalines were drawn after Bailey and Hachey (1950) to illustrate how the ocean water under-runs the coastal waters and reaches inland in the deep channels to bring ocean water near the shores. This is clearly evident in Hudson Strait, the Laurentian Channel, and the Fundian Channel. In the Hudson Strait and the Bay of Fundy, this condition has a profound influence on the waters in the two areas because they are thoroughly mixed through tidal stirring.

WATER CHARACTERISTICS

The Water Masses in the Greenland Area.

The distribution of the different water masses in the Greenland area is best illustrated by five vertical sections, the locations of which are shown in Figure 1 (I-V). The sections were selected from years in which the summer conditions in the sea around Greenland were regarded as normal according to the standards of recent years.

In the section (I) off southeast Greenland (Figure 16A), the Polar Current, with temperatures below 2°C. is found over the shelf, and has a width of only about 30 miles. Beneath, and outside the Polar water, the warm Irminger Current is found. The core of this current is found off the slope of the shelf, and has temperatures above 5°C. and corresponding salinities greater than 35$^{\circ}/_{oo}$. On account of its greater density, the Irminger Current has a tendency to sink beneath the less saline Polar water. Thus a relatively warm bottom water is found nearly everywhere along the East Greenland shelf, where depths exceed 200 metres.

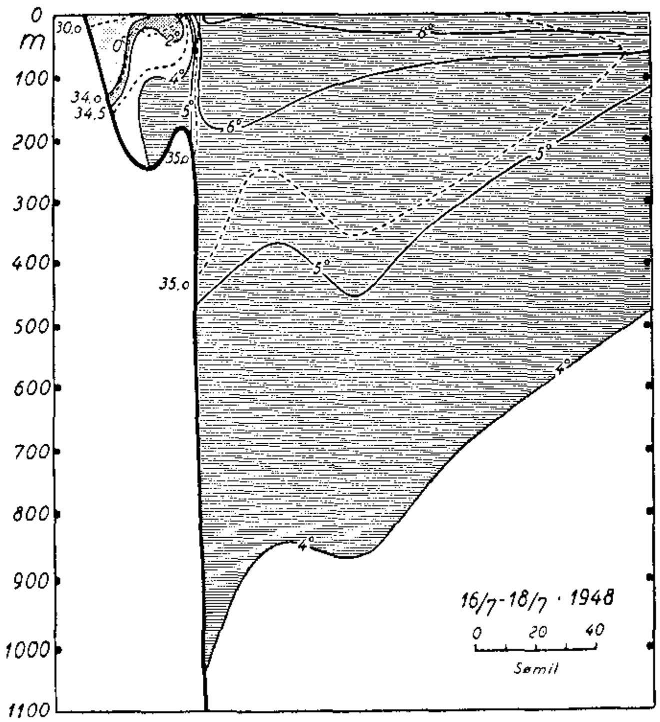


Fig. 16A. Temperature and salinity, Section I, off southeast Greenland, 16-18 July, 1948.

In Section II off Frederikshåb (Figure 16B), on the west coast, both the Polar water and the Irminger water are found, but due to mixing and summer warming the difference between the temperatures of the two currents is much less in this area than on the east coast. The remainder of the Polar Current is again found nearest the coast in the upper 100 metres, but the temperature in its core is now only slightly below 1°C. The warm current appears near the coast as an undercurrent, but reaches to the surface at approximately 60 miles offshore. The temperatures in the core of the warm current are between 4° to 5°C. and its salinity is less than 35°/oo.

Section III, (figure 16C), illustrates the oceanographic conditions of Fylla Bank, one important to the Greenland Fisheries. In this section, the current has lost its polar characteristics. In the deep channel between the bank and the coast, and off the western slope at a depth of

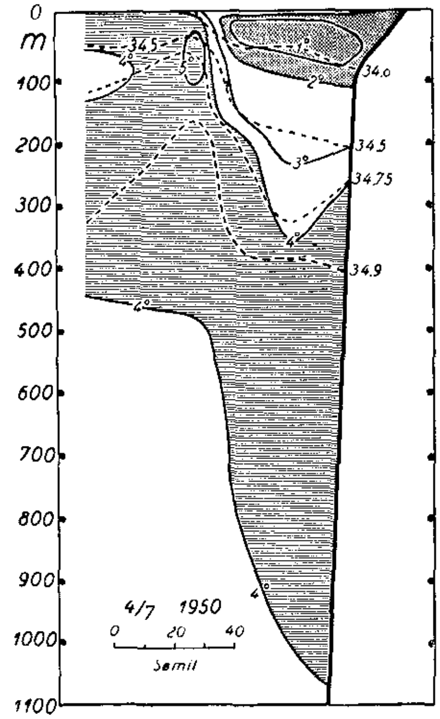


Fig. 16B. Temperature and salinity, Section II, off Frederikshaab, 4 July, 1950.

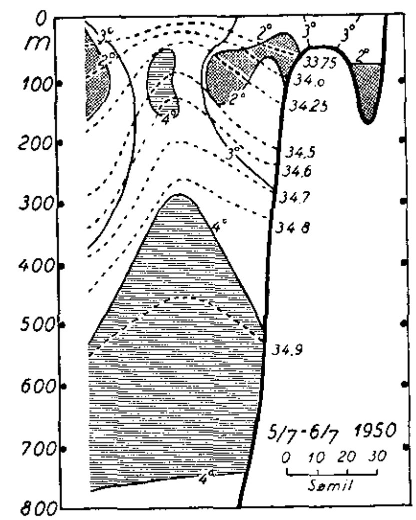


Fig. 16C. Temperature and salinity, Section III, Fylla Bank, 5-6 July, 1950.

about 100 metres, water colder than 2°C. is found. This is the last trace of the East Greenland Polar Current. Over the shallowest part of the bank summer warming has increased the temperature to more than 3°C. In the main portion of the section, water with temperatures between 2° and 4°C. is found, which is produced through the mixing of Polar water and Irminger water. Irminger water warmer than 4°C. is found in rather large quantities at depths between 300 and 800 metres. In the westernmost part of the section, colder water is encountered, which is a branch of the West Greenland Current which has turned west and southwestwards and mixed with water from the Canadian Polar Current.

In Section IV off Holsteinsborg (Fig. 16D), the western part is dominated by Canadian Polar water with temperatures less than -1°C. The West Greenland Current has decreased considerably in its passage from the Godthåb section, and this is because the main body of water follows the west going branch of this current. The remaining part of it is, however, of sufficient volume to keep the cold water of the Canadian Current away from the Greenlandic slope and shelf.

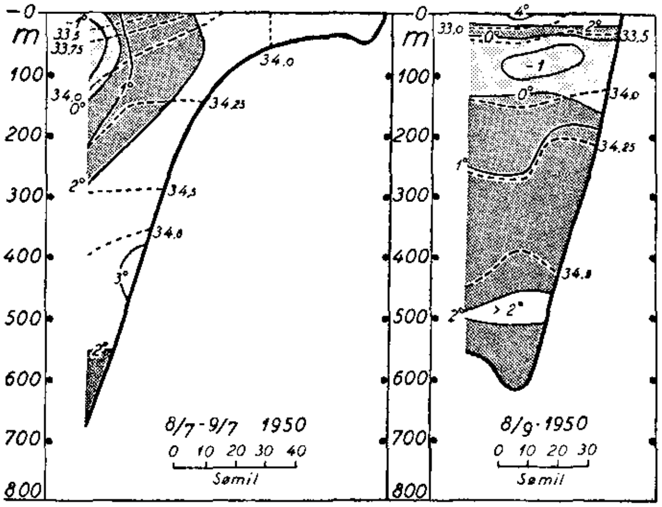


Fig. 16D. (left) and E. (right). **Temperature and salinity, Section IV (D), off Holsteinsborg, 8-9 July, 1950, and Section V (E), southeastern part of Baffin Bay, 8 September, 1950.**

The conditions as found in the southeastern part of Baffin Bay are illustrated by section V. (Figure 16E) Below the summer warmed layer, of approximately 30 metres thickness, a cold layer with negative temperatures formed by winter chilling, extends down to about 150 metres. Below this cold layer, the warm West Greenland Current exists. Its maximum temperature being slightly greater than 2°C., the current still has sufficient strength to cause positive temperatures to a depth of 1000 metres in the southeastern part of Baffin Bay.

