THE MARION EXPEDITION
TO
DAVIS STRAIT AND BAFFIN BAY
1928

SCIENTIFIC RESULTS
PART 3

U. S. TREASURY DEPARTMENT
COAST GUARD

Bulletin No. 19

THE MARION EXPEDITION
TO
DAVIS STRAIT AND BAFFIN BAY

UNDER DIRECTION OF
THE UNITED STATES COAST GUARD

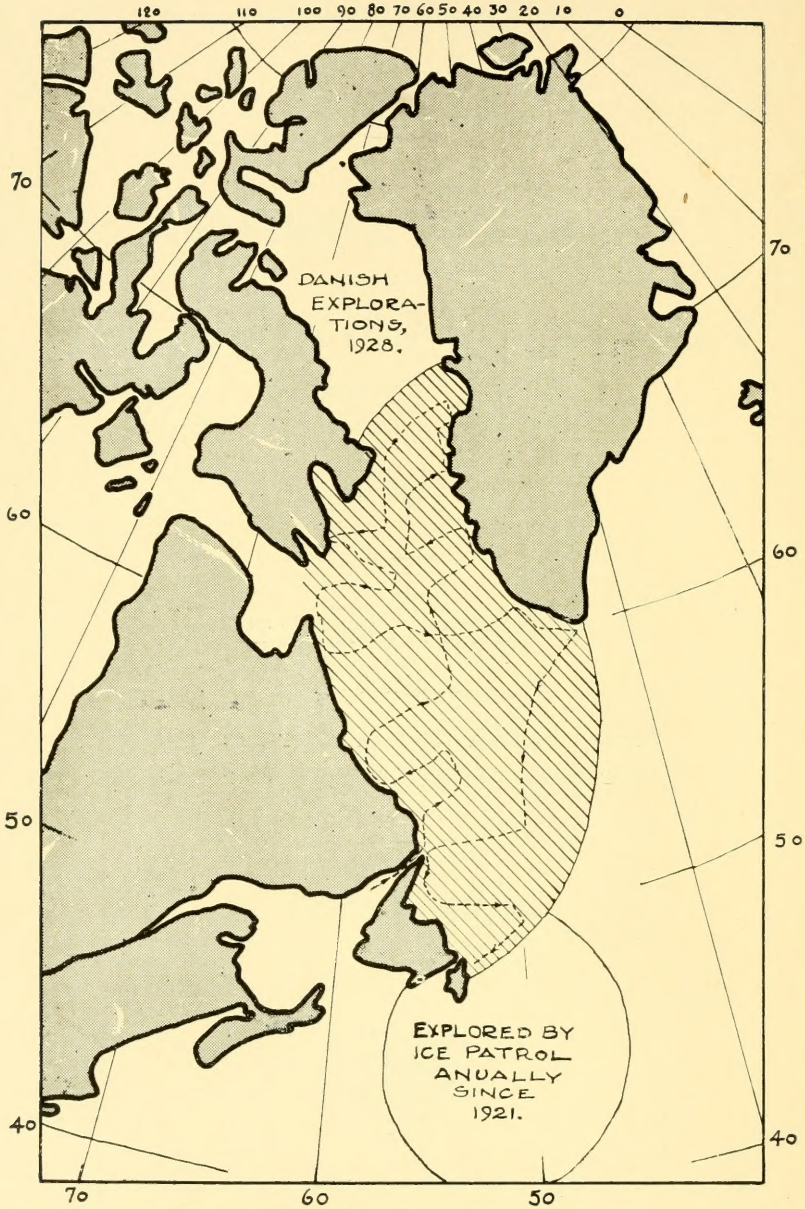
1928

SCIENTIFIC RESULTS
PART 3

Arctic Ice, with Especial Reference to its Distribution to the North Atlantic Ocean

EDWARD H. SMITH

UNITED STATES
GOVERNMENT PRINTING OFFICE
WASHINGTON : 1931



THE ICE AND HYDROGRAPHICAL SURVEY OF DAVIS STRAIT BY THE MARION EXPEDITION

FIGURE 1.—Baffin Bay, Davis Strait, and the waters around the Grand Banks embrace the iceberg regions of the western North Atlantic. The waters surveyed by the *Marion* expedition in 1928 form a connecting link in the history of icebergs from the time they leave the glaciers on the west side of Baffin Bay until they finally melt in the Gulf Stream south of Newfoundland. The track of the *Marion* from the time of leaving the Strait of Belle Isle until the arrival at St. Johns, Newfoundland, is shown on the shaded area.

ERRATA

The following changes have been found necessary since the publishing of this volume:

Figure 1.—Change the word "west" in the fourth line of legend to read "east."

Page 2.—Change the word "Joseph" in the twenty-sixth line to read "Josef."

Page 17.—Add the word "North" after the figures "85°" in the fourth line of the tenth paragraph.

Page 19.—Figure 9; add the word "erroneously" after the word "sometimes" in the legend.

Page 25.—Figure 13; change the name "Ellesworth" in legend to read "Ellsworth."

Page 38.—Change name "Angmagissalik" in lines 14 and 25 to read "Angmagssalik."

Page 55.—Figure 28; change the word "floes" in the last line of legend to read "glacons."

Page 64.—Figure 33; change the words "north side of Bennett Island" in second line of caption to read "south side of Bennett Island."

Page 87.—Change the word "as" in seventh line to read "so."

Page 93.—Change the word "water" in the third line of the legend to Figure 54 to read "reader."

Page 152.—Change the name "Placentia Bay" in tenth line to read "the Miquelon Islands."

Page 169.—Change the spelling of word "seperate" in forty-seventh line to "separate."

Page 170.—Change the word "sunk" in the last paragraph to read "sank."

Page 185.—Subparagraph (c); change the words "November to April" to read "December to March."

Subparagraph (d); change the words "August to January" to read "October to January."

Page 188.—Change the name "Julianahaab" in twenty-ninth line to read "Julianehaab."

Change the equations c' and d to read as follows:

$$c' = \frac{6(2\times\text{Dec.}+2\times\text{Jan.}+\text{Feb.}+\text{Mar.})^{\text{J-BI}}+(2\times\text{Dec.}+2\times\text{Jan.}+\text{Feb.}+\text{Mar.})^{\text{I}}}{25}$$

$$d = \frac{(\text{Oct.}^{\text{J-B}}+\text{Nov.}^{\text{I-B}}+2\times\text{Dec.}^{\text{I-B}}+\text{Jan.}^{\text{I-B}})}{5}$$

ERRATA

The following changes have been found necessary since the publishing of this volume:

Figure 1.—Change the word "west" in the South line of legend to read "east."

Page 2.—Change the word "Joseph" in the twenty-sixth line to read "Josef."

Page 17.—Add the word "North" after the figures "85°" in the fourth line of the tenth paragraph.

Page 20.—Figure 9: add the word "occasionally" after the word "sometimes" in the legend.

Page 25.—Figure 18: change the name "Ellsworth" in legend to read "Ellsworth."

Page 28.—Change name "Angmagssalik" in lines 14 and 25 to read "Angmagssalik."

Page 55.—Figure 22: change the word "floes" in the last line of legend to read "glaciers."

Page 64.—Figure 32: change the words "south side of Bennett Island" in second line of caption to read "south side of Bennett Island."

Page 87.—Change the word "as" in seventh line to read "so."

Page 98.—Change the word "water" in the third line of the legend to Figure 34 to read "reader."

Page 129.—Change the name "Pleasantia Bay" in tenth line to read "the Shipmelon Islands."

Page 169.—Change the spelling of word "separate" in forty-seventh line to "separate."

Page 170.—Change the word "sunk" in the last paragraph to read "sunk."

Page 185.—Subparagraph (a): change the words "November to April" to read "December to March."

Subparagraph (d): change the words "August to January" to read "October to January."

Page 188.—Change the name "Julianshaab" in twenty-ninth line to read "Julianshaab."

Change the equations c' and d to read as follows:

$$c' = \frac{(\text{Oct.}_{1-24} + \text{Nov.}_{1-24}) \times \text{Dec.}_{1-31} + \text{Jan.}_{1-8}}{3}$$

$$d = \frac{6(2 \times \text{Dec.}_{1-31} + 2 \times \text{Jan.}_{1-31}) + (2 \times \text{Dec.}_{1-31} + 2 \times \text{Jan.}_{1-31}) + \text{Feb.}_{1-14}}{}$$

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PREFACE

Over each polar region of the earth lies a great cap of heavy ice. During winter it grows and builds, expanding to a maximum volume; spring and summer witnesses its melting and retreat to a permanent central core. Throughout the year, however, cold ocean currents from out of the north bear the fragmented ice border to lower latitudes. This paper embraces a study of Arctic ice from the freezing of the water and its accumulation of snow through the various stages of development and re-formation to its final melting and distribution to the North Atlantic Ocean. The present treatise is mainly prompted by the writer's interest in the ice as it constitutes a grave navigational menace to the North Atlantic Ocean.

Our knowledge regarding the regional distribution of the ice, its state, behavior, and paths of drift has been accumulated over a long period of years; first, through the early voyages of exploration as maritime countries searched and pressed on for new territorial possessions and trade enrichment. These masters and navigators, finding the northern regions so commonly inaccessible, mainly on account of the dangers and barriers of ice, naturally recorded their experiences, gradually building up a history of the subject. Many of the early explorers were also endowed with scientific attainments and a keen appreciation of the value of accumulating all possible knowledge on the ice problems of the Arctic. The development of science later led to combining voyages of exploration with those having scientific objects and finally, in the more recent pages of history, we find expeditions setting forth equipped with excellent laboratories and a highly trained scientific staff, solely to gather knowledge free from economic bias.

Legends and sagas of the Scandinavians and other northern people devote much space to the struggle with ice. Historical records commence with the early gropings of the British and the Dutch to the seas north of Europe, and closely on the heels of northeastern discoveries came the unfolding of the northwestern North Atlantic. The sea ice and icebergs drifting southward out of Baffin Bay and Davis Strait were phenomena known to sailors and men of science even before the Pilgrims landed at Plymouth. Even after the Revolutionary War of the American colonies European enterprise was still actively engaged in discovering a northwest passage and a short route to the Orient. The early attempts at finding a northwest passage to India were followed by fur trading and whale fishing activities in the West Greenland and Canadian Arctic sectors. Much of our knowledge regarding the sea ice of Baffin Bay and the American Arctic Archipelago has been contributed by the whale industries of the Dutch, the English, and the Americans which flourished over a period of two centuries. The northwestern section also experienced great activity during the middle of the nineteenth century as a result of the prolonged search for Sir John Franklin and his ill-fated

expedition. Interest during the following years, kindled by international rivalry centered on the discovery of the pole, and this activity brought Davis Strait and Baffin Bay world attention during the latter part of the nineteenth and the early part of the twentieth century, after the decline of the whaling fisheries.

We now come to a history of the scientific investigations which have been carried out in connection with the great procession of ice which invades the North Atlantic along its western side. Naturally, one of the most important pieces of information on the southward distribution is information of the current which transports the ice to lower latitudes. The material is based on the following expeditions: *Scotia*, 1913; *Michael Sars*, 1924; *Dana*, 1925; *Chance*, 1926; *Marion*, 1928; and *Godthaab*, 1928. In addition to these investigations, confined solely to the inaccessible waters in summer time, we have the records of the international ice patrol for a period of several years, 1913 to 1929, covering the ice season, March to July, in the regions south of Newfoundland. As a result of the work of these expeditions we now are able for the first time to construct a fairly accurate picture of the system of oceanic circulation and therefore the paths taken by Arctic ice from its region of formation to its ultimate fate in the North Atlantic.

The international ice patrol was established in 1913 as a result of the tragic loss of life and property when the steamship *Titanic* sank April, 1912, off the Grand Bank. Besides the practical work of scouting for ice and warning all approaching ships of its position and drift, there has been the scientific program of collecting observations which would lead to an intelligent knowledge of the best manner in which to cope with the ice problem. During the course of its 16 years of service the ice patrol has accumulated a vast store of knowledge regarding the behavior of Arctic ice at its southern—melting end. An account of the ice from its northern sources to Newfoundland, based upon similar scientific methods as carried out by the ice patrol, became a universal demand, resulting in the sending out of the *Marion* expedition. The *Marion*, during the summer of 1928, completed a current and ice survey of the entire waters of Davis Strait from Newfoundland northward to the seventieth parallel of latitude, and during the same months the Danish ship *Godthaab* carried out a similar reconnaissance of a large portion of Baffin Bay. We are now ready (historical records combined with the oceanographical and ice observations of the *Godthaab*, *Marion*, and international ice patrol), to present a comprehensive exposition of Arctic ice, its state, behavior, and distribution to the western North Atlantic.

The writer first became associated with ice work when he joined the ice patrol in 1920. He has been a member of the scientific staff ever since and participated in the ice patrol cruises every winter and spring in the ice regions off Newfoundland from 1920 to 1924, inclusive. During the off season his time has been spent in study and compilation of the annual published reports at Harvard University, working under the supervision of instructors in oceanography and meteorology. In order to apply certain technical methods of oceanography and meteorology to the North Atlantic ice problem, he spent one year at the Geophysical Institute, Bergen, Norway, and

a few months at the British Meteorological Office, London, England, returning to the ice regions for the seasons of 1926 and 1927. In 1928 he was the leader of the *Marion* expedition to Davis Strait and Baffin Bay.

Two of the principal problems upon which the ice patrol has been working in order to provide greater safety for lives and ships on the trans-Atlantic routes have been (*a*) the forecasting of the annual number of icebergs which are liable to drift south of Newfoundland during the year, and (*b*) the construction of hydrodynamic maps to trace the behavior of currents and ice south of Newfoundland as well as to predict to approaching vessels the general movement of the ice. Much progress of an encouraging nature has been made in iceberg forecasting as a result of the statistical work at Harvard University just mentioned and at the British Meteorological Office. Since the establishment of this service four successes out of five years have been the record. The value of this work is obvious if North Atlantic steamship navigators can be warned in February or March that a heavy ice year, such as 1912, is soon in store, or they can be assured that little danger will exist, such as characterized the year 1924. The latter problem of constructing frequent current maps has resulted in a greatly increased knowledge of the movements of ice and ocean currents around Newfoundland, subjects even 10 years ago considered rather hazily. The present practice of two weekly or monthly maps of the circulation furnishes the best available information to the ice-patrol ships.

A comprehensive and connected account of Arctic ice has only now been made possible as a result of the *Marion* expedition to Davis Strait the summer of 1928. Part 1 of the present bulletin is an exposition of the bathymetry of Davis Strait. Part 2 treats the physical oceanography of Davis Strait with especial reference to the circulation of those waters. The present section deals with the drift of the ice in the established currents out of Davis Strait into the North Atlantic.