The above sections illustrate the oceanographic conditions as found during what may be regarded as normal years. The conditions, however, can differ considerably from these because they are greatly influenced by fluctuations in the strength of both the East Greenland and Irminger Currents, and by variations in the meteorological conditions, particularly the winter temperatures off West Greenland.

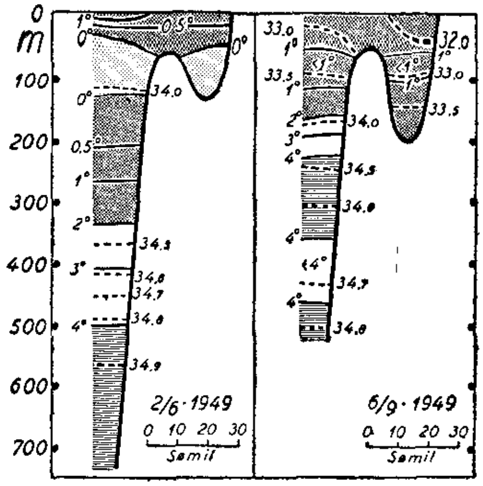


Fig. 17A. Temperature and salinity in sections across Fylla Bank in a "cold" year, 2 June and 6 Sept., 1949.

To illustrate the variations in the oceanographic conditions that take place in the waters off West Greenland Figures 17A and B, representing oceanographic conditions over Fylla Bank in a cold year, 1949, and a warm year, 1947, are to be compared with those shown in Figure 16C 1950. Conditions in 1947 were the warmest attained in the post-war period. In June, 1949, Fylla Bank was surrounded by ice-cold Polar

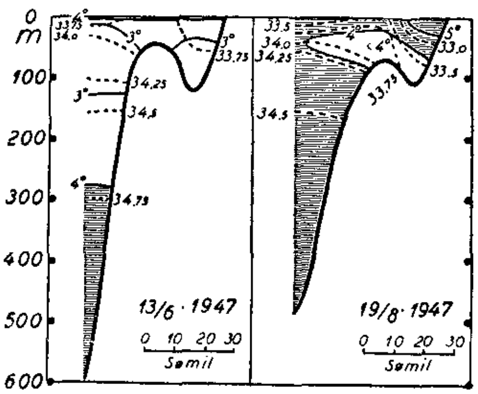


Fig. 17B. Temperature and salinity in sections across Fylla Bank in a "warm" year, 13 June and 19 August, 1947.

water, and even in September, temperatures at depths between the 50 and 100 metre level were less than 1°C. In 1947, almost no trace of Polar water was found, and August temperatures exceeded 4°C. over the greater part of the section.

Labrador Water.

On the basis of observations made in 1931 by the "Marion" and "General Greene" expeditions, temperature-salinity correlation curves were obtained for all observations below a depth of 50 metres in the Labrador Current (P. 112, Smith, Soule and Mosby, 1937). These correlation curves show that the main water mass of the Labrador Current is a mixture of two characteristic waters as follows:

(1) Baffin Land water exhibiting average temperatures of -0.5°C. and average salinities of 33.5‰.

(2) West Greenland water exhibiting temperatures as high as 3.8°C. and salinities as high as 34.6‰.

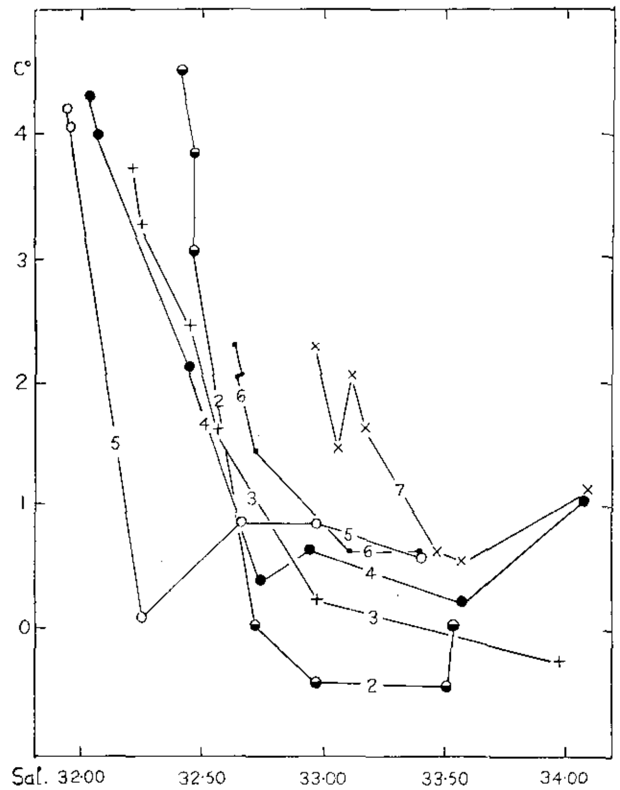


Fig. 18. Temperature-salinity correlation curves for Labrador coastal waters in September, 1948.

A temperature-salinity diagram constructed from the 1948 data collected by the "Haida" (Bailey and Hachey, 1950) is given in Figure 18. As the "Haida" data were collected only along the axis of the Labrador Current and to depths not exceeding 300 metres, the diagram gives preeminence only to two characteristic water masses as follows:

(1) Water exhibiting temperatures greater than 3.0°C. and salinities between 32.0 and 32.5‰.

(2) Water exhibiting temperatures between -0.5°C. and 1.0°C. and salinities between 32.7 and 34.0‰.

The first of these is therefore representative of the coastal contributions to the Labrador Current, and confined, in the main, to the upper fifty metres within the coastal belt. The second is obviously water of the Baffin Land Current, while no sampling of the waters of the West Greenland Current is in evidence.

Gulf of St. Lawrence and Strait of Belle Isle.

Temperature-salinity relationships of the waters in and adjacent to the Strait of Belle Isle are illustrated in Figure 19. The T-S diagram

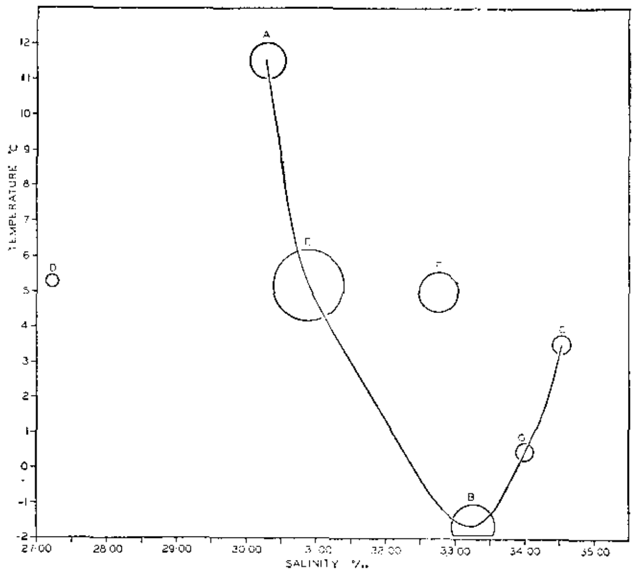


Fig. 19. Temperature-salinity relationships from the observations of the Belle Isle Strait expedition of 1923.

is based on observations by the Belle Isle Strait expedition of 1923 as presented by Huntsman, Bailey and Hachey (1954). The main water masses involved in the Strait as well as those which do not enter are indicated by circles.

On the basis of the T-S diagram, the waters in and adjacent to Belle Isle Strait are, in general, composed of three different water masses having the following characteristics:

(1) A-water, with temperatures greater than 11.0°C. and salinities of approximately $30.5^{o}/_{oo}$, found in the upper 25 metres in the western portion of Belle Isle Strait and in the Gulf of St. Lawrence, has the characteristics of waters from the surface layers of the Gulf of St. Lawrence.

(2) B-water, having temperatures and salinities of approximately -1.6°C. and $33.3^{o}/_{oo}$ respectively, is true Arctic water, which forms a layer 200 metres thick near the Labrador coast, and which decreases to a thickness of 100 metres near the edge of the Labrador shelf.

(3) C-water, exhibiting temperatures and salinities generally greater than 3.5°C. and $34.5^{o}/_{oo}$ respectively, is Labrador Sea water modified by waters of the West Greenland Current (3.5°C., $34.8^{o}/_{oo}$ Smith, Soule and Mosby, 1937). This water mass was not found in the Strait as such.

In addition to the above mentioned water masses, there are four distinct water masses that show influence of varying degrees upon the waters in Belle Isle Strait, with the following characteristics:

(4) D-water, with temperatures and salinities of 5.0°C. and $27.2^{o}/_{oo}$, respectively, found at the surface, is Labrador coastal water that has been largely influenced by land drainage.

(5) E-water, having temperatures between 4.5 and 6.5°C. and salinities between 30.5 and $31.5^{o}/_{oo}$ is Labrador coastal water. This water comprises the greater portion of the surface waters on the northern side of the Strait and near Belle Isle.

(6) F-water, exhibiting temperatures from 5.0 to 6.0°C. and salinities from 32.5 to $33.1^{o}/_{oo}$ comprises the surface layer of the Labrador Current. This water mass has the characteristics of the surface waters of the Labrador Current, which exhibits normal seasonal changes in its southward progress.

(7) G-water, with a temperature of 0.5°C. and salinites between 33.8 and $34.1^{o}/_{oo}$ found at the greater depths in the Labrador current, is Labrador Sea water (Dunbar, 1951).

In addition to the mixing that takes place between the different water masses that fall on the T-S curve, there is evidence of direct mixing between the isolated water masses and those on the T-S curve. A-water mixes with B-water, D-water with E-water, and F-water with B-water. There is however, no evidence of E-water mixing with F-water.

Scotian Shelf.

The waters of the Scotian Shelf are coastal in nature and are distributed in a three layer system described by Hachey (1942). The upper layers, of comparatively low salinity display a great variation in temperature with the seasons. The salinity of this layer increases outwards from the coast from less than $30^{o}/_{oo}$ to approximately $33^{o}/_{oo}$ and temperatures are generally higher offshore than at the coast. The intermediate layer appears throughout the greater part of the year, as a layer of minimum temperatures, generally less than 5.0°C. This layer results from waters which flood the area from the northeast (Hachey 1938), and is often divided into two distinct phases by the outer banks (McLellan and Trites, 1951). Inshore, the coldest water in the intermediate layer is associated with a salinity of approximately $32.5^{o}/_{oo}$, while beyond the outer banks it displays salinities very close to $33.0^{o}/_{oo}$. The deep layer in the coastal water region is warmer and more saline than the intermediate layer and is present only beyond the outer banks and over the deeper portions of the shelf.

Hachey (1942) has summarized the characteristics of the three main layers as follows:

TABLE II.

		Thickness (m.)	Temperature (°C.)	Salinity (°/°°)
"Upper"	Spring	0–85		<32.0
	Summer	20–60	>5	
	Winter	0–115	<5	
"Intermediate"	Spring	40–85	<5 (generally)	32.0–33.5
	Summer	55–90		
	Winter	30–85		
"Bottom"	Spring	Limited by bottom configuration.	>5 (generally)	>33.5
	Summer			
	Winter			

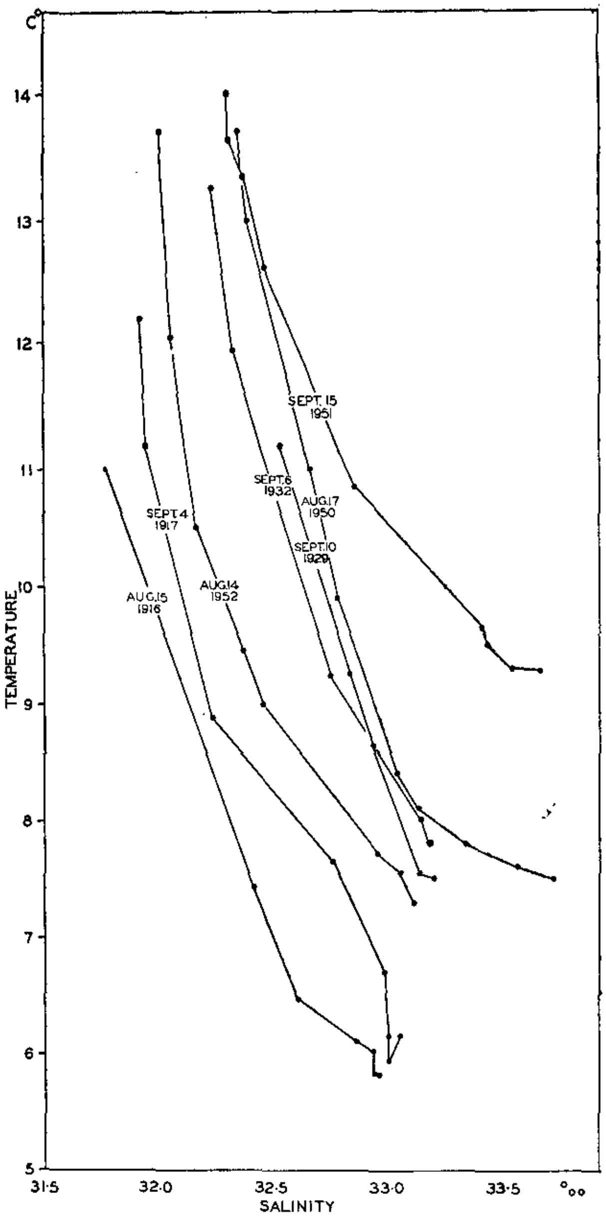


Fig. 20. T-S diagrams for the summer months at Prince Station 3 and at Station 3 for different years.

Readers who are interested in further details of the origins of the waters of the Scotian Shelf, are referred to a paper by McLellan (1954).

Gulf of Maine and Bay of Fundy.

The waters in the Gulf of Maine are chiefly derived from an influx of surface water past Cape Sable and a "draw-in" of water from the edge of the continental shelf. The surface water temperatures and salinities are extremely variable depending upon the seasons of the year, the surface salinities being less than 32.0°/₀₀. At approximately the 100 metre level, the waters may be considered as being unaffected by seasonal variations having an average temperature of 7.5°C. and an average salinity of 33.5°/₀₀. This is water from the edge of the continental shelf. Lying below this is "slope water" having a temperature of 10.3°C. and a salinity of 34.8°/₀₀. It is the mixture of these waters in varying proportions which form the waters in the Bay of Fundy and Gulf of Maine. Typical T-S diagrams from the Bay of Fundy are shown in Figure 20, the location of station 3 being Latitude 44°39'N., Longitude 66°28'W.

The Slope Water off the Scotian Shelf.

The slope water off the Scotian Shelf forms a well defined band between the coastal waters and the Gulf Stream. Its boundaries fluctuate widely with no apparent systematics, sometimes transgressing upon the shelf (McLellan, Lauzier, and Bailey, 1953).

The T-S curves for selected stations occupied during November, 1951 (Figure 21) give a picture of the way in which slope water is formed. The curve lying to the lower right of the figure represents the characteristics of Central Atlantic waters deeper than 400 metres (Iselin, 1936). Stations 25 and 70 taken well within the Gulf Stream, illustrate surface T-S relationships there. The deeper waters at these stations fall almost exactly along Iselin's curve, though water found at 600 metres in the Sargasso Sea was found at 400 metres at station 25. Station 31 is representative of the deeper coastal stations, and station 120, off the Newfoundland coast, shows the full development of the cold intermediate layer. The other stations belong to the slope water regime or the outer fringe of the coastal waters.