During the course of the researches of the international ice patrol many problems of a scientific nature regarding ice and ocean currents have arisen from time to time, and are presented in this thesis as original contributions; among them are the following: What is the prevailing circulation in the north polar basin? Do pack ice and icebergs from east Greenland ever contribute to the supply to North American shores? What is the cause of "north water" in Baffin Bay? In what proportion does the pack ice from the Polar Basin, Baffin Bay, and Hudson Bay mix to supply the North Atlantic? What effect does Hudson Strait exert on the Labrador current and its freight of ice? What is the cause of the annual variation in pack ice limits in the North Atlantic? In what proportions do polar glacial lands contribute to the North Atlantic iceberg quota? Of all the tidewater glaciers in the Northern Hemisphere, which ones are iceberg producing? What is the density of icebergs and their proportions of mass flotation? In what proportions do the wind and current enter to control the drift of the berg? What is the normal quantity of pack ice which melts annually, and what is the normal number of icebergs drifting south of Newfoundland? Is the Labrador current subject to a seasonal variation, and if so, what are the causa-

tive factors? What proportions quantitatively of the North Atlantic are cooled by northern ice and by cold water, and which of these agents is the more effective? Is the effect of ice melting in northern waters sufficient to account for the main system of oceanic circulation? Does the Labrador current extend down the east coast of the United States?

This affords me an opportunity to recognize the helpfulness of the following in connection with the compilation and treatment of the subject matter: Dr. Henry B. Bigelow, of Harvard University; Dr. C. E. P. Brooks, of the British Meteorological Office. Those in the Coast Guard to whom I am indebted are Lt. Comdr. N. G. Ricketts, Mr. Olav Mosby, and H. Addington.

THE MARION EXPEDITION TO DAVIS STRAIT AND BAFFIN BAY

ARCTIC ICE—WITH REFERENCE TO ITS DISTRIBUTION TO THE NORTH ATLANTIC OCEAN

EDWARD H. SMITH

Spread over the top of the world is a huge white covering of sea ice about 3,500,000 square miles in area, which during winter expands outward until at maximum, 27 per cent of the North Atlantic (the polar sea included), is ice decked. Add to this total 805,650 square miles of land areas, which in many places are submerged by several thousand feet of solid ice and we begin to realize the great extensions of this solid state of water.

The great polar ice cap covering 1,800,000 square miles of the sea never melts. During winter the basin becomes greatly congested through freezing and rafting of the fields and little open water is to be found. Lanes of open water appear along the coast for a few brief weeks in summer, and leads and pools appear throughout the pack. Two great icy arms extend from the central polar core, one along the east side of Greenland, the other down the American shore. Every year hundreds of square miles of ice fields perform journeys 1,500 miles or more in length, projecting barriers halfway from the pole to the Equator.

The annual discharge and summer melting amounts to about 1,100,000 square miles of northern sea ice. The great chilling effect attending the latent heat of melting combined with the cooling of northern currents are two of the Arctic's major geophysical factors. Annual variations in the amount of ice and cold water discharged into the North Atlantic are known to be a significant control of European weather.

The berg, of all ice forms in the polar regions, is the most spectacular. Greenland produces somewhere between 10,000 and 15,000 icebergs every year. Literally thousands of bergs are scattered at times in Greenland coastal waters.

The wanderings of the bergs are freer than those of the sea ice because, with their massive proportions, they sometimes survive long journeys. The icebergs that achieve farthest south are those traveling the 1,500-mile pathway down the Labrador coast.

Arctic ice, particularly the stream past Newfoundland, penetrates deeper into the North Atlantic, due partly to the shape of the latter, than into any other ocean. This invasion, moreover, for about four months of spring and summer creates a distinct menace to the main arteries of commerce on the most frequently traveled ocean of the world. The fact that the regions embracing the largest number of bergs are those enveloped in fog a large percentage of spring and summer, greatly accentuates the danger from ice.

HISTORICAL SURVEY

It is impossible to state who were the first peoples to direct their ships northward into the frozen seas, but Norse legends and sagas as early as the tenth century credit their viking navigators with several such adventures. The first historical records are of a voyage of discovery to Davis Strait in the year 1500, under the joint leadership of Scandinavians and Portuguese. Willoughby's voyage to the sea north of Europe, and Frobisher's to Baffin Land, then followed heading a list of Arctic explorations that is still continuing

The central polar cap of sea ice has been deeply penetrated by but few ships or men. Since this region is the site of and maximum development for all sea-ice forms, it has attracted the interest and attention of many explorers and scientists. Our knowledge of the polar cap is largely due to the following: Hayes, Hall, Weyprecht and Payer, Scoresby, Kane, Bering, Nordenskiold, DeLong, Nansen, Cagni, Peary, Toll, Kolchak, Stefansson, Bartlett, MacMillan, Storkerson, Vilkitski, Amundsen, Nobile, Byrd, and Wilkins. Bering in three voyages, 1728–1741, explored northward from the Pacific, discovering the strait bearing his name and surveyed the land of either continent, in the open water of the polar sea. The Scoresbys, Scotch whalers, in 1806 determined to see how far north they could go and forced their ship to latitude 81° 30′, in the longitude of Spitsbergen. Weyprecht and Payer, Austrian naval officers, embarked in 1872 on a voyage of discovery on the Barents Sea and eastward. They were caught in the heavy polar ice and carried from Novaya Zemlya to newly discovered islands which were named Franz Joseph Land. Weyprecht (1879), as a result of scientific observations, wrote a treatise on the various forms of polar ice. Nordenskiold, in 1878, commanded the *Vega* which made a complete passage along the Russian edge of the polar sea, entering the Arctic Ocean around the northern end of Norway and emerging through the Bering Sea into the Pacific. The fruit of this voyage was a great addition to our knowledge of the sea ice and the *Vega's* also still bears the unique distinction of being the only complete west to east navigation. Pettersson (1883) one of the *Vega's* scientific staff, made many important studies on the physical properties of sea ice.

The ill-fated DeLong expedition, which was aimed for the pole, met disaster when the *Jeannette* sank in June, 1881, nearly 300 miles north of the New Siberian Islands, at a point deeper in the polar cap ice than ever a vessel had penetrated before. The westerly drift of several hundred miles along the Siberian shelf revealed for the first time definite information on the movement of the great sea-ice cover. The expedition, however, which stands out above all others in Arctic annals was the Norwegian north polar expedition, 1893–1896, led by Nansen. The *Fram* was forced into the polar cap ice north of the New Siberian Islands in September, 1893. For nearly three years she remained tightly surrounded by heavy ice fields, all the while valuable observations were being taken on the drift, behavior, and physical state of the ice. This is the only craft that has ever been carried across the deeper part of the north polar basin, and mainly as a result of Nansen's observations we have learned it to be one vast central expanse of ice interrupted only by temporary leads and pools. Nansen's observations that the ice con-

sistently drifted to the right of the direction of the wind led to greater knowledge of the general laws for the frictional effect of the wind on free moving bodies as affected by earth rotation.

The sledge excursions of Cagni and Peary out onto the polar ice had for their main object spectacular and historic dashes for the pole. The scientific accomplishments, therefore, from the nature of the expeditions, are not as valuable as otherwise might have been the case, but considerable information regarding the state and movement of the cover, nevertheless, resulted. The Russians, amongst whom might be named Toll, Kolchak, Liakhof, Wrangel, Makarov, and Vilkitski, have all made noteworthy contributions on the ice of the Arctic Ocean. One of the most brilliant students and keenest of observers of the group is Kolchak, whose daring journey of several hundred miles in an open whaleboat to Bennett Island, far northward in the icebound ocean, was a notable achievement.

To Stefansson, Storkerson, Bartlett, and MacMillan during the period 1913 to 1917, we owe our knowledge of the ice conditions around the American margin of the polar cap, the tumbled, chaotic condition of which suggested for it the name of paleocrystic ice.

Lastly we wish to mention Amundsen who, in the *Maud*, attempted to repeat the original exploit of Nansen, and after several years, 1918–1925, abandoned further efforts to be caught up by the drifting pack. This period meant no inactivity, however, as Malmgren's (1928) observations on the properties of sea ice, and Sverdrup's (1929) discussion of the hydrography and the movements of the ice with wind and current greatly clarify many of the ice problems. Amundsen was the first to introduce and employ aircraft for Arctic exploration. In 1925 his uncompleted flight with planes from Spitsbergen to Alaska awakened interest and opened new opportunities to study Arctic ice from the air. Byrd, Nobile, Wilkins, and now the proposed program of aero-Arctic, just in the stages of preparation, as well as navigation by a submarine, mark the dawning of a new day in northern exploration.

Our knowledge of the ice in the regions of the northwestern North Atlantic; i. e., the region of Baffin Bay; its tributaries; Davis Strait; and south to Newfoundland, has been gained as these countries have passed through the following eras: The early voyages of exploration represented by Frobisher, Davis, Hudson, and Baffin; exploitation of whaling, fur trading, and missionary colonization; international commercial rivalry seeking a northwest passage to India; the traverse of explorers on their way to and from polar regions; and finally, the present era of scientific investigation. Martin Frobisher, groping for the elusive northwest passage, made the first historical crossing of Davis Strait, when in July, 1576, under the shores of southern Baffin Land, he came amidst drifting icebergs and dense fogs. The largest iceberg was measured and found to be 330 feet high. Next appears Davis, combining the qualities of an able British seaman and a patient, scientific observer, and carrying out a series of explorations of Davis Strait, following the ice from the southern end of Baffin Bay to well south on the coast of Labrador. As a result of his voyages cartographers were able to draw the first intelligent maps of Davis Strait and in a general way indicate the drift of the ice. Returning from a voyage of discovery to New York, Hudson was commissioned in 1610 to seek a passage by the

wide opening pointed out by Davis as "furious overfalls,"[1] through which Hudson eventually sailed. The first record of navigating Baffin Bay was in 1616 when Baffin successfully followed along the Greenland coast as far northward as Kane Basin before turning back and sailing out along the Baffin Land shore. Baffin's voyages to the northwestern extremes of the North Atlantic showed the hitherto unknown openings from the polar ocean into Baffin Bay.

The Dutch whalers being partly forced out of Spitsbergen by the keen competition of the British were alert to adopt Baffin's suggestion of fishing in Davis Strait, which they entered as early as 1620. The British followed later, and heeding the reports of Ross and Parry, extended their fishery northward to the very headwaters of Baffin Bay, and in a short time became exceedingly proficient at ice navigation. The designation of the pack as "middle ice" and "west ice," and the description of the movements of the pack we owe to Captain Marshall and Doctor O'Reilly (1818) of the whalers. The long mysterious disappearance of Sir John Franklin and his ships and men, 1845–1851, brought forth no less than nine relief expeditions to the Canadian Arctic over a period of 10 years. Much knowledge regarding the icy character of the waterways in the archipelago was obtained through the drift of ice-beset ships of these expeditions and in consequence it was learned that a surprising amount of ice formed in the Arctic Ocean is drawn into Baffin Bay yearly.

Smith Sound, also Kennedy and Robeson Channels, have often been designated the American route to the pole, followed by Kane, Hayes, Hall, and finally successfully on ship and sledge by Peary. The results of these activities, separated from each other by a number of years in the nineteenth century, increased our knowledge of the state of the ice at the head of Baffin Bay. Whaling has practically ceased in the northwestern sector. Fur trading, however, is still carried on by the Hudson Bay Co. but the returns are small compared with years ago. Denmark still continues her colonization of the Greenlanders along the eastern side of Davis Strait. Except for the annual visit of the Canadian Government ship *Beothic* to the mounted police and Hudson Bay posts in Ellesmere Land and Baffin Land, and the regular steamers as far north as Upernivik, Baffin Bay is to-day quite deserted. The recent interest being displayed by Canada to establish a shipping port on the west side of Hudson Bay at Port Churchill is the only activity in the region of Davis Strait and to the westward. If this project is successful, much information on the ice and currents of this region is, in time, bound to accumulate.

Greenland supports the greatest reservoir of land ice in the Northern Hemisphere. The German geologist Giesecke, in the course of a mineralogical survey beginning in 1806, described many of the ice fjords. Rink, the Danish geologist, and later governor of Greenland, 1848–1851, studied the motion of glaciers and the calving of icebergs, his publications receiving a wide distribution and arousing lively interest in scientific circles. The names of the other scientists who have carried out investigations on the ice fjords, the glaciers, and the calving of icebergs, practically all on the west coast of

[1] Chart of Davis Strait published Apr. 20, 1875, by the British Admiralty, London.

Greenland, are: Steenstrup in 1883 (for a number of years); Helland, Norwegian geologist, in 1872–1876; Rabot and Fries in 1875; Hammer in 1879–80; Ryder in 1886–87; Astrup in 1893; von Drygalski in 1891–1893; Engell in 1902; Porsild in 1918; and L. Koch over a period of recent years. The results of the labors of the foregoing men, especially Steenstrup, Drygalski, and L. Koch, have permitted us in this paper to name the glaciers which are iceberg-producing and estimate the relative numbers which are normally released every year. Our knowledge of the shape and thickness of the ice cap is furnished by several crossings of the great inland ice, first by Nansen in 1888, and then by DeQuervain in 1912; Rasmussen in 1912; J. P. Koch and Wegener in 1913; and more recently by Lauge Koch. Several others have also made important incursions, for instance, Jensen in 1878, Nordenskiold in 1883, Garde in 1893, Mikkelsen in 1910, and lately Hobbs and Wegener. The only other glaciated area in the northwest sufficient in size to contribute any considerable supply of icebergs is Ellesmere Land, the inland ice of which has been explored by Greely in 1883; by Sverdrup in 1902–1904; and by MacMillan in 1914.

As a by-product of early exploration, whaling, and trading, as we previously described, there has been assembled a great deal of knowledge on the behavior and regional distribution of floating ice in the western North Atlantic. Later scientific expeditions concentrating mostly along oceanographical lines have considerably advanced our understanding of drifting ice. A large number of accurate oceanographic observations regarding the system of circulation which prevails, for example, in the waters of Davis Strait and over the Grand Bank, means greater enlightenment on the movements of the ice. Baffin Bay, due to inaccessibility, has witnessed few subsurface investigations: Hamberg in the *Sofia* 1883, Nielsen in the *Tjalfe* 1908–9; and a more thorough and systematic oceanographical examination by Riis-Cartensen in the *Godthaab* in the summer of 1928 complete the local list. The account and results of this last work have not yet been published but from the printed data appearing in the International Hydrographic Bulletin we have here interpreted some of the indicated movements of the water and the ice. The hydrography of Davis Strait is known from the following investigations: Wandel in the *Fylla*, 1884, 1886, 1889; Knudsen in the *Ingolf*, 1895; Nielsen in the *Tjalfe*, 1908–9; Mathews in the *Scotia*, 1913; Hjort in the *Michael Sars* in 1924; Jensen in the *Dana*, 1925; Iselin in the *Chance*, 1926; Riis-Cartensen in the *Godthaab*, 1928; Ault in the *Carnegie*, 1928; and our own voyage in the *Marion*, 1928. The results of the earlier of these expeditions threw little light on the scheme of oceanic circulation there and on the drift of the ice, while the work of the *Scotia* and of the *Carnegie* was only incidental to much larger fields of operations. The most intensive work in Davis Strait has been accomplished by the *Tjalfe*, *Godthaab*, and *Marion*. As a result of these latter operations we now possess a fairly clear picture of the paths that the ice follows from Baffin Bay to Newfoundland and also of the approximate rate of its southern dispersal.

Various aspects of the ice problem in the North Atlantic south of Newfoundland have been the subject of exhaustive investigation by

the international ice patrol.[2] This service was established as a result of the great number of sea disasters which occurred around Newfoundland and culminated in the tragic loss of 1,500 lives and the steamer *Titanic* in 1912. Besides the practical work of locating the ice and warning ships of its presence, the ice patrol in its 16 years of service, especially since 1921, has assembled a large amount of ice and oceanographical data. The area so frequently surveyed includes the Grand Bank from the forty-eighth parallel southward to 39° latitude and east and west between meridians 45 and 55. As a result the behavior of the ice south of Newfoundland can now be treated in great detail.[3]

Using the available data several researches have been published on the subject of Arctic ice and its drift into the North Atlantic Ocean. Thus Rodman (1890) published a report on the ice and its movement to the North Atlantic. Howard (1920) prepared an article for the Monthly Pilot Chart issued by the British Meteorological Office, and Hennessey (1929) has continued the work in the Marine Observer. Mecking (1906–7), with the assistance of the Deutsche Seewarte, made a special research of ice conditions in Davis Strait, and with especial reference to the effect of meteorological conditions on distribution of the ice in the Atlantic. Bowditch (1925) and the Newfoundland and Davis Strait and the Arctic Pilots (U. S. Hydrographic Office, 1884, 1909, 1921), all contain chapters devoted to drift ice and icebergs for the information of seamen. Smith (1927a) for several years published a paper devoted to ice on the back of the Monthly Pilot Chart, United States Hydrographic Office.

All of the foregoing accounts have been seriously handicapped by the lack of a sufficient knowledge of the cold currents north of Newfoundland and of the consequent behavior of the ice there.

"MARION" EXPEDITION

The need for ice and oceanographic observations, from Newfoundland northward to Baffin Bay, had been felt by many scientists for a number of years. Accordingly, in June, 1928, the United States Coast Guard, in charge of the ice patrol, assigned the patrol boat *Marion* for a northern expedition. The *Marion* sailed from Sydney, Cape Breton, on July 16. (See narrative account in Part 1 of this bulletin.) A total of eight weeks was devoted to a detailed ice and ocean-current survey of the waters from the latitude of St. Johns, Newfoundland, to the mouth of Baffin Bay, and extending all the way east and west between North America and Greenland. The success or failure of any mission to the waters of Davis Strait and Baffin Bay depends upon the ice conditions prevailing in these waters during the brief warm period of the year, and the *Marion* expedition, the summer of 1928, was very much favored

[2] The British auxiliary bark *Scotia* inaugurated the first systematic plan of scientific exploration of the ice regions off Newfoundland in the spring of 1913; Mathews (1914) has written a very illuminating account of the oceanography of the Grand Bank.

[3] Previous to 1928 the observations of the ice patrol had been confined to the area south of Newfoundland with one exception. In July, 1914, the Coast Guard cutter *Seneca* sailed from St. Johns, Newfoundland, bound for the sea which separates Labrador from Greenland. The purpose was "to observe the origin of the ice which annually appears on the banks of Newfoundland, and to investigate the agencies by which it is transported from the north." (Johnston, 1915, p. 31.) Unfortunately 1914 was a bad ice year, and after proceeding about 200 miles north of St. Johns, the *Seneca* came into ice so heavy and bergs so numerous that it was dangerous to attempt further progress.

in that the ice was scarce. Another circumstance which greatly enhanced the value of the observations of the *Marion* expedition was the fact that the *Godthaab* expedition simultaneously carried out an oceanographic survey in the more northern waters of Baffin Bay, allowing direct comparison.

The principal task of the *Marion* expedition was the collection of a record of temperature and salinity—the raw data—from as many selected points of observation and depths as possible in the waters of Davis Strait. This material, consisting of over 2,000 surface and subsurface observations, has been subjected to Bjerknes's (1910, 1911) hydrodynamic formulae according to the methods employed on international ice patrol and described by Smith (1926, pp. 1–50).

As a result of the *Marion* expedition the prevailing oceanographic circulation of Davis Strait has been mapped from the lower end of Baffin Bay to the latitude of St. Johns, Newfoundland. A complete report on the dynamic oceanography is to be published in Part 2 of this bulletin, entitled "The Marion Expedition Under the Direction of the U. S. Coast Guard, 1928. Scientific Results. The Physical Oceanography of Davis Strait." (In press.)

The dynamic topographic maps which are described and illustrated in Smith (1931) have been used in the present paper as the basis of describing and interpreting the direction and movement of the icebergs in Davis Strait. (See especially in this connection fig. 95, p. 147, in the present paper.) Our conclusions on the general movement of icebergs in Baffin Bay have been based on the dynamic topographic map (fig. 91, p. 139) constructed from the *Godthaab's* observations. The dynamic topographic map showing the stream lines of the gradient currents furnishes an excellent presentation of the normal courses taken by the icebergs, if we assume that the deep-drafted long-life icebergs outside coastal promontories are in the main controlled by ocean currents. On several previous occasions I have found a good agreement between the stream lines of the currents as represented by the dynamic topographic maps (see figs. 103, 104, 105, 106, p. 162) and the movement of the bergs in the Grand Bank region. See, in addition, Smith (1927, p. 118) (1927a, pp. 70–93). Since the current maps of the *Marion* expedition are similar in every respect to those constructed on ice patrol, there is every reason to believe the agreement between calculated currents and ice drifts holds equally true for Davis Strait and Baffin Bay.

The *Marion's* 2,000 observations of salinity and temperature, covering the waters of Davis Strait (see station table data, Part 2 of this bulletin), thus constitute original basic data from which to deduce the circulation and the behavior of the ice. By synthesis with the various earlier data it is now possible to present a connected picture of Arctic ice and of its southward drift.

FLOATING ICE

Ice which is sighted floating at sea may have been formed either on the salt water itself or upon the land. Floating ice, therefore, as it is pertinent to the present discussion, separates into two great divisions as it hails from independent sources namely, land ice and

sea ice.[4] The former when found drifting in the ocean, is usually in the form of icebergs which from their relatively great mass and density, form especially dangerous obstacles in the paths of navigation. Sea ice of course is the ice formed of salt water, and because of its great horizontal proportions, covers a fairly wide expanse in the northern hemisphere.

The Properties of Sea Ice

Sea ice differs from fresh-water ice because of the presence of salt in the former and on account of this attribute great contrasting differences occur between the physical properties of the two kinds of ice.

We may learn much regarding the physical properties of sea ice by observing what happens at the boundary between ice and water when the water approaches freezing conditions. Water, according to Johnstone (1923, p. 185), is composed of H_2O molecules mixed together in single, double, and triple combinations. The triplex molecule, is the one closely associated with the formation of ice. It is believed these triplex, so-called ice molecules, are present in water in varying proportions, even above the freezing point, depending upon the particular temperature of the liquid at the time. They, therefore, lie suspended in the water, similar to salt in solution, until the saturation point is reached, which occurs at a temperature of $-1.85°$ C. (28.7° F.), for common sea water (salinity 34 0/00) when the liquid is transformed to a solid state. The subtraction of an amount of heat energy equivalent to the heat of fusion of that given mass of sea water, often the addition of a minute piece of ice itself, is sufficient to initiate actual freezing. The first sign of a change of physical state under the microscope is a cloud of small disklike particles which floculate and grow, finally passing from an original colloidal state into true crystalline form. If they are allowed to develop in quiet water, the crystals assume most beautiful feathery designs, but more often they collect in a network of fine spicules and elongated prisms, which extend down and out as streamers and plates. If the water be exposed to a very sudden chilling by the wintry atmosphere, such as often occurs within the bounds of the polar ice cap, the colloidal stage of ice growth is masked by the direct creation of the more familiar crystalline structure.

The processes of solidification of sea water differs in one important particular from that of pure fresh water. The first crystals to appear in salt water are comparatively fresh because the pure H_2O molecules tend to separate and to congeal first. As freezing temperatures continue, the process spreads to the unfrozen liquid lying between the strings and plates of the initial ice crystals. This brine, which has now become more or less imprisoned, tends to sink by virtue of its greater specific gravity, causing a sagging of the entire pulpy mass, and much of the brine in this manner actually drains out. Gradually, as the fabric attains a firmer structure, the remaining salty liquid is completely entrapped and so prevented from escaping out of the ice body. Therefore, we have finally a solid mass consisting of innumerable pure ice crystals frozen together, between

[4] Fresh-water ice formed in rivers and later discharged downstream into the sea exhibits a maximum field of dispersal in the Siberian offing of the Arctic. Even here, however, river ice is of such comparatively limited extent, compared to the magnitude of the regions under discussion, that no discussion of it is needed.

which are caught a smaller number of salt crystals, together with an appreciable quantity of concentrated brine. The relative amount of the latter is dependent upon the temperature; the colder, the more brine is imprisoned in the cellular structure.

It is an interesting question whether all the salts contained in sea water are to be found in sea ice; also if the salts in the ice are combined in the same relative proportions as they are in the sea water from which the ice was frozen. Laboratory experiments to test the solubility of its various salts require a supercooling and agitation of the liquid, seldom, if ever, experienced under natural conditions, but such investigations furnish interesting information regarding the eutectic point of the salts and also regarding the salinity of the resulting solid. According to Johnstone (1923, p. 186) every sea-water salt has been found to possess its own individual temperature of solubility which if depressed, causes that particular salt to precipitate from the mother liquid. First to crystallize, as previously stated, are the water molecules which for a sample of 33 per mille salinity takes place at −1.8° C. (28.8° F.). When −8.2° C. (17.3° F.) is reached Na_2SO_4 begins to precipitate, and at −23° C. (−9.4° F.) solid NaCl separates from the remaining brine.

The behavior of sea water when supercooled has led to the establishment of a theory that yearly great quantities of SO_3 are removed from polar regions and deposited in the Atlantic as ice floes saturated with that substance move out of the north to melt in lower latitudes. Malmgren (1928, p. 7) had an excellent opportunity to test this theory by determining the ratio of SO_3 and Cl, for samples of sea water and of sea ice from several of the freezing zones of the polar sea. If a selective process of salt crystallization prevails, SO_3 and Cl would be the substances first to exhibit variations. In all cases, however, the sample pairs of ice and mother liquid showed almost the same ratio of SO_3 and Cl.

Wiese (1930) reports Liakionoff's investigations of this phenomenon in the Barents Sea the summer of 1930. Liakionoff found contrary to Malmgren that a deficit of Cl and a surplus of SO_3 actually prevailed both in the melting water and in the sea ice itself. The ratio ranged from Krummel's value of 0.1150 for ordinary sea water to a maximum of 0.1700 in old sea ice. These results agree moreover with those of Ringer and Pettersson, and therefore it appears that a selection of the salts in sea water occurs upon freezing. Future investigations will determine this with more certainty. The freezing point of sea water is defined as that temperature at which the first ice crystals appear in the liquid.

The freezing points of samples of water of varying salinity are given herewith:

Salinity 0/00	Freezing point		Salinity 0/00	Freezing point	
	° C.	° F.		° C.	° F.
25	−1.35	29.6	32	−1.74	28.9
26	−1.40	29.5	33	−1.80	28.8
27	−1.45	29.4	34	−1.85	28.7
28	−1.51	29.3	35	−1.91	28.6
29	−1.59	29.2	36	−1.97	28.5
30	−1.62	29.1	37	−2.02	28.4
31	−1.68	29.0			

The point of solidification of sea ice; i. e., the temperature at which it becomes a true solid, is however, quite different from the freezing point of sea water. The fact that sea ice is a conglomeration of pure ice crystals and of particles of brine, combined in varying proportions, due to varying temperature and salinity, naturally prevents one single definite point of solidification. Strictly speaking, the solidification process in such ice can not be considered completed until the last salt solidifies, namely $CaCl_2$, the eutectic point of which is −55° C. (−67° F.). For general purposes of calculation, therefore, the freezing point of sea ice is assumed to occur at the eutectic point of its major salt NaCl (about 77 per cent of the salt content of sea water), i. e., at approximately −22° C. (−7.6° F.). The closer the solidification point of a sample of sea ice lies to the freezing temperature of sea water, the more homogeneous will be the composition of the ice, and the less the deposition of brine and salt. The lower the temperature to which sea water is exposed, the more rapidly will the net of ice crystals form.

One of the most astonishing things about sea ice is the fact that it is so fresh. Malmgren (1928, Tables 4 and 5), has made several determinations of the salinity of the ice of the polar cap, showing it to be subject to a relatively wide range—0.05 0/00 to 14.59 0/00. A salinity of 7 0/00, therefore, may be taken as the salt value of ordinary sea ice.[5] It would be interesting to learn what is the average salinity of the Arctic water from which ice of 7 0/00 is formed. The salinity of the surface layers of the north polar ocean, outside of the continental slope, as Nansen (1928, p. 11) has pointed out, are greatly diluted by the land drainage from the Eurasian side. Nansen estimates that the inflow of fresh water from Siberian rivers alone is sufficient to cover the ocean from the Asiatic coast to the pole each year, with a surface film 3½ feet thick. The distribution of salinity of the water as found along the *Fram's* track was 28.40 0/00 to 29 0/00 to a depth of 30 meters, 90 miles north of the New Siberian Islands, and only 31.7 0/00 to 33 0/00 at 30 meters, within 300 miles of the pole. In view of these observations we may conclude that the surface layers in the polar basin, i. e., the water subjected to freezing, vary little from a mean salinity of 31 0/00.[6] The average salinity of sea ice, approximately 7 0/00, shows, therefore, that only about one-fourth of the total quantity of salts of sea water enters its ice. Nansen, on the drift of the *Fram*, made no determinations of the salinity of sea ice, but from our information on the distribution of the salinity in the Arctic Ocean, the *Maud's* ice experiments can, without criticism, be compared directly with the *Fram's* hydrographical observations.

The rate at which sea water freezes is a very important factor in determining the percentage of salt imprisoned in the ice. The more rapid the process, the more sudden the pure ice crystals impregnate the water, catching within their meshes a large quantity of brine drops. The saltiest piece of ice that Malmgren (1928, p.

[5] Weyprecht (1897, p. 58) during the drift of the *Tegetthoff* found the salinity of a piece of thin ice formed rapidly under low temperature in the Arctic to be 25 0/00, and this is probably as salt as sea ice ever is.

[6] No other ocean in the world can show such low salinities more than a few miles out from land. The inflow of land water around the basin's border fails to explain a salinity uniformly so low over the entire expanse. Probably the absence of evaporation and the presence of ice are the two factors chiefly responsible, combined with the seasonal cycle of freezing and melting, repeated throughout the years since the present cap became permanent.

11) found in the Arctic Ocean was 14.59 0/00, this being a sample cut from a sheet of thin ice that had been exposed to the very low temperature of about −40° C. (−40° F.). It is true that the ice cover is continually being rended apart exposing the water to extremely frigid temperatures. But the total amount of the salty ice thus produced is comparatively small, because in relatively few instances does ice form under supercooled conditions.