At the surface, all observations lay close to a straight line running from the surface characteristics of the most coastal (Station 31) to those of

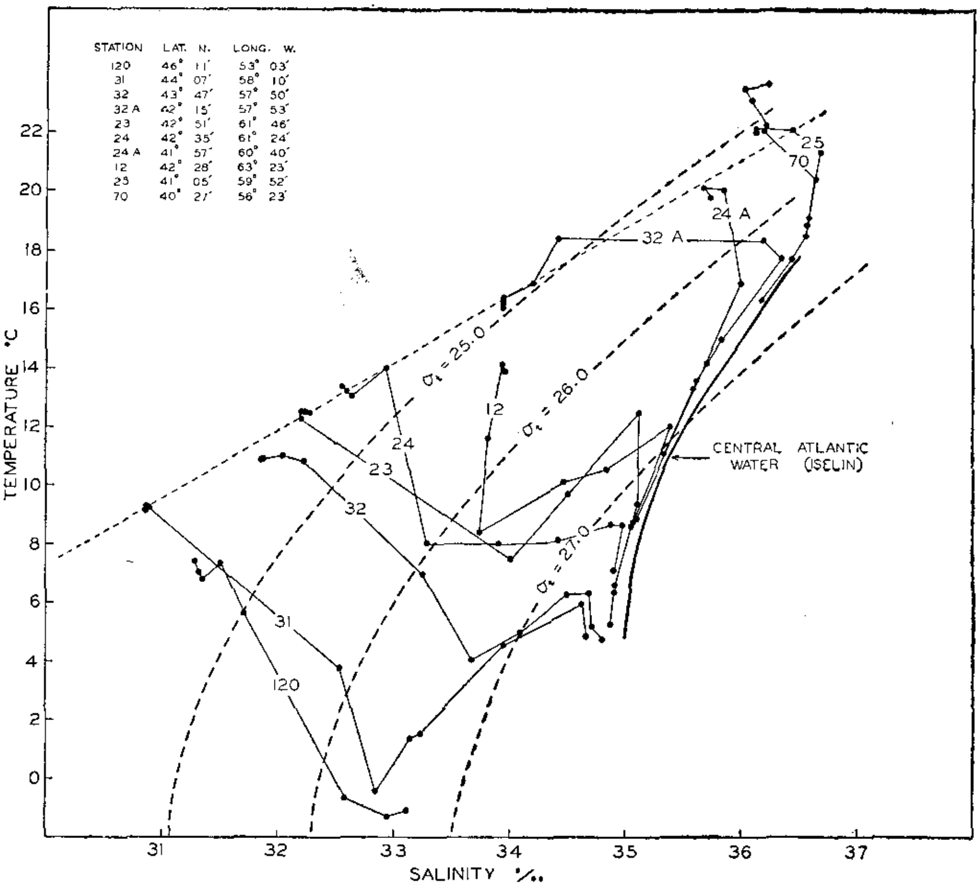


Fig. 21. Temperature-salinity relationships for the selected stations occupied in the slope water off the Scotian Shelf in November, 1951.

the Gulf Stream (Station 25), indicating that direct mixing of these waters takes place. The observations at 30 metres at stations 24 and 32A fell on the same line, and each was warmer and more saline than the corresponding surface waters. This shows that some of the water formed by surface mixing sinks and flows shoreward under the lighter, more coastal waters. A definite bend in the T-S curves of all but the most seaward stations indicates the influence of the cold water layer.

The deep waters of all stations have T-S characteristics which fall on a smooth curve that nearly parallels Iselin's curve for Central Atlantic water. The depth at which these water types occur, however, is much less than that at which similar types are found in the Sargasso Sea. Water closely resembling that found at 1000 metres in the Sargasso Sea is found at 400 metres in the slope water. Iselin (1936) has remarked upon the apparent upwelling and the fact that water of a given temperature is less saline in the slope water than in the Central Atlantic.

A straight line drawn for T = 1.3°C., S =

32.95‰, a point representative of the core of the cold water, to a point representative of similar depths in the Gulf Stream (19.0°C., 36.6‰) falls along a section of the deep water curve paralleling Iselin's Central Atlantic curve. Here, then, we have two types of water, with approximately the same density, efficiently mixed to produce waters of greater density which sink and flow under the lighter waters. The subsurface layers of the slope water must be formed in this way, and by a mixture of water so formed with Central Atlantic waters upwelled against the continental slope (McLellan, Lauzier and Bailey, 1953).

SEASONAL AND LONG TERM VARIATIONS IN TEMPERATURES

Seasonal Variations.

The annual variations in temperature and of which the seasonal variations are a part, are controlled at the surface by a large number of factors, chief of which are radiation, the character of the currents and the prevailing winds. In the surface layers the temperature variations, as pointed out by Sverdrup, Johnson and Fleming (1949) are due to the variation in the amount of heat that is absorbed at different depths, to the effect of heat conduction, to variations in the currents related to lateral displacement of water masses, and to the effect of vertical motion.

In the northwestern-north Atlantic, there are too few observations for the production of curves showing annual variations, and the calculation of annual means such as were produced from shore station data (Hachey and McLellan, 1948, and Bailey, MacGregor and Hachey, 1954). However, Smith, Soule and Mosby, (1937), used observations from various cruises to discuss the annual cycles for the various sectors.

In the West Greenland sector, observations of Smith, Soule and Mosby (1937), show that throughout the year, cold-low salinity water (East Greenland-Arctic) prevails in the surface layers next to the coast, while further offshore at deeper levels, warmer and saltier water persists, (Irminger-Atlantic). Although complete data are not available, it is probable that the temperature in the Irminger-Atlantic current off Cape Farvel rises from a minimum in February of about 4°C. to a maximum of slightly over 8°C. in September. In the fresher water near the coast, the temperature probably rises from -1.3°C. at the end of the winter to approximately 3° or 4°C. at the end of the summer.

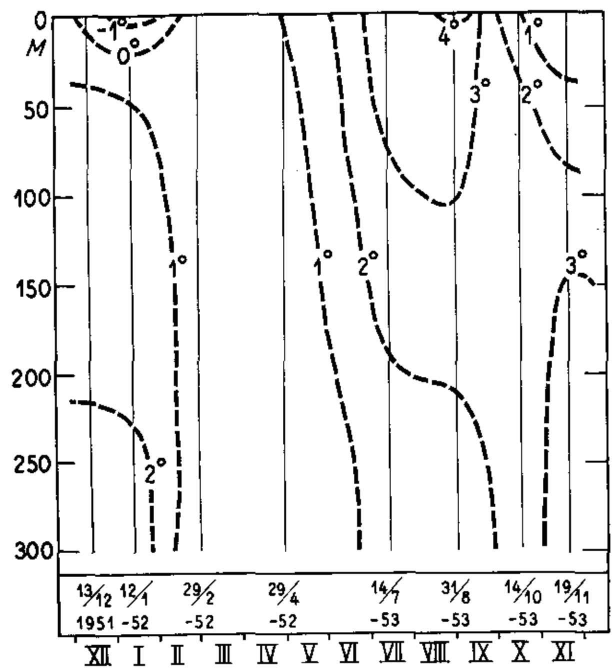


Fig. 22. Variations of temperature throughout the year in the entrance to Godthaab Fjord.

In the coastal area of West Greenland, a fixed hydrographic station at the entrance to the deep Godthåb Fjord has been occupied throughout the year. It is believed that many of the features in the variation of temperature at this station are representative of those for the middle part of the West Greenland area. Figure 22 illustrates the variation of temperature throughout the year. The most striking feature of the diagram is that the bottom water is shown to reach its maximum temperature as late as November, and at times as late as January. The warm bottom water must originate from the Irminger Current and it appears to reach its maximum temperature in late autumn or early winter in the latitude of Godthåb. Observations from Julianehåb Bay indicate that similar conditions prevail there. At this locality the maximum bottom temperature seems to be reached approximately two months earlier than at Godthåb Fjord. The explanation is presumably that the water mass of the Irminger Current reaches its maximum temperature in the Irminger Sea in August, but it is late autumn or early winter before this water arrives in the vicinity of Godthåb Fjord. Since this current is deep off West Greenland, it is not subjected to heat exchanges with the atmosphere and thus undergoes only slight temperature changes in its northward progress.