The two principal regions of ice formation in the Arctic Ocean are:

(*a*) The shallow waters of the Eurasian shelf and in the northern North American country.

(*b*) The underside of the permanent polar ice cap.

In the case of (*a*) the sea surface normally becomes covered early in the autumn before the atmosphere has greatly cooled, and tends to remain more or less screened during the winter. Cases (*a*) and (*b*) are now similar and new ice is added throughout the colder months of the year, on the underside of the cover, where the water is beyond the reach of severe atmospheric chilling.

The fact that the amount of salt in sea ice varies directly with the rate of freezing (inversely as the temperature), causes the top of a glaçon or floe, the part exposed to the coldest temperature, to be composed of the saltiest ice.[7] Several writers on polar phenomena have reported the presence of solid salt on the surface of ice sheets, which sometimes coats them so heavily that it has been likened to the early morning collection of hoar frost.[8] The growth in thickness of young sea ice from the top surface downward proceeds at a slower and slower rate the further the underside retreats from the source of freezing. Similarly the insulating effect of a snow cover in retarding the freezing processes is proved by the formation of fresher ice on the underneath sides of those floes that are heavily snow decked. The heat conductivity of ice is very poor, about one-hundredth part that of iron. The upper surface of old ice being comparatively fresh is a much poorer heat conductor than is young, salty ice, and similarly, retards freezing. A set of salinity observations by Malmgren (1928, p. 6) through the vertical cross section of a young sheet, gives the following distribution of salt with depth:

Depth of ice from surface (cm.)	0	6	13	25	45	82	95
Salinity of the ice (0/00)	6.74	5.28	5.31	3.84	4.37	3.48	3.17

The fact that the surface layers of the sea are normally fresher than those deeper down appears to have no appreciable effect in varying the salinity of the ice, top to bottom, because its thickness includes only a very small part of the range of salinity with depth of water.

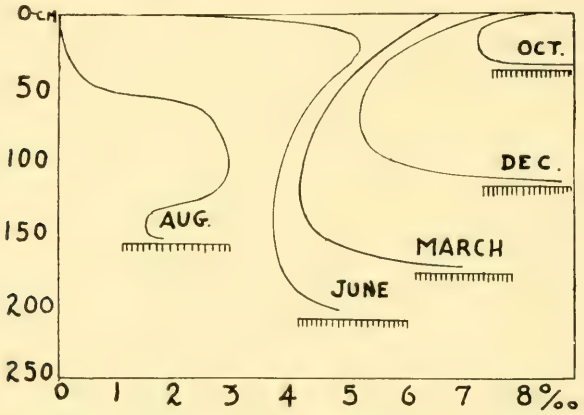
Since the underside of the cover is not only the freezing surface but the plane of contact between the ice and the concentrated brine, one might naturally reason that the thicker the ice became, the richer it would grow in salts. The fact that the salinity of ice decreases with the depth, however, indicates (*a*) that the processes of precipitation are more than sufficient to counterbalance the increase in salt concentration of the mother liquid; (*b*) that the mother

[7] A thin veneer of new ice may often overlay young ice formed early in autumn from the thaw water of the Arctic pack.

[8] Weyprecht (1879, p. 58) states that the salt is not pure, but consists of fine needles of ice tipped with small salt crystals.

liquid undergoes dilution continually as a result of mixing with the surrounding waters; or (*c*) that freezing is retarded to a very slow rate.

One of the most interesting features connected with the physical properties of sea ice is the seasonal change in salinity which develops with age. Young ice; that is, the variety which has formed during the autumn and winter, is observed to decrease slightly in salinity at all depths shortly after freezing; it then continues nearly constant in this respect throughout the winter, with a second and greater loss on the arrival of the succeeding summer. When air temperatures over northern seas rise above −4° or −5° C. (24.8° or 23° F.) the minute brine particles and salt crystals that have lodged throughout the interstices of the pure ice crystals begin to melt away from the more solid structure. Since the heat which comes from the sun is too feeble to melt the ice crystals in the permanent polar cap, at least to any great exent, the brine drains off and down leav-



THE CHANGE IN THE SALINITY OF SEA ICE WITH AGE

FIGURE 2.—Young ice formed in October in the north polar basin has a thickness of about 45 centimeters and contains approximately eight parts per thousand of "salt." The same ice by the following summer has increased about three times in thickness, but its salinity has decreased about four times the original proportions. (Figure from Malmgren, 1928.)

ing the tops of the floes of a composition closely approaching that of fresh ice. The structure at the same time gradually becomes granular and after a few years it is difficult to distinguish an old floe from glacier ice. The penetration of solar warmth into the ice in the Arctic regions by the end of summer, is sufficient to reduce the salinity 2 or 3 parts per thousand to depths of 3 to 5 feet. The pressure ridges so prevalent in the regions of paleocrystic ice are found to be more completely washed free of salts than any other northern sea ice. The brine that drains out of the uppermost layers is prevented from settling directly downward by the bottom layers which, constantly immersed in cold water, remain so solid that the only drainage is through the many narrow-cut channels that furrow the floes. If a series of careful measurements of the thickness of the ice cover be made during the early part of summer when melting begins, they will show at first a gradual decrease, then a brief thickening as the water thawing from the tops of the floes and from the

snow freezes upon reaching the frigid sea. But as increased solar warming raises the temperature, so it increases the rate of melting, and the floe finally becomes thinner.

The loss of salts during the warmer months of the year leaves the polar cap mantled with fresh ice possessing a higher melting point and therefore greater permanence. Old sea ice accordingly survives much longer than the young, provided both samples be of the same volume, when exposed to similar melting conditions. Many people have regarded with considerable skepticism the statement that sea ice furnishes a supply of very good drinking water. Experienced ice navigators, however, are well aware of the freshening of sea ice with age, since the fresh-water pools on old pans have long been utilized as a never-failing supply. The rule is to moor the ship with ice anchors to a well-selected old floe and then run a hose to a near-by pool, where water has collected either from the weathered side of the ice itself or from the preceding winter's snow cover. It is important to pump only from pools located well in from the edge of the floe to avoid the possibility of contamination by salt spray.[9]

The physical behavior of salt water when subjected to freezing temperatures is different from that of fresh water under similar conditions, because the former contracts right to the freezing point.[10]

The fact that salt-water ice is formed when the freezing mass is at maximum density, causes it, therefore, to be slightly heavier than fresh-water ice. The fact that it does not float lower is due mainly to the considerable quantity of air which is in the water when it freezes. The density of pure, fresh-water ice, according to Barnes (1928, p. 25) is 0.91676, and that of sea ice formed from the water of the salinity that normally comes under freezing conditions ranges from 0.857 to 0.924.[11]

The relatively great range in the specific gravity of sea ice is due, according to Malmgren (1928, p. 17), partly to large quantities of air or water which the ice may absorb during the summer directly from the atmosphere. The reason that water is not sucked up from below into the spongelike vacuoles (caused by the escape of salts) is because much of the lower layers of the sea ice in polar regions remain unmelted even in summer and, therefore, waterproof. If, however, sea ice drifts out of the Arctic into waters warmer than 0° C. (32° F.), melting will begin below the water line, causing ingestion and a drowning of the ice. Arctic pack ice, therefore, in the North Atlantic floats lower than it did nearer its source.

As a corollary to the above, if sea ice is exposed to temperatures as low as −20° C. (−40° F.), nearly all the brine globules freeze into salt crystals, causing the ice to swell to a maximum volume.

The physical appearance of sea ice is quite different from that of fresh-water ice: the former exhibiting an opaque slaty whiteness.

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[9] Pettersson (1883, p. 304) collected sea ice floating in the polar drift current northwest of Spitsbergen which contained less than one-fourth the amount of chlorides found in the drinking water of Stockholm.

[10] As a matter of fact the saltiest water, i. e., pure ocean water, seldom reaches a freezing temperature, because it is usually protected by warm convectional currents. The freezing regions of the hydrosphere are confined mostly to the shallow waters of continental shelves and to epi-continental seas, where the salinity of the surface layers rarely exceeds 33 0/00; so that freezing takes place at a temperature of −1.8° C. (28.8° F.).

[11] We performed the following experiment: A quart of sea water of salinity 34.3 0/00 was frozen into a cake of solid ice, with no opportunity for any of the salts to escape, and despite its high specific gravity the block floated in a jar of fresh water. The height of flotation was noticeably lower than in the case of fresh-water ice, but the experiment definitely shows, nevertheless, that ordinary sea ice has reserve buoyancy. It is often observed, for instance, floating off river mouths where the water is comparatively fresh.

Sea ice has none of the hardness and extreme brittleness that characterizes fresh ice under similar coldness. The fact that salts lodge between the pure ice crystals prevents the latter from interlocking tightly, materially reducing the tensile strength, and permitting a maximum amount of bending without fracture.[12]

Ice Terminology

The following definitions are apparently more or less standard in ice terminology:

Slush or sludge.—The initial stages in the freezing of sea water when it is of a gluey or soupy consistency and when the surface of the water takes on the appearance of cooling grease, with a peculiar steel-gray or lead tint.

Pancake ice.—Small cakes of new ice approximately circular with raised rims. The rims give them a striking resemblance to pancakes. The diameter of the cakes is from 1 to 3 feet and their thickness is up to 2 to 4 inches.

Young ice.—A compact sheet formed by the repeated freezing together, breaking up, and refreezing of cakes of pancake ice. Its initial thickness is 1 to 3 inches and this may increase to a maximum during a winter in the Arctic regions of 6 to 9 feet.

Fast ice.—Horizontal ice formed by the freezing of the sea out from the shore. The 12-fathom contour is approximately the outer limit of the spread of fast ice along open coast lines in the polar basin.

Ice foot.—That part of the fast ice which forms and builds on the shore itself and therefore is unaffected by vertical motions such as the tides.

Anchor ice.—All submerged ice attached to the bottom irrespective of its mode of formation.

Pack ice.—Sea ice which has drifted from its original position.

Polar cap ice.—Oldest and heaviest of ice pack, characterizing the central portions of the north polar basin.

Ice field.—An area of ice other than fast ice of such an extent that its limits can not be seen from the ship's masthead.

Ice floe.—An area of ice other than fast ice from one-third of a mile in diameter to the size of an ice field.

Glaçon.—Any piece of pack ice ranging in size from a cake 2 to 3 feet in diameter to a floe.

Hummock.—A piece of ice formed by marginal crushing and heaping up of the sea ice.

Hummocking.—The process of pressure on sea ice resulting in a heaping up of the sea ice.

Floeberg.—A massive hummock; the results of great pressure and piling up of the heaviest forms of sea ice.

Rotten ice.—Any pack ice which has become much honeycombed through the latter stages of melting so that it lacks the strength of other ice.

Crack.—Any fracture or rift in sea ice but not sufficiently wide to permit navigation. Kinds: (*a*) Tidal; (*b*) temperature; (*c*) shock and pressure.

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[12] Weyprecht (1879, p. 49) states that soon after the formation of young ice in the autumn it records every footprint as easily as newly fallen snow.

SLUSH AND SLUDGE BEGINS TO FORM

FIGURE 3.—The initial stages in the freezing of sea water. Slush and sludge of soupy consistency is forming in the spaces between the old glaçons. (Photograph by H. G. Ponting in the Scientific Reports of the British Antarctic Expedition.)

YOUNG ICE

FIGURE 4.—Glaçons of young ice which will soon cement together again in a continuous sheet with the further advance of winter temperatures. The several small circular cakes with the raised rims which compose the glaçons are called pancake ice from their close resemblance to "pancakes." This form represents the intermediate step in ice formation between slush and young ice. (Photograph by H. G. Ponting in the Scientific Reports of the British Antarctic Expedition.)

ICE FLOES AND ICE FIELD

FIGURE 5.—A view of sea ice to illustrate the terms "ice floes" and "ice field." A floe is any ice area from one-third of a mile in diameter to as far as can be seen from a ship. An ice field is an area so great that its limits are beyond the vision of the masthead lookouts. This scene has all the appearance of the southern edge of the pack which reaches the Grand Bank, south of Newfoundland every spring. (Photograph by F. A. Matisen in Kolchak, 1909.)

GLAÇONS

FIGURE 6.—Sea ice in the form of glaçons; the term used to designate any piece of sea ice ranging in size from a cake 2 to 3 feet in diameter to an ice area of one-third of a mile in diameter. (Photograph from Arctowski (1908a).)

Lead or lane.—A navigable passage through any kind of pack ice.

Pressure ridge.—The marginal elevation and heaping up of any kind of pack ice when opposing forces press it together.

Polynya (Russian).—Any sizeable water area, not a crack or a lead, which is surrounded by sea ice.

Pool.—A depression in the fields, floes, or glaçons that contains fresh water.

Frost smoke.—The foglike clouds that form over newly opened water areas in sea ice in the far north.

Water sky.—Dark streaks on the clouds due to the reflection of polynyas, or of the open sea, in the vicinity of large areas of ice.

Ice blink.—The whitish hazy glare on the clouds or near the horizon produced by the reflection from large areas of sea ice in the vicinity.

Iceberg.—A large mass of glacier ice found in the sea.

Growler.—A low-lying piece of glacier ice not so large as a berg.

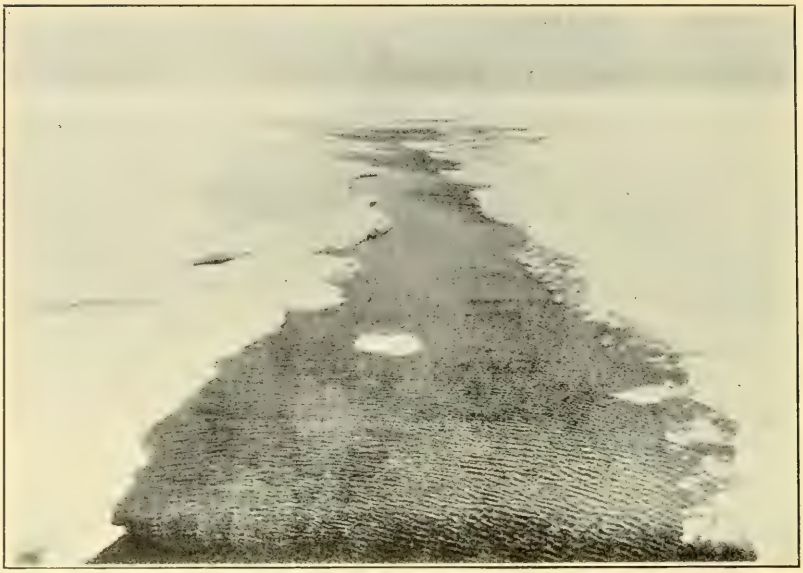
Classification of Sea Ice

The ice cover that spreads over northern seas, for clearness and simplicity, may be classified in accordance with its distribution, i. e., in the form of concentric belts, focused around a center called the ice pole, normally located in the vicinity of latitude 83° to 85°, longitude 170° to 180° west. The southernmost boundary of the outer circumpolar region coincides with the southernmost sea area which becomes ice covered, either because of freezing temperature or through drift of the ice from colder regions. The southern limit of sea ice naturally deviates from the latitude parallels because of many varied conditions, such as the distribution of land and sea masses, the bathymetrical features of the ocean basins, the ocean currents, and the winds. All these factors tend to modify a symmetrical arrangement which otherwise would designate the geographical pole as the iciest spot on the earth.