During late February the density of the winter cooled surface layer increases to the point where vertical mixing through convectional overturn reaches to the bottom. At this time, the warm bottom waters disappear from the entrance to Godthåb Fjord, but in the inner part of the fjord, where the stability of the surface layer is very great, the warm bottom water persists throughout the spring and summer months.

The inflow of warm water to West Greenland during the winter time is one of great importance to the fish population in that area, since it offsets the effects of winter chilling and prevents sea temperatures from falling below a critical value.

On the middle West Greenland Banks (Fiskenæs Bank, Fylla Bank and Lille Hellefiske Bank), the yearly variations are broadly as follows: In early spring the water over the shallowest part of the banks (less than 50 metres) is, in general, cooled to approximately 0°C., but off the western edge of the banks where the water is influenced by the Irminger Current, the temperature is nearly always positive below 50 metres. During the spring and summer months, the upper layers are heated by radiation from the sun. At the same time, the intensity of the Arctic component of the West Greenland Current increases, and this often causes decreasing temperatures over the deeper parts of the banks during June and July. On the western edge of Fylla Bank in depths of about 100 metres, the year's minimum temperature will often occur as late as June or July. During August and September temperatures rise at all depths, and observations from the entrance to Godthåb Fjord make it seem probable that the temperature in the Irminger Current on the western slope of the banks continues to rise during the autumn and early winter.

On Store Hellefiske Bank the oceanographic conditions differ somewhat from those described above. The West Greenland Current here is weak, so that temperature conditions are more dependant upon local meteorological conditions than upon the sources of the water masses. As a result summer temperatures are often higher on this bank than on the more southerly ones.

The main feature of the variations in salinity is the decrease in the salinity of the surface layers during the summer months due to land drainage. In the deeper layers, the salinity of the Irminger Current increases during the late summer and autumn, and will often reach $35^{\circ}/_{oo}$ off the western slope of Fylla Bank during August and September.

In the areas north of the Grand Banks there are not sufficient data to make any real estimates of seasonal variations of temperature. It is reasonable to assume that winter chilling cools the surface waters to the freezing point (a function of salinity) and that the temperatures reach those shown in Figure 12. In the Grand Banks sector, it has been generally considered that the Labrador Current exhibits freshet conditions during the spring, through the release of water from melting snow and ice, and then dwindles or disappears from the Grand Banks region in the fall and winter. When the distances involved and the rate of transport are considered, however, a flood wave comes much too early to be associated with summer or even vernal warming. The seasonal range in the minimum temperature of Labrador Current may be wide. Temperatures as low as -1.6°C. have been recorded at a depth of 100 metres, and as high as 2.9°C. along the east side of the Grand Banks. Fluctuations in this area are of such magnitude that the determination of an annual cycle is practically impossible.

On the Scotian Shelf, Hachey (1942), used data taken in 1938 to illustrate the oceanographic conditions at various seasons in that area.

In winter, temperatures at the surface are as low as -0.5°C., and at the bottom as high as 7.9°C. Salinities at the surface are as low as $30.55^{\circ}/_{oo}$ and at the bottom as high as $34.45^{\circ}/_{oo}$. East of Sable Island Bank, at all depths, low temperatures (less than 2°C.) and low salinities (less than $32.9^{\circ}/_{oo}$) prevail.

In spring, vernal warming is shown by a warm surface layer extending even to 74 m. depths and distinct from the subjacent "intermediate cold water layer". Penetration of warmer water at

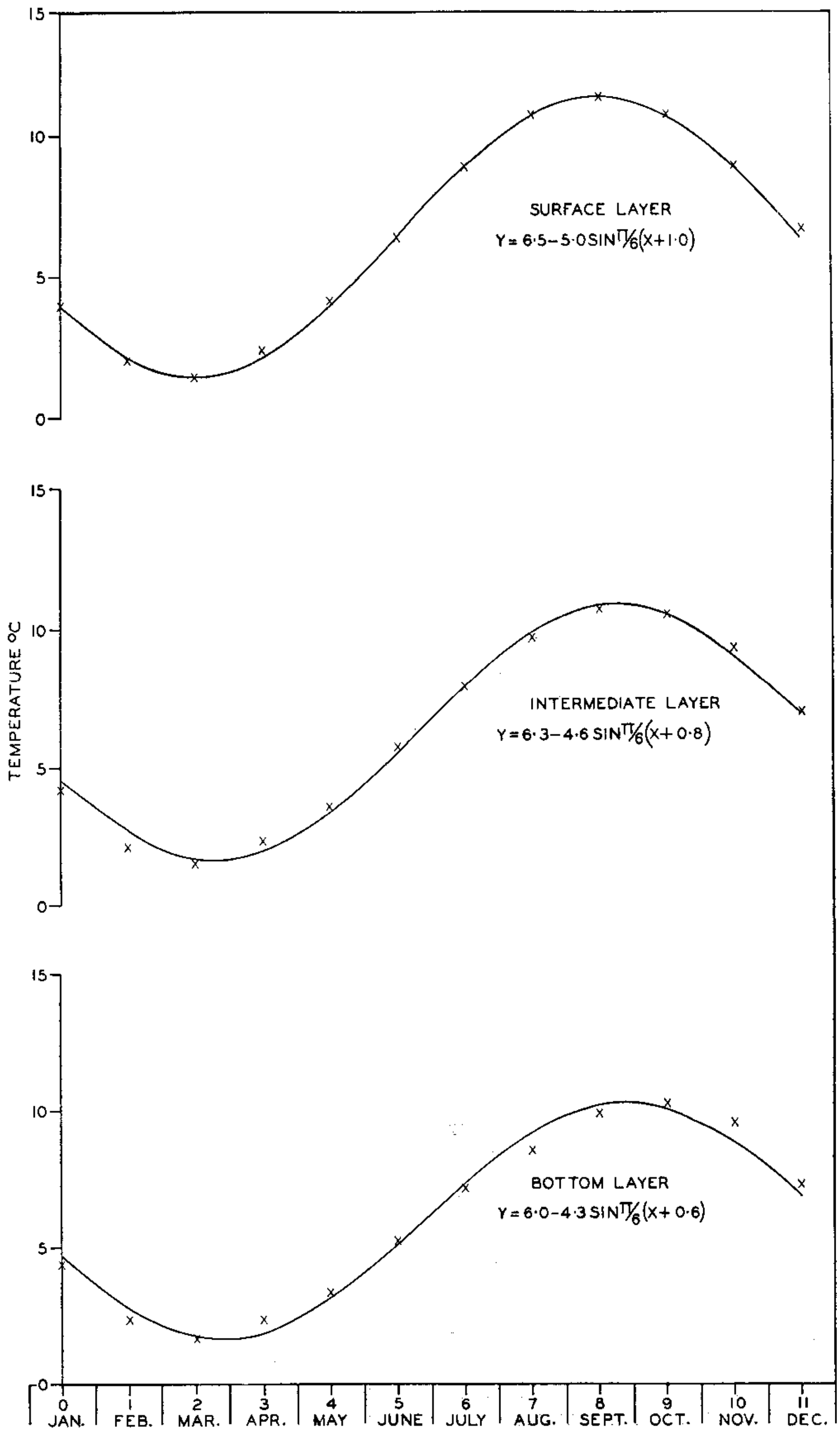


Fig. 23. Normal theoretical temperature-time curves for the surface, intermediate, and bottom layers at Station 5 (Bay of Fundy).

the bottom in certain areas is shown by temperatures as much as 2° higher than in winter.

In summer, there is full development of the warm "upper layer", and complete isolation of the cold "intermediate layer".

Inshore waters on the Scotian Shelf have temperatures at the surface, ranging from -1.0° to 19.8°C., and at 50 m. as low as -1.6°C. The winter minimum generally occurs in February and the summer maximum usually in September.

At several selected points along the Canadian Atlantic coast, Hachey (1939) demonstrated how the normal monthly mean surface water temperatures vary from season to season. Following the same general method, Bailey, MacGregor and Hachey (1954) determined the equation for the normal mean monthly temperature curves for three different layers in the Bay of Fundy, as shown in Figure 23.

The salient features are presented in Table III.

TABLE III.

Important features of the normal mean monthly temperature curves in three layers in the Bay of Fundy.

	Max.	Date	Min.	Date	Annual Mean
Surface Layer	11.46°C.	15 Sept.	1.47°C.	15 Mar.	6.51°C.
Intermediate Layer	10.74	30 Sept.	1.54	15 Mar.	6.29
Bottom Layer	10.26	15 Oct.	1.67	15 Mar.	6.01

An interesting fact is that the maxima occur approximately two weeks later in successive layers. As the warming of the deeper waters is carried out almost wholly by vertical mixing this time-lag is a measure of the efficiency of the tidal mixing in the area. In the case of the minima, on the other hand, autumnal and winter cooling reduces the stability of the upper layers so that more efficient mixing takes place, and the time-lag is practically zero.

The T-S cycle (Figure 24), was used by Bailey, MacGregor and Hachey (1954) to demonstrate the annual cycle in the oceanographic conditions in the Bay of Fundy. It is seen that the entire cycle of oceanographic conditions at station 5 (Lat. 44°56′N., Long. 66°48′W.) are clearly defined by the T-S diagram. The normal values of temperature and salinity are illustrated as well as the normal annual ranges. Bailey (1953) has illustrated the T-S cycle for 1952. In viewing Figure 24, it may be noted that between the middle of February and the middle of March, there is normally a short period when the water column at station 5 is nearly homogeneous with respect to temperature and salinity. The surface layer continues to cool while the bottom layer begins to warm. Until April-May with the land drainage at a maximum, the salinity in each layer becomes progressively fresher with the greatest changes taking place in the surface layer. By the middle of April, the waters are isothermal, but show a considerable difference in salinity. In May, minimum salinities are reached in all layers while temperatures are increasing. During the summer months both temperatures and salinities increase in value. In the bottom layer an increase of salinity in August causes a very noticeable change in the trend of the T-S cycle. This would seem to be related to a phenomenon of replacements of Bay of Fundy waters. Undoubtedly, a steady replacement is involved during the time of "spring freshets", but vertical mixing on a large scale almost completely masks the possible increases in salinity that such replacements might entail. Thus it is not until the middle of summer when the river discharges are at a minimum that replacements can be readily detected. By the middle of September, the temperature has reached a maximum in the surface layer, and the salinities have increased in all layers. The maximum temperature of the surface layer is reached in late September, and by the end of the month all layers have reached their highest salinities. Temperatures and salinities decrease at a comparatively steady rate in all layers until February, when the waters have nearly the same characteristics from top to bottom.

Long Term Variations.

It is generally contended by a large number of authors that the present upswing of water

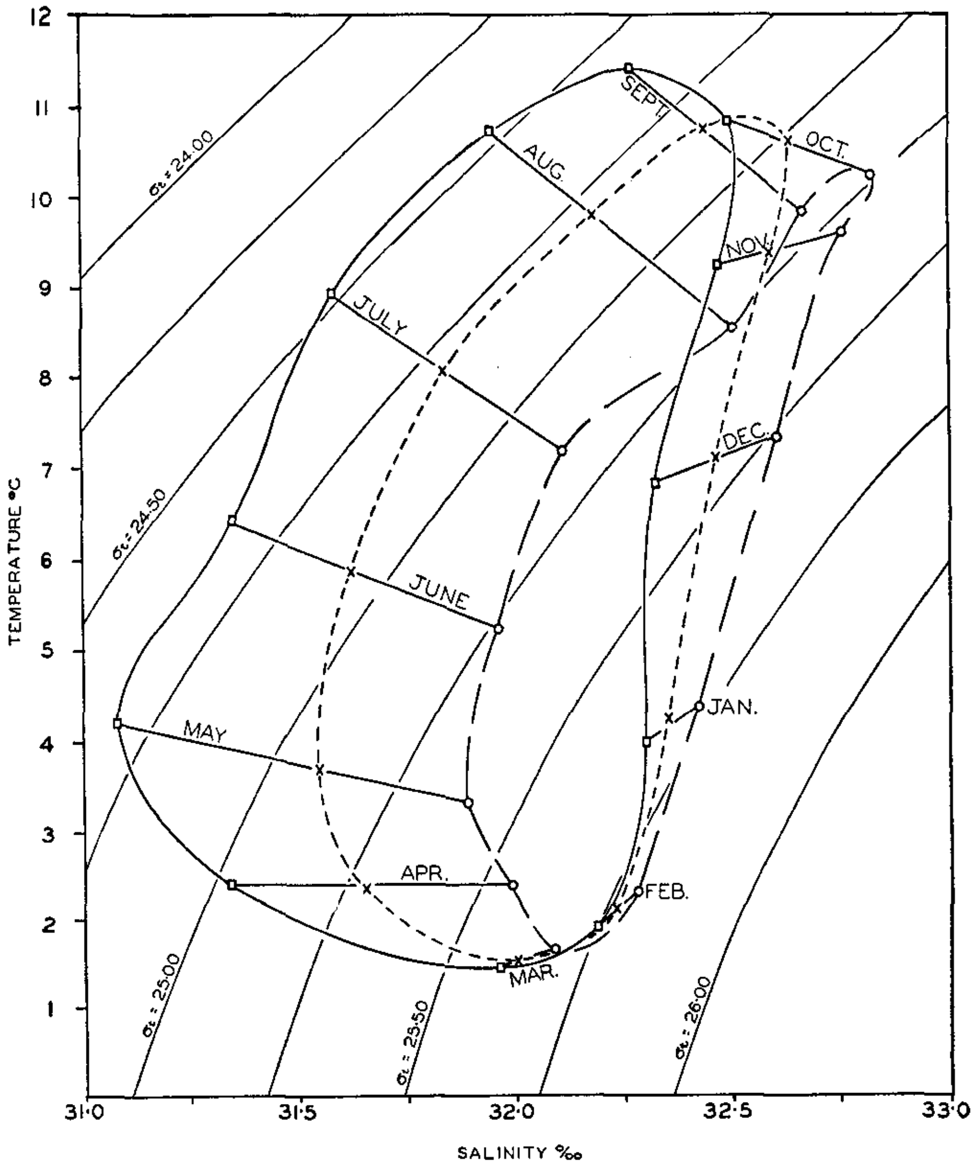


Fig. 24. Normal T-S curves for each month, and the normal annual T-S cycle for the surface————, intermediate - - - - - - -, and bottom—— —— —— layers, Station 5 (Bay of Fundy).

temperatures in the North Atlantic is mainly due to an increased atmospheric circulation. A great many papers have been written regarding one phase or another of this subject and its attendant climatic changes, and its effects upon the distribution of plants and animals in various regions. Regular oceanographic observations were few in West Greenland waters before the middle of the 1920's, but the Danish Meteorological Institute has collected, since 1876, surface temperature observations taken along shipping routes, and presented them as monthly means for each degree square. These data have been worked up by Smed (1953), who calculated the monthly anomalies for each one degree square and averaged the greater areas. Shown in Figure 25 are the five-year running means of these anomalies for Smed's areas A and B. The figure shows the very pronounced increase in temperature in the 1920's. From the 1930's, the temperature has decreased somewhat, and at present is about 0.5°C. lower than during the maximum. While the direct temperature measurements at sea are only available for the past 65 years, infor-

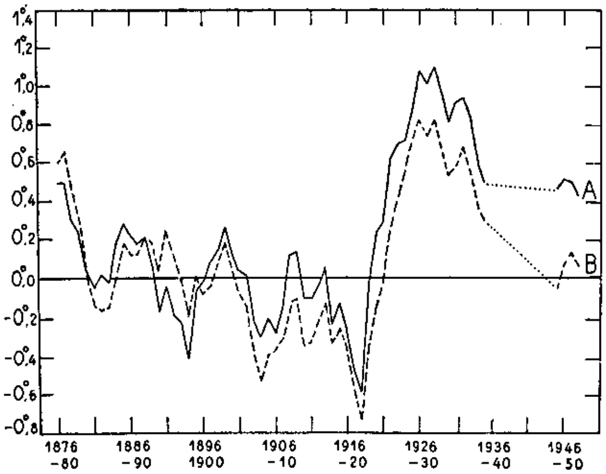


Fig. 25. Five year running means of surface temperature anomalies for Smed's (1953) areas A (W. Greenland) and B (S. Greenland).