The ice cover of northern waters may be classified as follows: (*a*) Fast ice; (*b*) north polar cap ice; (*c*) pack ice.

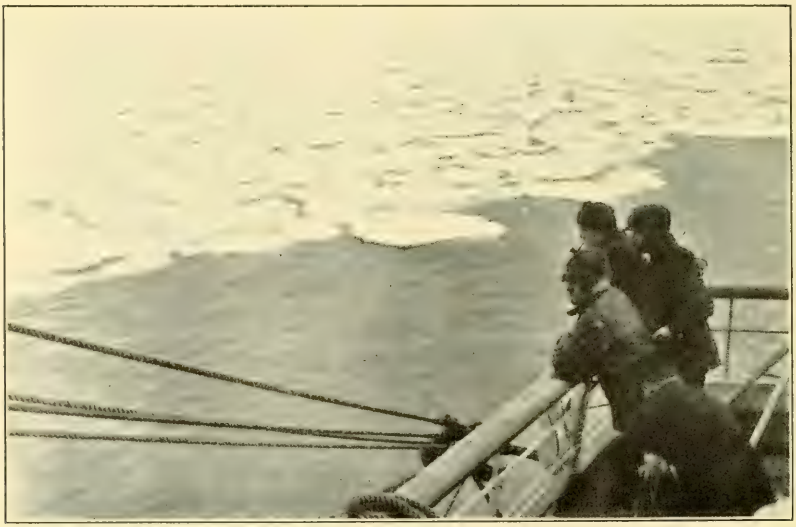
Fast ice (Transehe 1928, p. 99), or winter ice (Koch 1926, p. 101), refers to the sheet that forms in coastal zones during winter, and that is rendered immobile by attachment through the ice foot to the shore. North polar cap ice is the ice which throughout the year covers the deep central, major portions, of the north polar basin. It is chiefly distinguished by great solidity; by the great size of its fields; and by the massiveness of its rafted hummocks. Pack ice consists mostly of fast ice that has broken away, supplemented in certain regions by minor additions from the polar cap ice. It is found around the outer borders of the Arctic Ocean; dominates its marginal seas, the sounds of the American Archipelago, Baffin Bay, Davis Strait, and the waters of east Greenland.

The annual cycle which witnesses the formation or depletion, as the case may be, in the varying areas occupied by the three types—fast, polar cap, and pack—displays the following features. We start with the formation of fast ice in winter, as well as the solidification of and accretion to both the already existing pack ice and the polar cap. The approach of spring and summer causes the fast ice to



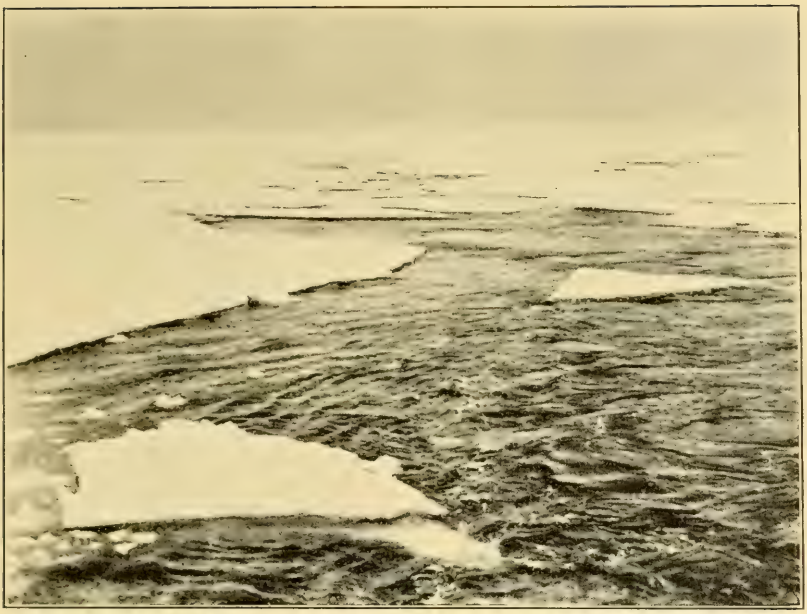
A CRACK IN PACK ICE

FIGURE 7.—Winds and currents often open cracks in the pack ice such as shown here. A crack differs from a lead in respect to its width, a crack not being of sufficient width to permit navigation. A crack is the antithesis of a pressure ridge. (Photograph by H. G. Ponting in the Scientific Reports of the British Antarctic Expedition.)



A LEAD

FIGURE 8.—A lead in pack ice. Ships in the pack are conned by the ice navigator from a station aloft. This is the best vantage point to study the system of leads and the location of open water. (Photograph by H. G. Ponting in the Scientific Reports of the British Antarctic Expedition.)



A POLYNYA

FIGURE 9.—A small polynya in pack ice. From its small extent it is sometimes referred to as a pool. (Photograph by H. G. Ponting in the Scientific Reports of the British Antarctic Expedition.)

FROST SMOKE

FIGURE 10.—A polynya from which frost smoke is rising. The famous "North Water" in the northern part of Baffin Bay remains open throughout many winters despite the very low temperatures and the presence of surrounding ice. (Photograph by H. G. Ponting in the Scientific Reports of the British Antarctic Expedition.)

break up and partly melt, with the surviving portions augmenting the still larger areas of pack ice. Narrow margins of fast ice, however, may remain unbroken throughout a summer or perhaps for several summers, in which case, of course, they maintain their identity

THE REGIONAL DISTRIBUTION OF SEA ICE AND LAND ICE

FIGURE 11.—North polar cap ice. Pack ice. Fast ice. The polar cap ice is the hub of all northern sea ice. Glacial ice. The largest areas of land ice reaching sea level are: Greenland, Ellesmere Land, Baffin Land, Melville Island, Patrick Island, Spitsbergen, Franz Josef Land, and Nicholas II Land.

for that particular year. Pack ice has three methods of disposition in the Arctic Ocean, (*a*) it rafts and solidifies, inserting itself as a part and parcel of the heavy polar cap, or (*b*), it joins one of the ice streams which are always slowly moving, with little interruption, out

of the north into lower latitudes, or (*c*), it melts and disintegrates during summer near, or at, its source. Polar cap ice comes under the influence of powerful conflicting forces which causes it sometimes to persist for four or five years, and possibly longer, before leaving the polar basin.

Three Forms of Sea Ice

FAST ICE

The ice cover of northern seas thousands of years ago probably had its genesis in fast ice fringing polar shores. Once the nucleus of the permanent cap was built from the dismemberment and offshore scattering of the fast-ice belt—and survived the first few summers—the latter lost much of its geographical importance. The wide distribution of fast ice in coastal waters, however, serves to make it representative for such zones.

The initial appearance of slush and sludge takes place in water lanes of the Arctic early in September, first in the sounds of the American Archipelago, then in the marginal Arctic seas, and in Baffin Bay in swift succession. Small pancakes soon cement, rim to rim, in a continuous sheet of young ice and preparations are made by man and beast to meet the rapidly approaching grip of winter. Hudson Bay and Fox Channel in the American sector see their first fast ice during October, with freezing spreading to Newfoundland and to the Gulf of St. Lawrence late that same month, or early in November.

Fast ice continues to "make" until it reaches a maximum area in December after which, until May, it builds very little farther out from the coast, but continues steadily to increase in thickness.

The regions of growth for fast ice largely reflect bathymetrical conditions. Other important influences are the degree and duration of low air temperatures; the decrease of salinity due to snow melting; to river discharge and to precipitation; the amount of storminess; and the presence of grounded hummocks, icebergs, and floes of pack ice.

The seats of fast ice are the broad continental shelves and the flat spacious embayments. The most striking example of such a region is the remarkably wide Siberian shelf which has a mean width of 400 miles and a depth of 12 to 50 fathoms, its outer edge forming a "steep" slope facing the polar ocean.[13] These regions produce a vast amount of fast ice (see fig. 11, p. 20) because (*a*) the shallow depths favor early chilling, and (*b*), the sea freezes more rapidly where it has been diluted by Siberian rivers.[14] It is estimated that approximately 150,000 square miles of fast ice form every winter from Wrangel Island to Northern (Nicholas II) Land. It has an average thickness of 6½ feet, ranging all the way from 0 to 9 feet.

Another large area of fast ice, second only to Siberia, is the sheet covering the labyrinthlike waterways of the American Arctic Archi-

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[13] Transehe (1928, p. 102) states that fast ice extends about 275 miles out from the coast opposite the Yana River, longitude 135° E.

[14] Wiese (1922, p. 271) calls the Siberian shelf and the fringing Arctic seas "The factory of northern polar ice." According to his theory the air temperatures along the Siberian coast are an index of the size of the Greenland Sea pack 4½ years later. This interval represents approximately the time required for a piece of ice to drift from Arctic shores to the Greenland Sea, and a high correlation of 0.83 appears to support such a contention.

pelago. While small openings over some of the deeper channels and straits may remain clear for long or short periods, the proximity of the large number of islands, combined with their irregular outlines, promotes a maximum amount of freezing. Solar warming of these same land masses during summer, on the other hand, tends to accelerate melting and disintegration of the ice, but summer in the Arctic is a brief period, and practically three-fourths of the year the temperature is below freezing. The ice is held fast in the archipelago region longer than in many localities by the narrow constrictions, the straits, and the sounds. Kane Basin and Smith Sound, waterways separating Ellesmere Land from Greenland, freeze fast, shore to shore, every winter. The southern limit of this solid sheet forms a bridge opposite Etah over which the Eskimos cross to hunt caribou in Ellesmere Land.

Melville Bay in winter is normally covered by fast ice, its seaward edge extending in a curve from Cape York to Wilcox Head, a distance of 800 miles, and out from the Greenland coast 30 to 40 miles. This ice sheet exerts a very important influence on the progress of icebergs out of Baffin Bay to the Atlantic when it often seals the inner part of Melville Bay (see fig. 43, p. 85) imprisoning not only the bergs calved from local glaciers but also thousands of others en route. Melville Bay fast ice in this manner may interrupt the berg supply to the Atlantic for one year and possibly for several years. Hudson Bay is fringed every winter by a belt of fast ice, but rarely if ever are its central portions completely frozen over. Hudson Strait in winter exhibits a fringe of fast ice 5 or 6 miles in width except that opposite steep cliffs, where the water is deep and remains open throughout most winters.

Fox Channel, situated immediately west of Baffin Land, is a region believed to be prolific in the production of fast ice. This estuary is broad and flat and noted for the great daily rise and fall of the tide. In many shallow bays the fast ice by resting on the bottom is broken up and therefore is more easily carried away by the currents. As soon as the floes move offshore the water is uncovered to freeze again, and in this manner large quantities of pack ice are believed to be manufactured there during winter.

Fast ice forms in the vicinity of MacKenzie Bay, Alaska, as early as August 15, and at Point Barrow it bridges from the shore to the pack by the last of September. The large fast ice area off the mouth of the MacKenzie and the rapidity with which it disappears on the arrival of the spring freshets has long been known to the whalers.

Fast ice, according to Helland-Hansen and Nansen (1909, p. 307), extends considerable distances into the Greenland Sea during severe winters but our information of this particular region, due to the great hazards attending wintertime navigation, is meager.

Varying meteorological conditions cause similar wide variations from year to year in the area and thickness of fast ice. Under this heading may be listed the duration and degree of low air temperatures; precipitation in the form of rain or snow; the prevailing atmospheric circulation; and the amount of storminess. The maximum production of fast ice takes place during cold quiet winters, while greater areas of water remain open from November to April during windy years. Storminess and rough water associated with

warm winters not only prevents the formation of the sheets but also continually break and loosen them from the shore. Melville Bay remained well covered with fast ice during the years 1915 and 1916 largely on account of favorable weather conditions. If a cold, quiet winter be preceded by an unusually wet autumn, conditions are ideal for the production of a maximum amount of fast ice. Copious precipitation not only freshens the surface layers, thereby raising the freezing point, but it also establishes a very pronounced state of stratification in the surface layers causing ice to form as in a saucer. No correlations have ever been published, to our knowledge, on the relation between rainfall and fast ice, but it is logical to assume that the fresher are the surface layers, the more swiftly will grow the ice.

AN ICE FOOT AT LOW TIDE

FIGURE 12.—The ice foot is defined as the ice formation which builds up along the water lines of the northern shores during the colder months of the year. (Photograph by C. Wagner.)

The outer limit of fast ice is usually determined by waves in the sub-Arctic and by buffeting and frictional pressure in the Arctic. Transehe (1928, p. 112) in this connection, has called attention to the important rôle played by the rafted masses of sea ice called "stamukhi." These old grounded hummocks with a draft of about 12 fathoms, which become strewn along the slopes of the Arctic basin during the summer, serve in winter as a seaward rampart protecting the fast ice from destruction. The outer bounds of the fast ice in the polar sea, therefore, coincide quite closely with the 12-fathom bathymetrical contour. The shoreward border of fast ice which builds out from the land, unmoved by tides or waves, is called the ice foot. The beaches, ledges, and cliffs of northern lands are naturally chilled to very low temperatures during winter, causing the

ice line to climb as high as the spray dashes and to penetrate down as far as the frigid temperature of the land is below freezing. This thickening of the sheet combined with the rise and fall of its seaward side, causes it to fracture first in one tidal crack running adjacent and parallel to the shore and later in several.

Fast ice not only "makes" first in autumn but it also is the earliest of all sea-ice forms to melt. Beginning in the early part of spring in the latitudes of Newfoundland, and sweeping northward with the advancing sun, disintegration reaches along Arctic shores during May. The land snow, in melting, flows out on top of the fast ice collecting in large pools and lagoons of "offshore water." The fast ice platform, for a few weeks, holds tightly until the heat from the sun penetrating downward opens up the floes, and allows these pools to drain away. Then sand and detritus from the land now bare, honeycomb the ice. Soon the sheet gains a slight movement along the coast, uncovering a lane of open water which persists there for the entire summer. The so-called famous Siberian Sea Road of the Russians, has this origin.

Fast ice is of particular importance as it exerts a direct influence upon the rate of production of icebergs, and also upon their subsequent movement once they have calved from the glaciers. The effect runs all the way from entirely preventing production to blocking, temporarily, the passage of the bergs toward the open sea. Several fjords in northern Greenland, Frederick Hyde, and Bessels, for example, are covered between the glacier front and the fjord entrance by a very old, heavy ice which Koch (1926, p. 100) calls "sikussak," which effectually prevents the glaciers from ever discharging any bergs into the fjord. The word "sikussak" is Eskimo, meaning "very old ice," and was first used by Rasmussen (1915, p. 358) for sea ice which after two to five years has gradually become fresh throughout, and roughly granular in structure, so that it is indistinguishable from glacier ice. Sikussak has a very limited geographical distribution and is found only in calm, undisturbed fjords of the north Greenland type. It is at least 25 years old, therefore the oldest fast ice known.

Fast ice which has not acquired the age of sikussak may lie in front of glaciers, penning up the icebergs for an indefinite period. (See fig. 39, p. 80.) The Great Humboldt Glacier in northwest Greenland experiences such a blockade which may last 20 to 30 years, or which may be broken by an especially favorable summer when the accumulation of hundreds, possibly thousands, of bergs is freed to drift southward. Some of the rich iceberg years in the North Atlantic may reflect some of these events.

Along other coast lines and in other ice fjords fast ice may blanket the bergs for only a part of the year. Thus the iceberg fjords of West Greenland, particularly those in Disko and Northeast Bays that supply so much of the North Atlantic quota, are normally covered by fast ice only during the colder months, from November to June. When the ice breaks away, usually in May or June, the winter's collection of bergs is released to float out into Disko Bay. Fast ice, therefore, causes a seasonal pulse in the presence of icebergs in Davis Strait where otherwise the regular flow from the ice cap itself would cause a more or less steady distribution the year round.

NORTH POLAR CAP ICE

Polar cap ice is the end product of all the forces which develop a very strong, massive ice cover in the central north polar sea. This covering constitutes about 70 per cent of the polar basin; approximately 2,000,000 square miles. Fields and floes break away from the main polar core, either to mix with the inshore ice over the continental shelves, or to be discharged through one of the several ocean straits. The great sea-ice cap, upon closer examination (see fig. 11, p. 20), has the same general shape as the polar basin, with its margin closely paralleling the course of the 1,000-meter isobath. Like the deeper part of the polar basin, the elliptical-shaped cap lies much closer to the Greenland-North American side than it does to Europe and Asia, with its long axis running from Spitsbergen to Point Barrow, its center, often called the pole of inaccessibility (see fig. 14, p. 27), offset about 400 miles toward Alaska.

POLAR CAP ICE

FIGURE 13.—The polar cap ice at the North Pole on May 12, 1926. In summer, small polynyas are stated to be found every 5 to 10 miles. (Photograph taken from the dirigible *Norge* by L. Ellesworth.)

This hub of the permanent sea ice is continually fed by pack ice around its periphery, and reinforced by accretional freezing and snowfall, all at a rate during the last century at least, equal to the combined processes of melting, crushing, and discharge through the various exits. Freezing on the underside of the cap progresses throughout nine months of the year, most rapidly during the winter. Examination shows that the underside of an ice sheet is often rough and brush-like in character. This is the transition zone between ice and water. The thickness of the solid sheet depends mainly upon two factors, viz, the condition of heat and the magnitude of the forces that raft and pile up the ice. Sea water solidifies in the Arctic Ocean to an average depth of 6½ feet.[15] But winds and cur-

[15] The maximum thickness of hummocked ice observed by Nansen in the Arctic Ocean was 11 feet 10 inches, but he sets the mark of 8 feet as the estimated mean thickness of the entire cover. Stefansson (1922), north of Banks Land, reports the thickness of old fields, free of pressure ridges and rafting, as 12 to 14 feet.

rents may telescope the fields and floes, layer upon layer, to a height of 30 to 40 feet, and to depths of 100 to 200 feet.

Our knowledge regarding the movement and behavior of the great polar cap, is confined, with few exceptions, to observations along its frontier from the following sources: (*a*) From the drift of ships which have been imprisoned in the margin: (*b*) from a few extended sledge excursions of polar explorers: and (*c*) from the records of drift objects released at various points. These are tabulated below:

(*a*) Ship:	Drift
Jeannette (DeLong)	750 miles. (From the vicinity of Wrangel Island westerly to 80 miles northeast of Bennett Island.)
Fram (Nansen)	1,400 miles. (From north of the New Siberian Islands westerly to north of Spitsbergen.)
Tegetthoff (Weyprecht)	250 miles. (Ice Cape, Novaya Zemlya westerly to southeast coast of Franz Josef Land.)
St. Anna (Brusilov)	850 miles. (Southern part of Kara Sea via its eastern side to north of Ice Cape, Novaya Zemlya.)
Young Phoenix	375 miles. (From Point Barrow easterly to Return Reef, thence westerly past Point Barrow, and finally disappeared in the northwest.)
Karluk (Bartlett)	500 miles. (From Point Barrow westerly to 75 miles north of Herald Island.)
Maud (Amundsen)	750 miles. (From 125 miles northwest of Wrangel Island westerly to 50 miles north of New Siberian Islands.)

(*b*) Sledging journeys out on to the polar cap ice from their very nature do not give as good opportunities for observation on the movement of the ice as do the more accurately measured tracks of imprisoned vessels. Peary, on dashes to the northward of Ellesmere Land toward and to the pole, believed that the pack outside the continental edge was sliding eastward. Cagni, pushing poleward from Franz Josef Land was carried steadily toward the west. Storkerson and Wilkins on the opposite side of the polar sea also experienced a westerly drift.

(*c*) In 1898 the Geographical Society of Philadelphia supported a project to release a number of drift casks or buoys at various points north of Berling Strait in order to learn the set and drift of the ice. Three of the buoys, which we will call "a," "b," and "c," were recovered as follows:

Cask "a."—1,400 miles. Set adrift August 21, 1901, in 72° 18′ north, 175° 10′ west, about 85 miles northeast of Wrangel Island, was recovered August 17, 1902, near the mouth of Kolyuchin Bay on the Siberian coast.

Cask "b."—3,500 miles. Set adrift September 13, 1899, on the pack ice west-northwest of Point Barrow in 71° 53′ north, 164° 50′ west, was recovered June 7, 1905, 1 mile east of Cape Rauda Nupr on the northern coast of Iceland.

Cask "c."—3,500 miles. Set adrift July 24, 1900, at Cape Bathurst in 71° 00′ north, 128° 05′ west, was recovered November 3, 1908, on Storo Island, Finmarken, Norway.

Wreckage from the *Jeannette* drifted from the vicinity of Bennett Island, to Julianehaab, Greenland, a distance of about 3,600 miles. Siberian tree trunks and other objects of Asiatic origin are quite

THE RECORDED DRIFTS OF BESET SHIPS, FLOE PARTIES, AND WRECKAGE IN THE ARCTIC

FIGURE 14.—1. Wreckage of whaleship *Dauntless*, 1817. 2. Floe party from German polar expedition, 1869–70. 3. The Dutch whaling fleet, 1777. 4. Wreckage of the *Jeannette*, 1879–1881. 5, 6, and 7. Casks released near Point Barrow, 1898. 8. The *Tegetthoff*. 9. The *St. Anna*. 10. The *Fram*. 11. The *Karluk*, 1913–14. 12. The *Young-Phoenix*. 13. The *Fox*, 1858. 14. The *Polaris* floe party, 1871–1873. 15. The Resolute, 1854. 16. A tool box incased in ice. The outline of the central area of polar cap ice is obtained by recording the distance to which vessels have forced their way toward the pole. The pentagonal figure located between the geographical pole and Alaska is known as the ice pole or pole of inaccessibility.

frequently washed ashore along the eastern and southern parts of the Greenland coast.

The total amount of observational data collected to date is so small that no complete and reliable account can yet be given of the movement of the ice in the polar cap. Nearly all of the data, as can be seen, pertain to the Siberian sector, and practically no satisfactory information is available for the American side of the great interior. The known indraft of water through Bering Strait and the well-recognized discharge of ice through Greenland Strait, when considered in conjunction with the above records of drifts, just mentioned, definitely establish, however, a westerly movement of the ice on the Siberian side. Since the course of the ice can be traced with reasonable assurance from Point Barrow around the Siberian continental shelf, and finally out into the Greenland Sea, it is logical to believe that the American margin assists in feeding what would otherwise become ice-deficient regions immediately to the west. All of the drifts that have been recorded were from east toward west, except Peary's, and the route over which he traveled toward the pole undoubtedly lay within the area of active drainage to the North Atlantic. It seems, therefore, difficult to escape the conclusion that the outer margin of the polar cap ice, on the Siberian side at least, participates in a slow, but definite, anticyclonic movement.[16]

Cum sole motion in the Northern Hemisphere is, however, not in harmony with the theory which, as Nansen points out, holds that gradient currents flow as a rule with the land on their right hand. Why should the Arctic Ocean ice drift in a direction opposite to the current? The drift of flat ice, as proved by observations in many parts of the world is well known to be largely controlled frictionally by the wind and in a region of weak gradient currents, such as the north polar basin; the wind is probably dominant.[17]

Two different types of circulation, belonging to the planetary wind system, affect north polar regions. A central polar dome of residual high atmospheric pressure causes anticyclonic winds around the pole. But the southern margin of the Arctic extends into the belt of cyclonic westerlies, and as modified by the distribution of land and water and by the seasonal cycle, these prevailing conditions are often interrupted. The polar cap of high atmospheric pressure must tend to give a westerly and northwesterly drift to the ice, and the subarctic westerlies would tend to impart an east and southeasterly component to the fields while the congestion in various parts of the cover, as noted by Sverdrup, must exert a third modifying effect. Until we secure a greater amount of meteorological, ice, and oceanographical data from these regions we shall be unable to state conclusively the actual movement of the polar ice cap and to explain its causes.

The accepted views that the general drift of the ice is anticyclonic does not necessarily contradict the theory of gradient currents for the Northern Hemisphere because the winds or various conditions within the pack itself may constitute the deciding factor. All the

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[16] We should not lose sight of Amundsen's unsuccessful attempt to be carried from east to west across the polar basin in the *Maud*, 1918–1925. The early part of the drift began auspiciously and the vessel was carried as far as the New Siberian Islands, but whether owing to unfavorable local conditions, or actually to some temporary suspension of the normal east to west movement of the ice, no further headway was made.

[17] On the Siberian shelf the winds, according to Sverdrup (1928, p. 45) are the main control, with the ice drifting approximately 33° to the right of the wind. These conclusions are based on a wealth of observational data obtained during the drift of the *Maud* with the recorded winds in the Siberian sector. (See Sverdrup 1928, 1929.)

current data and drift records, except the *Fram's*, so far secured in polar regions have been confined to the shelves inside the continental edge. The circulation of shelf water, as confirmed by studies off Newfoundland (Smith, 1924a) are controlled by the wind, by the tidal currents, and by other dynamic forces in irregular succession. Gradient currents, on the other hand, consistently hug the steepest part of the continental slopes, and their influence, we have found, ceases with astonishing abruptness as shallow depths are entered. Since the total collection of data to date, except the *Fram's*, has been made on coastal shelves, they furnish little or no evidence for or against the presence of a gradient current of cyclonic direction in the north polar sea. On the other hand, we have every reason to believe, that with the large amount of land drainage from Siberia, such a current is developed there. The fact that the *Fram* remained on the edge of the continental slope—the seat of gradient currents—at only two points and briefly, weakens the evidence favoring the nonexistence of a current. It is of interest to note that the two points where the *Fram* approached nearest the slope, viz. (*a*) into the basin north of the New Siberian Islands, and (*b*) north of Franz Josef Land, coincide with the two most tortuous parts of her drift. These irregularities may indicate the effect of the gradient current, which is probably not strong enough to overcome the wind-driven movements of the pack.

The main escape of polar cap ice toward the Atlantic is through the opening between Spitsbergen and Greenland. Ice streams much more attenuated and distinctly secondary in size emerge around the southern side of Spitsbergern; through the Ellesmere-Greenland Strait to Baffin Bay; and by still more minor passages through the sounds of the American Archipelago. The aggregate volume of these several discharges and their relative proportions are still matters of conjecture. Observations on the rate of drift of ships and other objects together with a consideration of the age of the ice, leads to an estimate of four to five years for the average period that a given sample of cap ice remains in the Arctic Ocean.

Few who have visited the north polar regions can fail to appreciate the great magnitude of the forces that are constantly at work, forcing the cap together in some places but rending it apart in others.[18] The momentum often attained by the fields, hundreds of square miles in area, driven forward by a gale is great. Small wonder that in meeting the edges are tossed high aloft by the impact, and that along the line of collision veritable embattlements are formed in a confusion of ribs and ridges. The rugged features found in one sector, contrasted with the flat sheets prevailing in another, mirror the effects of wind and current. Whether the polar cap ice suffers its greatest dynamic deformation in the winter when the ice is knitted the tightest, or during the autumn when it is loosest, is an open question.[19]

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[18] Makarov (1901) has carried out some very interesting investigations on the state of concentration of the cover and estimates that approximately 10 per cent is continually in the process of opening, closing, or freezing.

[19] Transehe (1928, p. 97) states that the hummocked, telescoped condition is best developed in late summer and autumn. Open water is most plentiful then, permitting the fields to travel across leads with maximum momentum. It is claimed that during winter and spring the battering and buffeting diminishes on account of the close, tightly knit condition of the cover. Sverdrup (1928, p. 106) regards the subject in a different light, pointing out that when the ice cover is open it affords one floe which is being struck an opportunity to give way to the next floe, and so on, the ensuing shocks being proportionately reduced.

As a result of the prevailing winds, currents, and of the configuration of the basin, certain parts of the polar cap ice are characteristically congested. These are known as the regions of paleocrystic ice. In this respect the most famous district is north of Greenland and Grant Land, while other sections are along the east coast of Novaya Zemlya, around Franz Josef Land, and off the southeastern coast of the Beaufort Sea. The term "paleocrystic ice" was first used by Nares in 1876 to describe the tumbled, chaotic mass of blocks and domes of old sea ice, developed after many years of shock and pressure. A well-known area of paleocrystic ice is the sector on the northwest coast of Greenland, where the polar cap ice is forced through the funnellike opening in Baffin Bay.

Paleocrystic ice lying across the so-called American route to the pole proved a great obstruction to the early northern explorers.

A PALEOCRYSTIC ICE REGION

FIGURE 15.—The horizontal fields of cap ice in certain regions undergo tremendous shocks and pressures due to the prevailing winds, currents, and the configuration of the basin. The resulting form is called paleocrystic ice. This photograph was taken in April, 1902, north of Cape Helca, Grant Land, the northern extremity of Ellesmere Land. (See fig. 11.) (Photographed by Rear Admiral R. E. Peary.)

Sledge progress across such upheaved scarps is snaillike, full of toil, and fraught with danger, both to dogs and men. After several years of pressure ridging and hummocking with successive layers of snow increasing the bulk, floebergs develop. These when floating, attain heights as great as 30 feet above the sea, and they often show a striking resemblance to icebergs. Several Arctic explorers have mistaken floebergs for icebergs; even Peary, seeing them north of Grant Land, thought he had evidence of undiscovered glaciation in that direction. It is an interesting question whether or not many floebergs actually join the supply of icebergs drifting into the North Atlantic. Information on this subject is negligible because the proper place to distinguish the one ice form from the other and to determine their relative numbers is at the head of Baffin Bay, where few or no observations have ever been taken. The presence of floe-

bergs, however, at the extreme northern end of the south flowing current places them in a favorable position eventually to drift to Newfoundland.