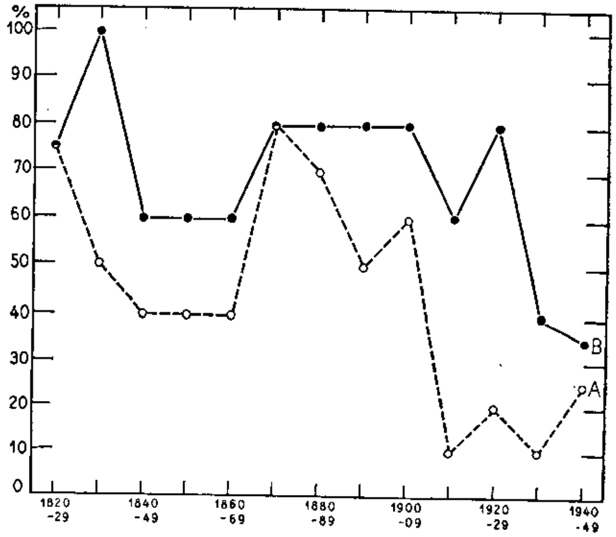


Fig. 26. Frequency of years in which the Storis reached as far north as Godthaab (A. broken line) and to Fiskenaesset (B. full line).

mation on the amount of Storis carried by the East Greenland Polar Current to West Greenland, is available for a considerably longer period. Speerschneider (1931) collected information about Storis from log books of ships visiting Greenland. Since there is assumed to be a close connection between sea temperatures and the presence of Storis a consideration of the ice conditions can be expected to yield information concerning climatic changes at sea.

Speerschneider (1931) produced in tabular form a summary of the maximum extension of ice along the west coast of Greenland. On the basis of this summary and from information provided by the Danish Meteorological Institute for the years 1930-1949, Figure 26 has been drawn. The points for curve A give, for each decade, the percentage of the years in which Storis reached at least to the neighbourhood of Godthåb during its maximum extension northward, while those for curve B give the analogous curve for Fiskenæsset. The most striking feature in curve A is the strong decrease in the frequency of Storis off Godthåb from about 1910, and off Fiskenæsset from about 1930. Furthermore, there seems to have been a period with relatively favourable ice conditions from about 1840 to 1870.

In the Bay of Fundy at St. Andrews, N. B., temperatures have been observed twice daily since 1921. Analyses of these data were made by Hachey and McLellan (1948), in which they compared the annual mean surface water temperature at St. Andrews with various points along the Atlantic coast of North America. From this comparison, it is obvious that the outstanding cycles and trends as exhibited in the surface water temperatures at St. Andrews, N. B. are common to the waters of the main portion of the Atlantic coast of North America. With this in mind, it is of interest to regard an analysis of the St. Andrews data on the basis of a twelve-month running average. The period shown in Figure 2 is from 1940 to 1951. Lauzier (1952) compared the same curve, but from 1936-1951, with comparable data from Sambro Light Vessel (off Halifax, N. S.). During the 1940-50 decade, the waters of the Bay of Fundy have shown a fairly definite trend towards higher temperatures with a maximum in 1949, with an appreciable decrease in 1950. From 1921 to 1940, the variations of the running average ranged between 59.0 and 91.8 degree-months, and since 1941 maxima of 93.5, 100.9 and 103.4 were reached in 1947, 1949 and 1951, respectively. The minimum recorded in 1950 was a high as the maximum of the 1921-1940 period.

Temperature and salinity relationships in Hudson Bay in 1930 (Hachey, 1931) show the characteristic Arctic water exhibiting temperatures between -0.5° and -1.8°C. and salinities between 31.0 and 33.2‰, while in 1948 (Bailey and Hachey, 1950), temperatures were between

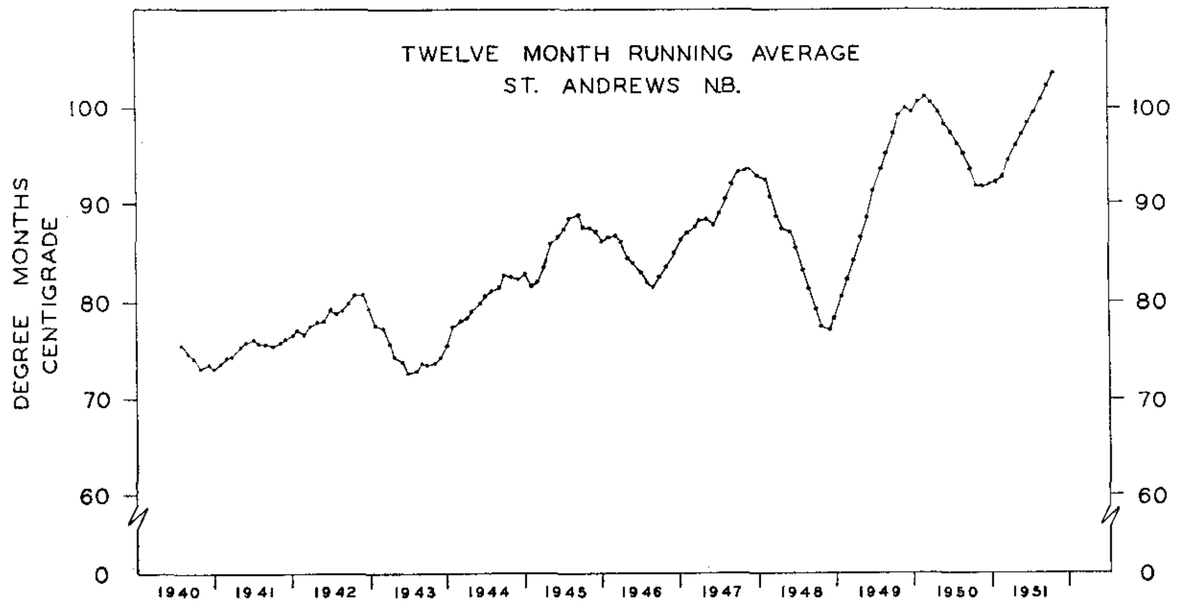


Fig. 27. **Twelve months running average of surface water temperatures at St. Andrews, N. B., 1940-51.**

-0.5 and -1.5°C., and salinities between 32.0 and 33.5°/₀₀. In addition, it may be pointed out that salinities greater than 32.8°/₀₀ were found only at one station in 1930, but at all stations in 1948. Oceanographic conditions at the 175 metre level in the Bay of Fundy, for a period from 1916 to 1918, were compared with those for a period from 1950-1952 by Bailey (1953). The data showed an increase in temperature of approximately 3.2°C. and an increase in salinity of approximately 0.5°/₀₀. These differences as observed in Hudson Bay and Bay of Fundy reflect an increased Atlantic influence in the deeper waters off the Canadian Atlantic coast.

It is of interest to note that in Greenland waters, 1949 was considered a "cold year" while data from St. Andrews represented it as a "warm year". On the other hand, 1947 was represented as a warm one in both areas. These observations would seem to suggest that although an increasing Atlantic influence has been generally experienced, the local reactions differ to some extent.

INFLUENCE OF TEMPERATURE ON THE FISH POPULATION

Temperatures and Occurrence of Cod.

The occurrence of cod in West Greenland waters has been periodical and the two best known rich cod periods are 1845-1851, and 1924 to the present.