In contrast to the regions of hummocks and of pressure ridges, large areas of open water, known as polynyas, also form within the

THE DIFFERENT TYPES OF SEA ICE IN THE NORTH GREENLAND SECTOR

Fig. 16. -, sea-ice discharged into the Greenland Sea. , sea-ice, congested in the paleocrystic region. , paleocrystic area in some years. , fast ice. , many years old ice. , Sikussak.---, Peary's "Big Lane" marks the line of shearing in the sea-ice cover. (Map after Koch, 1928).

polar cap ice; these tend to occupy certain characteristic positions in the polar basin, two regions now definitely established being:

(*a*) A belt, or belts, several hundred miles long, parallel and just northward of the New Siberian Islands.

(*b*) Peary's "Big Lead," north of Grant Land and Greenland.

The fact that polynyas are best developed off the New Siberian Islands in winter when high atmospheric pressure prevails over

Siberia indicates that the wind is a major factor in keeping the polar cap ice in motion.

The second famous polynya was discovered and described by Peary (1907, p. 97) as the "Big Lead" (see fig. 16, p. 31). It coincides in position very closely with the continental edge running along the eighty-fourth parallel from Grant Land to Cape Bridgeman, Greenland. The "Big Lead" is smaller than the Siberian polynya, being

A FLOEBERG

FIGURE 17.—A floeberg (see p. 30) in the northern part of the Hall Basin between Ellesmere Land and Greenland. Its geographical position indicates its origin as an old pressure ridge. Floebergs, although from an entirely different source, are often mistaken for icebergs. Some of the bergs drifting in the North Atlantic may well be floebergs from the polar basin. (Photograph by L. Koch, 1929.)

seldom over a mile or more in width yet having an estimated length of at least 300 miles.[20]

PACK ICE

Although seamen usually refer to any flat ice drifting at sea as "field ice," such a practice is not recommended because it tends to confuse one not thoroughly familiar with ice terms. It is better to follow the classification of Priestley (1922, p. 393), who defines pack ice as any sea ice which has drifted from its original position under the influence of winds and currents. The word "pack" does not necessarily mean the ice is tightly packed together, because the pack may have large or small areas of open water, as the case may be, within its bounds. Its northern sources are the fast ice and the

[20] Koch (1926, p. 102) is of the opinion that the "Big Lead" marks the line of shearing of the ice cover while the northward part, out in the deep polar basin, drifts eastward the southern side does not, partly because there is no definite gradient current in over the shelf, but chiefly because it is prevented by the promontory formed by the north Greenland coast. This theory is supported not only by the jamming of the ice between Greenland and Grant Land, but also by a marked change in the age of the material. Thus old ice alone was observed by him south of the "Big Lead." Koch also observed a westerly current running in between the ice floes along the northern coast of Greenland west of Cape Bridgeman.

reworked, possibly several years old pack that has survived several summers.[21] In the north polar ocean the pack ice occupies a belt intermediate between the fast ice and the polar cap ice. The pack on its offshore side may add to the polar cap or it may remain in the intermediate zone, or it may freeze again in the region of fast ice. Outside the polar basin, however, pack ice never builds into heavier forms, but either drifts southward to melt or remains in the north to survive longer. Pack ice is distinguished from the polar cap ice by its lightness.

Outside the Arctic Ocean the qualities that distinguish pack from polar cap ice are not so apparent and the farther south we go the

SUMMER IN THE NORTH POLAR BASIN

FIGURE 18.—An old field of pack ice in the polar regions showing the disentegrative effects of the heat that comes from the summer sun. Note the hummocked moutoneé contour of the ice surface and also the fresh-water pools in many of the depressions. (Photograph from the Russian hydrographical expedition, 1910–1915.)

smaller is the difference. Pack ice is distributed from the Arctic to lower latitudes in two main streams which reflect not only the ocean currents but also the general trend of the coastal shelves. The fact that these shallow waters are more effectively chilled than deep permits the ice to survive instead of melting near its sources. The

[21] A source of limited supply, not often mentioned, is anchor or bottom ice. As its name implies, it forms on the bottom in those regions where swift-running frigid currents prevail, and at depths seldom exceeding 40 to 45 feet. During clear cold winter nights radiation causes a loss of heat even from the bottom, and anchor ice will form. The first rays of the morning sun loosens this, allowing it to rise to the surface. Barnes (1928, p. 105) states that thousands of tons of this type are added each year to the Gulf of St. Lawrence, and similar contributions must be significant in all northern seas. Rodman (1890, p. 16) relates an instance where anchor ice brought a tool box to the surface in the vicinity of Nain, Labrador, the box being recognized as one belonging to a ship lost years before in Hudson Strait, several hundred miles to the northward (see fig. 14, p. 27).

bathymetrical influence, masked the greater part of the year in polar seas, becomes more noticeable further south, where in temperate latitudes at the end of winter the boundaries between the warm deep ocean and the cold shallow shelves are very clearly marked.

Most all pack-ice streams show the following features in cross section: (1) An outer zone of scattered loose glaçons; (2) a mid zone of heavier floes more compact but with occasional cracks and leads; and (3) an inner heavy band, possibly polar cap ice, pressed closely against the shore ice. Offshore winds broaden the stream, and scatter the outer floes, while on-shore breezes narrow the ice and pack it against the coast. In summer a lane of open water often develops adjacent to the shore, the warm drainage from the land overflowing along the coast breaks up the fast ice and speeds the forward movement of the pack.

The ice streams are most voluminous and extend farthest southward during spring or early summer and shrink toward their sources during late summer or early fall. This recession is due not only to the heat of summer melting the ice but also the absence of the favorable winds and the strong currents which prevail in spring. Since the supply of pack does not immediately increase on the resumption of freezing air temperatures, due to the specific heat of the water, there is consequently a lag in the swelling of the ice streams, and this interval marks the minimum of pack. The seasonal variations in the limits of pack ice are also dependent upon the amount of fast ice that is contributed.

Pack ice invades the North Atlantic along two main routes, (*a*) along the eastern side of Greenland and (*b*) along the eastern side of North America. (See fig. 11, p. 20.) The East Greenland ice stream in its upper reaches is split by Spitsbergen. Its main trunk, bearing the heaviest of all sea-ice forms, pours directly through the Greenland Sea, while an eastern arm from the cachments of the Barents and Kara Seas moves toward Greenland roughly along the seventy-fifth parallel. The ice stream to the western Atlantic is fed through the tortuous waterways of the Arctic Archipelago, which not only lengthen the journey but materially reduce the volume of contribution. Neither ice stream depends solely on its Arctic Ocean connections for its supply, as two of the most prolific regions of ice production are in the Greenland Sea on the east and Baffin Bay on the west.

The following table gives the approximate length, velocity, etc., along the two main ice paths to the North Atlantic:

Ice stream	From—	To—	Distance, miles	Velocity, miles per day	Approximate time in months
East Greenland pack	75° N., 0° W.	62° N., 51° W.	1,850	7.5	8½
Eastern North American pack	74° N., 70° W.	45° N., 49° W.	1,960	12.5	5½

The Spitsbergen Pack

Spitsbergen marks the transition between Arctic and Atlantic influences—to the northeast intense Arctic cold prevails, while on the southwest the coast is warmed by the Gulf stream drift. This

warm current makes accessible the harbors of the west coast for a period of about four of the warmest months of the year. Spitsbergen pack refers to the ice which in the winter moves westward past the south cape of Spitsbergen, and which generally blocks the northeastern quadrant throughout the summer. Such of the pack as lies southward of a line Spitsbergen-Franz Josef Land to Nicholas II Land, including the Kara Sea, and east to Cape Chelyuskin, drains into the Atlantic through the northern part of the Barents Sea and south past Spitsbergen. During winter and early spring the Spitsbergen pack is swollen by the break-up of fields in the Barents Sea and is, therefore, at its flood. During this season the pack may spread so far to the south as to inclose Bear Island for a month or more at a time. But the continuity of the Spitsbergen pack is always threatened by the inthrust of warm waters from the Atlantic. The first encroachment of spring severs the pack and forms open water in the offing of the west coast. The ice in Barents Sea with the progress of spring and summer retreats steadily northward until it persists only in the shelter to the northeast of Spitsbergen. In this last region, on the line of conflict of such opposing forces, ice conditions are subject to wide fluctuations. Despite the warm current the great productivity of the area to the eastward guarantees in most years a generous supply of ice to the great east Greenland pack. Those particularly interested in the Spitsbergen area are referred to Hoel (1929); Iversen (1927); Kolchak (1909); and Makarov (1901).

The East Greenland Pack

Pack ice is seldom, if ever, absent from the waters of northeast Greenland and the Greenland Sea. The east Greenland pack is fed by (*a*) the direct discharge from the Arctic Ocean; (*b*) by the ice from the Barents and Kara Seas; (*c*) by winter ice formed in the Greenland Sea; and (*d*) by fast ice made locally along the coast. The ice from all these separate regions is alike in general character and appearance, except that from the polar basin which, as already described, is easily distinguishable. This old, heavy ocean pack fills the northwestern sector of the Greenland Sea but the major portion of the covering of the latter consists of younger, lighter fields.

Winter witnesses the influx of heavy pans and floes into the Greenland Sea reinforced by great quantities of ice formed locally. This accumulation spreads gradually southward along the coast of east Greenland initiating in successive months the beginning of the ice season. Throughout the winter and spring large masses of the ordinary pack, together with some of the Arctic Ocean type, continue to push southward, and to spread away from the coast. Denmark Strait is normally more or less choked in spring, while in a bad year the ice completely encircles the northern coast of Iceland at that season. The boundary of the ice cover in spring displays a characteristic tendency to spread eastward immediately north of Iceland where it probably comes under the control of the east Iceland current. An equally impressive feature is a V-shaped retrenchment immediately west of Spitsbergen, an unmistakable effect of the warm Gulf stream drift. The average outer limit of the east Greenland pack in spring runs from the embayment near

THE VARYING BOUNDS OF PACK ICE

1 FIGURE 19.—The annual maximum, minimum, and mean limits to which pack ice in substantial volume has been recorded over a period of years. The data for the pack ice in the east Greenland current has been taken from a 15-year record of the Danish Meteorological Institute. The data for the ice in the Labrador current comes from the international ice patrol and other sources.

Spitsbergen to Jan Mayen, in latitude 71° N., longitude 8° W.; thence southwestward toward the coast of Iceland, thence westward across Denmark Strait toward Angmagssalik, thence, narrowing, to Cape Farewell where the ice tends to congregate around Farewell.

The first of the east Greenland pack appears at Scorsby Sound in October, at Angmagssalik in November, and off Cape Farewell in December or January. But it seldom sets northward around the cape to blockade Julianehaab Bay until the strong northerly winds abate in April.[22] The wind is an important factor in its distribution. Thus during northerly winds a strip of open water will appear next to the coast, while the outer edge of the pack may lie 75 to 100 miles offshore. Normally, however, during the ice season the outer edge of the pack around Cape Farewell lies about 60 miles out from the coast. If the prevalent south wind is blowing, the "storis"[23] reaches Ivigtut in March, Fiskernæs in April, where in bad ice years it may reach as far north as Godthaab, 360 miles north of Cape Farewell, from May to August. The pack, in a light ice year, however, will not reach further north than Cape Desolation, 140 miles northwest of Cape Farewell. The ice tongue which so often stretches northwest from Cape Farewell tends to bend slightly away from the coast if undisturbed by the wind, and sailing directions from Ivigtut advise going north around it, unless the wind is on-shore, and coasting back inside. The pack around Cape Farewell consists of glaçons of all sizes, and also of old hummocked floes as great as 100 feet in width and 10 to 20 feet thick. In June and July when the "storis" reaches its greatest abundance around Cape Farewell the edge of the fields has been met 100 to 200 miles offshore. The best evidence of the rate of progress of the east Greenland pack is the drifts of ships which have been caught within its clutches. Such an experience befell several vessels of the Dutch whaling fleet in June, 1777, which, according to Irminger (1856, p. 36), drifted southward from 76° north, parallel to the coast through Denmark Strait at the average rate of 11 to 12 miles per day (see fig. 14, p. 27). The *Hansa*, one of the vessels of the German north polar expedition 1869–70, was crushed October 23, 1869, in latitude 70° 50′ N., well within the grip of the pack off northeast Greenland. The survivors drifted on the ice and in a whaleboat all the way down the coast, and around Cape Farewell, and finally landed the following year at Frederickshaab, in southwest Greenland. The average rate of drift was 4 to 5 miles per day. A bottle thrown overboard from a ship near Denmark Harbor, latitude 77° 12′ N., longitude 16° 00′ W., in northeast Greenland, was recovered a year later in the

[22] Brooks and Quennell (1928, p. 9) show the monthly variation in the amount of pack ice drifting past Iceland during the period 1901 to 1924:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
4.0	4.3	8.2	7.5	7.3	7.3	3.8	1.0	0.0	0.2	0.1	0.2

These figures show that the ice season at Iceland normally extends from January to July, with March, April, May, and June being the four heaviest ice months.

[23] Storis, literally "large ice," is the term used by the Danes to refer to the pack in the east Greenland current which is heavier than the floes that have formed locally.

"storis" near Godthaab, west Greenland. Nansen drifted on the pack in July, 1888, from near Angmagissalik, latitude 65° 35′ N., longitude 38° W. to 61° 35′ N., 42° W., near Cape Farewell at the rate of 24 miles per day. (See Helland-Hansen-and Nansen, 1909, p. 306.) Wreckage of the famous *Jeanette* after leaving the polar basin must have drifted southward along Greenland because several pieces have been recovered on the southwest coast. Siberian tree trunks and many other unmistakable types of oriental driftwood have been picked up along the shores of southern Greenland—additional evidence of the drift of the ice.

Upon the approach of summer the east Greenland pack recedes inversely as it advances. The southwest coast off Julianehaab is usually free from ice by early August; Cape Farewell in late August or early September; and Angmagissalik during September.[24] Scoresby Sound district is more likely to be free in late September than at any other time, but in severe ice years it may not uncover at all; or other parts of the east Greenland coast, for that matter. During late summer or fall, when the east Greeland ice pack shrinks to a minimum, open water may be found close in, or even along the coast in favorable places. Angmagissalik, in latitude 66 north, on the other hand, has occasionally been isolated by ice the entire year. The supply ship usually finds communication easiest during the months of September and October, but sometimes it has not been able to land there until early November, while in one year, Wandel (1893, p. 252) mentions that the coast around Angmagissalik was free of ice from September 10 until November 25. It is interesting to know that Angmagissalik was selected in 1894 as Denmark's chief trading post in east Greenland because the ice belt at this point is most penetrable. It is rare indeed for the pack ice to retreat as far north as the Arctic Circle, but there are records of such occurrences.