Figure 26 shows that during both periods relatively favourable ice conditions prevailed. These rich cod periods seem thus to coincide with periods of relatively warm climate. This becomes still more apparent when the beginning of the latest cod period is compared with the surface temperatures shown in Figure 25. At the same time as the temperature rose sharply, the cod, previously a rare fish, appeared in greater and greater quantities and over larger and larger areas off the West Greenland coast. Therefore the cod fishery developed into the main source of income for the Greenlanders. In this connection, it is alarming that the temperatures since the late 1930's have shown a decreasing trend, especially in the south Greenland area. A continued decrease may cause a catastrophy in the Greenland cod population.

During the various seasons of the year, the occurrence of cod is greatly influenced by variations in the oceanographic conditions. In the spring, when the water on the shallow parts of the banks is cold after winter chilling, the cod are found mainly in the deeper waters and especially over the western edge of the banks, where the warm Irminger Current is dominant. During June and July, when the deeper parts of the banks are often covered by colder water, the cod will migrate to the shallow parts of the banks or they will be found pelagic in the upper 50 metres in the summer warmed layers.

Different attempts have been made to determine the most profitable fishing temperatures. Birger Rasmussen (1952) states that profitable fishing for cod cannot be expected if bottom temperatures are below 2°C. Faroe line fishermen, who used reversing thermometers, have reported that practically no fish were caught when the temperature of the water was below 0°C., and that the fishery first became profitable at temperatures above 1°C. Birger Rasmussen (1953) found in 1952 that the greatest concentration of pelagic cod occurred in temperatures between 3° and 4°C.[1]). During the English Investigations in the Cape Farewell region in 1952, G.C. Trout (1953) found that the bulk of the cod stay in water with temperatures less than 2°C.

The relationship between occurrence of cod in Greenland waters and sea temperature seems to be rather complicated. The cod seem to avoid water with temperature below 1°C., but at higher temperatures the concentration of cod is probably more related to food concentration than to water temperature.

Influence of Temperature on the Strength of Cod Year-Classes.

Since the cod in Greenlandic waters is living near its northern limit, it is reasonable to try to explain the great fluctuations in the strength of the year-classes in terms of fluctuations in the temperature conditions during the larval stage. The cod spawn in spring on the western part of the mid Greenland Banks and in June, the larvae are found mainly over Fylla and Lille Hellefiske Banks. In the following, bottom temperatures in June for the shallow part of Fylla Bank (40 m.) are given for all years in which observations are available, and arranged in order of increasing temperatures:

Year:	1938	08	49	09	25	48	37	28	50	36	26	24	34	47
Date:	1	24	2	8	13	10	2	18	10	10	12	24	24	13
Temp:	0.04	0.10	0.10	0.60	0.80	(1.30)	1.34	1.77	(1.8)	1.95	2.21	2.50	2.69	3.39

The temperatures in brackets are interpolated from neighbouring months and must be treated with reserve.

It is remarkable that the years with the highest temperatures (1936, 1926, 1924, 1934, and 1947) coincide with the best year-classes of cod in recent years, while only unimportant year-classes have arisen from the colder years. Figure 28 shows the yield up to 1946 of cod year-classes in the Greenland fishery as a function of temperature over Fylla Bank in June. The close connection between the strength of the year-class and the temperature makes it probable that the variations in temperature during the larval stage are the principal cause of the fluctuations in the strength of the various year-classes.

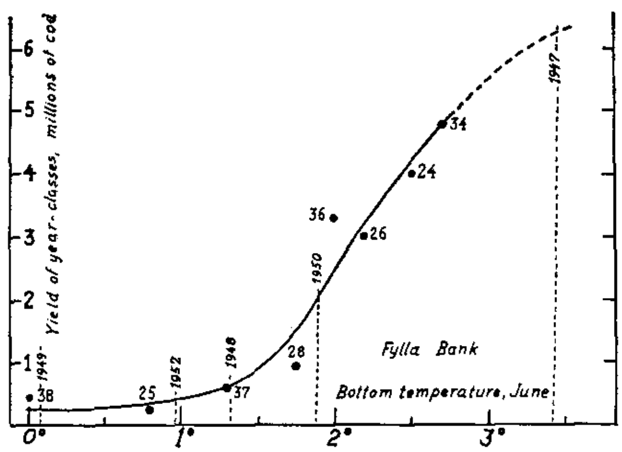


Fig. 28. Yield of cod year-classes in the Greenland fishery as a function of temperature over Fylla Bank in June. Temperatures of recent years are marked by dotted lines.

1) According to the Norwegian Research Report for 1953 (see p. 42) the greatest concentrations of pelagic cod were in that year observed in water layers with temperatures between 2.2 and 2.4°C. (Ex. Secr.).

NUTRIENT SUPPLY

The production of organic matter by phytoplankton depends to a very high degree on the presence of nutrient salts in the surface layers of the seas. In the deeper layers, below 100 metres, these salts are always present in sufficient concentration, but in the surface layer they are frequently almost completely removed by the phytoplankton. In Greenland waters, as in most waters of high latitude, winter cooling of the surface layers will cause a vertical mixing which brings the deeper layers, rich in nutrient salts, to the surface. In the spring, when the water is stabilized, a rich plankton production sets in, and the nutrients are removed except in areas where vertical mixing brings a continuing supply of the nutrient-rich deep water to the surface. The distribution of inorganic phosphate at 20 metres in July, 1953, for West Greenland waters is shown in Figure 29. The highest concentrations are met off the Greenland slope where the turbulence in the strong current causes vertical mixing. Good conditions for plankton production on the western side of the banks should prevail as far north as Fylla Bank. In the northern part of the area only small concentrations of phosphate are encountered, with the exception of the westernmost station in the Davis Strait section, which lies near the edge of the Canadian Polar Current.

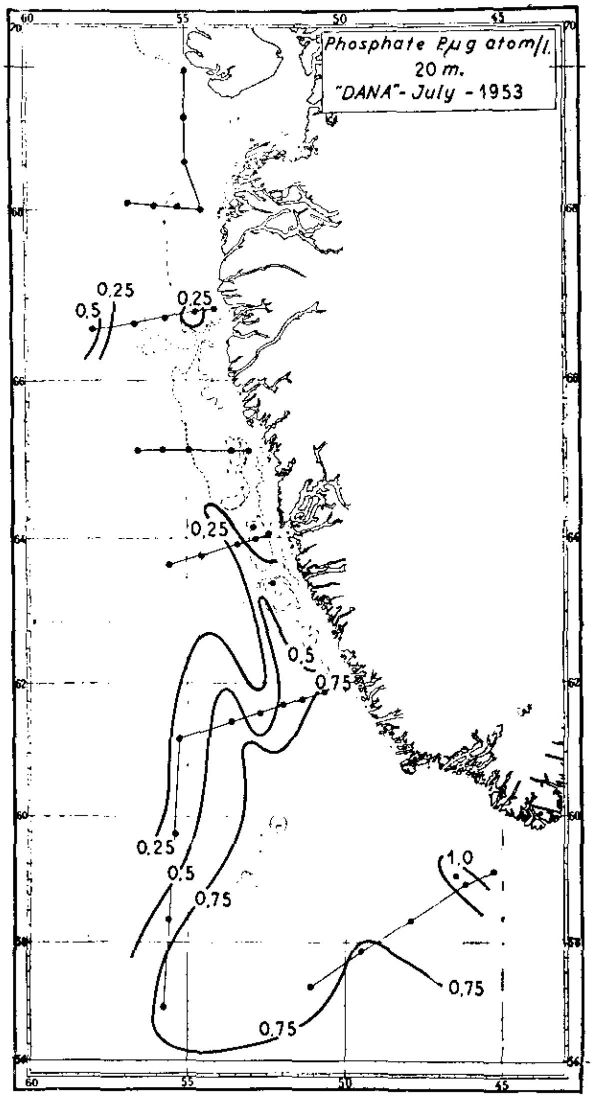


Fig. 29. Distribution of Phosphate off West Greenland, July, 1953.

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