The fact that Greenland is one of the earliest discovered lands, affords opportunity to investigate possible changes that have slowly developed in the character and behavior of the drift ice. The legendary accounts of the early voyages of the Norsemen during the eleventh century suggest that Greenland waters were icier then than they are to-day. These adventurous colonizers apparently cruised directly from Iceland to Greenland in their open Viking ships and followed the coast southward to Cape Farewell on courses to-day completely blockaded. The earliest reports, in the first century of the young colony, mention good pasturage and large fine farms in Greenland, but later, conditions apparently changed for the worse and we learn about the advance of great masses of ice. Recent archeological excavations in southwest Greenland[25] have disclosed the root-entwined frozen bodies of some of these early settlers, evidence of a record of a change of climate. In the thirteenth century the slow advance of the pack is again corroborated by the southward migration of the Eskimos; the chain of evidence being traceable in the economic relation between the gradual encroachment of the pack and the consequent disappearance of seals and of man.

[24] No pack ice whatsoever was sighted around Cape Farewell by the *Marion* expedition cruising in that vicinity, September, 1928.

[25] See Hovgaard (1925, p. 614). Porsild, in discussing the archeological finds, told me that the roots were those of annuals, not perennials; therefore the evidence is not conclusive.

There has been much discussion whether any of the pack ice which drifts southward to Cape Farewell continues either toward Flemish Cap, or westward to join the eastern North American pack. Apparently no direct observations have been made to support such a conclusion. Nevertheless, statements to this general effect continue to appear in literature. The possibility that ice may journey from Cape Farewell to Labrador or to Newfoundland is worth reviewing. Chamberlin (1895, p. 653) in describing the Arctic current which curls around Cape Farewell and sets northward along the west coast, says: "Then gradually the current recurves to the west and south, and descends the Labrador coast, its burden of ice being progressively melted and dispersed." On his own voyage to Greenland, however, he left the pack at a point no farther north than the Frederickshaab Glacier in latitude 62° 30′ N. Johnston (1915, p. 40) is of the opinion that a branch of the east Greenland current invades the North Atlantic to the forty-third parallel east of Flemish Cap, and that ice sighted east of longitude 44, is from east Greenland. Jagdt (1928, p. 9) believed that parts of the east Greenland pack sometimes break away, off west Greenland and subsequently are carried over to the American side. Schott (1904, p. 305) investigating the possible relationship between the pack ice off Newfoundland and off east Greenland, found that bad ice years in the one area coincided with good ice years in the other. For example, in 1881, enormous masses of ice swelled the east Greenland pack to capacity, while at the same time Labrador saw little ice. If east Greenland ice feeds the waters of Labrador or of Newfoundland, why did those regions lack ice that year? Mecking (1906, p. 102) after an exhaustive investigation of Davis Strait, concludes that the east Greenland pack never joins the American. Jensen (1909, p. 32) on the *Tjalfe* expedition, June 5, 1908, encountered the seaward edge of the pack in longitude 56° 30′ W., on the sixtieth parallel, at a position halfway between Greenland anl Labrador. The National Geographic Magazine for November, 1925, published on a supplementary map of the Arctic regions the maximum limit of ice, indicating a complete bridging of Davis Strait. But the westernmost record in the files of the Danish Meteorological Office, however, an institution which keeps in close touch with ice conditions in Greenland, is at longitude 54° 30′ W., July, 1918, when the pack barely reached halfway across Davis Strait. Irminger (1856, p. 41), says "there does not exist even a branch of the current which runs directly from east Greenland toward the Newfoundland banks." And Wandel (1893, p. 255) in speaking of the east Greenland packs, says, "elle ne se renuit jamais avec celle du Courant du Labrador ou le Vestis, quand, par exception, le detroit de Davis est barré c'est sans aucun doute le derniere qui a eté poussé vers l'Est."

The most favorable time for pack ice to cross to the American side is in winter, i. e., at the season when it is at its minimum. And in July, when the pack is most abundant, Davis Strait waters are so warm it can not long survive there. It seems safe to conclude, therefore, that pack ice from the east Greenland current never crosses to American waters.

The pack ice in the Arctic Ocean, around Iceland, in Kara Sea, in Barents Sea, and in the Greenland Sea has been studied by

Meinardus, by Brennecke, by Wiese, and by Brooks and Quennell with the object of determining what effect variations in these ice areas have on the weather of Europe. Meinardus (1906, p. 151) compiled a table giving the deviation and severity of the pack off Iceland from 1801 to 1904, and since that time similar data have been compiled monthly by the British Meteorological Office. The basis of Meinardus's figures were the number of days that ice was sighted from the coast of Iceland—when the masses were particularly heavy the values received double weight. The investigation discloses a very clearly marked periodicity in the character of the east Greenland pack of 4½ years.[26] The annual variations in the ice off Iceland are associated with similar variations in the wind; for example, in a winter with unusually strong, fair winds more ice than normal is to be expected to drift past Iceland. The data selected by Meinardus to demonstrate this were the difference in atmospheric pressure between Stykkisholm, Iceland; and Vardo, Norway, which, if large, forecasts more ice than usual in the east Greenland current the following spring. Wiese (1924, p. 289), independently investigating the variations in ice conditions in the Barents and Kara Seas, found an exceptionally high correlation between autumn air temperatures there and the volume of pack ice along east Greenland 4½ years later—a low temperature presages much ice and vice versa. The well-marked periodicity of 4½ years is explicable when we realize that it represents the interval necessary for the ice to complete the journey to Iceland from its sources. Brooks and Quennell (1928, p. 3) have collected a long series of statistical data on sea-ice conditions in the following regions: Off Iceland; Greenland Sea; Barents Sea; Kara Sea; and Arctic Ocean. The work of these meteorologists constitutes the most thorough investigation to date on the effect of northern ice on European weather. More ice off Iceland, or in any one of these several seas than normal, causes in the same months an excess of pressure around Iceland and a deficiency of pressure from Paris to the Azores. One of the most interesting discoveries was that heavy ice conditions during spring in northern waters are liable to be followed by a deficiency of pressure the following autumn around the British Isles. The cause is in the liberation of more water than normal, by melting, to mix with the Gulf stream during the summer. The regional variations in sea temperature produce corresponding thermal variations in the atmosphere bringing stormy weather to northern Europe. It seems well established, therefore, from the foregoing that variations in the pack ice of the northeastern North Atlantic exert an important control over European weather, the effect of the ice on the atmospheric pressures for the countries north of the British Isles being stronger even than that of the Gulf stream.

The Eastern North American Pack

One of the largest streams of ice that emerges out of the north follows a path along the east side of Baffin Land, along the Labrador coast, and eventually spreads out past Newfoundland. (See fig. 19, p. 36.) The geographical positions of the North American lands and

[26] Brooks and Quennell (1928), p. 6) recalculating make this figure 4.76 years.

the bathymetrical features of the shelves are important factors in the track of this stream of pack ice. Baffin Bay, a shallow elongated basin covering 650 miles of latitude, connected with the Arctic Ocean by the narrow opening between Greenland and Ellesmere Land, and in a less direct way through the maze of Arctic sounds to the westward, is one of the chief reservoirs of the ice. Davis Strait is the bottle neck through which a great proportion of the pack emerges into the North Atlantic. Another probable source is the region of Hudson Strait and Fox Channel. The Labrador current is the great agency of transportation southward along Labrador, past Newfoundland, and over the Grand Bank. It is the low temperature of the water over the continental shelf from Baffin Land to the Grand Bank, approximately 150,000 square miles, the surface layers of which, are chilled to 0° C. (32° F.) or lower, that permit the southward drift of the pack. The extent of these frigid shelf waters and of those of Baffin Bay furnishes a clew to the aggregate annual output of sea ice. If two-thirds of the total 467,000 square miles of the ice area is normally covered to a depth of 6 feet, the eastern North American pack consists of approximately 450 to 475 cubic miles of ice yearly. (See fig. 121, p. 200.)

The several tributary sounds located on the western side of Baffin Bay and Davis Strait contribute relatively great quantities of pack to the eastern North American ice stream. During the colder months these openings supply[27] ice to the southward moving masses, but in summer their discharges create areas of open water. These channels from north to south are as follows: Smith Sound, Jones Sound, Lancaster Sound, Hudson Strait, and Strait of Belle Isle. The sounds of Baffin Bay contribute the greatest quantities of ice, while the straits to the south are responsible for the greatest amount of dissipation and wastage. Summer melting is usually viewed as a phenomenon spreading from south to north. But in case of the North American pack, account must be taken of the disruptive influences along its western flank. Thus the discharge from Hudson Strait very much hastens the dismemberment of the pack in that offing. The *Marion* expedition in 1928 had an excellent opportunity to survey the results of these disintegrations. As early as June 11 the *Godthaab* expedition arrived and found the waters open off Hudson Strait and the ice lying quite far back both to the north and to the south. Sometime shortly prior to the *Godthaab's* visit, the strong currents pouring in and out of the strait had apparently severed the ice stream and isolated the large Labrador field. On August 18, about two months later, the *Marion* expedition found the offing of the strait still clear; and the Labrador field had disappeared by that time, though the southern edge of the Baffin Land pack remained about the same as it had been a month earlier.

In 1928 the pack around Cape Dier, Baffin Land, consisting of 18,000 square miles of ice, never penetrated farther south than Cumberland Gulf after June, showing that the rate of its southward advance was offset by the rate of melting in the sun-heated waters discharged through Hudson Strait. The isolated Labrador field of

[27] Mecking (1906, pp. 22–31 and appended map) after consulting the accounts of the various early explorations of the American Archipelago has constructed a map showing the prevailing circulation. Mecking shows that practically all of the currents flow, and therefore the ice masses move, from west to east; that is, from the polar regions into Baffin Bay and Davis Strait.

some 15,000 square miles disappeared through melting and diffusion soon after it was cut off from its northern supply. The current and temperature maps of Davis Strait compiled by the *Marion* expedition indicate a communication between Atlantic water from Cape Farewell and water from Hudson Strait, and it is this mixing across the main pathway of the polar current that interrupted the ice stream sometime during May, 1928. The dissipation of ice by the active discharge through Hudson Strait exerts a heretofore unsuspected influence on the character of the east North American ice stream. The increasing warmth of summer combined with the depletion of supply are the underlying factors that prevent the march of arctic ice from continuing past Newfoundland throughout the year, but

THE EAST COAST OF ELLESMERE LAND

FIGURE 20.—One of the iciest of shores lies on the west side of Smith Sound. This is a 12-mile-wide margin of the pack photographed on August 15, 1928, off the east coast of Ellesmere Land just north of Cape Faraday. The pack is composed of ice formed both locally and that brought here from farther north by the cold current. Some summers the pack never uncovers the Ellesmere Land coast in the Smith Sound region. (Photograph by Commander E. Riis-Cartensen of the *Godthaab* expedition.)

an important accessory is the disruptive influence of Hudson Strait. Were the Baffin Land-Labrador littoral not interrupted, the amount of pack ice carried into the western Atlantic would be much greater than is actually the case.

BAFFIN BAY PACK ICE

Except for the traverses of occasional polar explorers our knowledge of the ice of Baffin Bay has largely been derived from the logs and records of the whalers. Pursued first by the Dutch, the industry received added impetus after Ross's voyage, to reach the height of its prosperity in the middle of the last century. To-day little or no whaling is done in Baffin Bay. According to the experience of the

whalers the bay is characterized by two distinct areas of sea ice, respectively called "west ice" and "middle ice" from their locations. Mecking (1906, p. 104) basing his assumptions largely upon the drift of the *Fox* describes the bay as split by two packs. The easternmost is composed largely of ice from Melville Bay and from the east side of Smith Sound. The west ice is said to come from the west side of Smith Sound, from Jones Sound, from Lancaster Sound, and from the Baffin Land coast. The "middle ice" loomed especially large in the eyes of the whalers because it often obstructed their path, and threatened their profits. The separation of the sea ice of Baffin Bay, however, into two definite sections is not borne out by a careful analysis of conditions as a whole, nor would it logically

BAFFIN BAY PACK ICE

FIGURE 21.—The eastern edge of the Baffin Land pack July 3, 1928, in latitude 67° N., longitude 58° W., 70 miles east of Cape Dier, Baffin Land. At this time the western half of the neck of Davis Strait being ice decked and the eastern half, open water, reflects the underlying circulation of these interesting waters. The fact that the ice rises above the main deck of the *Godthaab* is striking evidence of the great thickness of the pack ice in Baffin Bay. (Photograph by Commander E. Riis-Cartensen of the *Godthaab* expedition.)

result from the behavior of the ice or from the factors influencing the latter. What the whalers called west ice is the most tightly packed part of the cover, naturally to be found hugging the Baffin Land coast. Middle ice probably refers to that part of the pack that the winds and slow cyclonic circulation of the bay tend to collect in the central and even in the Melville Bay section. The designation of "middle" to the position of the pack is, moreover, somewhat accentuated by the widening of a lead of open water around the shores of Baffin Bay in late summer. The west ice represents the heavy backbone of the pack, while the middle ice is merely the outer fields subject to wider annual variations. The supply for both comes from the upper reaches of Smith Sound, from the water arms of the American Archipelago via Jones Sound, Lancaster Sound, and Eclipse

Sound, and from the fast ice formed locally around the shores of Baffin Bay. The floes converge as they feed into the narrow neck of Davis Strait, and passing out to the south, relieve the congestion in the upper waters.

If large quantities of fast ice break up in Melville Bay, and if the winds drive across additional masses, the navigation that particular spring and summer will be greatly hampered, and the only means of proceeding northward in such a year is to hug the Greenland shore to Cape York, hence to steer westward. There are records of ships which required several weeks to make the passage under conditions such as these, or even suffered the misfortune of becoming nipped in the pack.[28] If little ice is formed or if the normal amount fails to break out of Melville Bay, Smith Sound, and the Arctic Archipelago, the pack will be of small extent and the so-called North Water will enlarge. In such years whale ships have reported crossing Melville Bay in the incredibly short time of 20 hours. The pack-ice cover normally is believed to fill four-fifths of Baffin Bay, with an area about 165,000 square miles, and often it is so extensive that it reaches over to the west coast of Greenland in some places north of Davis Strait.[29] In occasional winters pack ice is said to fill Baffin Bay solidly from shore to shore.[30]

The Baffin Bay pack has its greatest extent in March and its least in August and September. In some winters the ice area may grow to a size that completely fills Baffin Bay, while in other years polynyas are numerous and extensive; for example, off Smith Sound, Jones Sound, and Lancaster Sound. Lancaster Sound is, however, occasionally frozen solidly from shore to shore,[31] but at such times even the natives deem any attempt to cross to North Devon an extremely hazardous undertaking because a sudden shift of the winds or the currents may break the bridge. The neighborhood of Cape Warrander on Lancaster Sound is said to have more open water than any other locality in Baffin Bay. But only a short distance farther west Barrow Strait becomes covered as early as September. Baffin Bay has never been crossed by sledge but many experienced explorers have held the opinion that such a feat would be possible during an exceptionally icy winter. It is of interest to learn also that the ice cover of Baffin Bay is more or less completely renewed every year.

One of the most widely discussed features of Baffin Bay is the ice-free area at its head called North Water. Coming suddenly upon this opening after a week or more of struggling through the heavy middle pack, it is not surprising that North Water has excited the curiosity and interest of explorers for two centuries. (See fig. 10, p. 19.) The earliest and still most common explanation which has now become quite firmly established in the minds of many connects North

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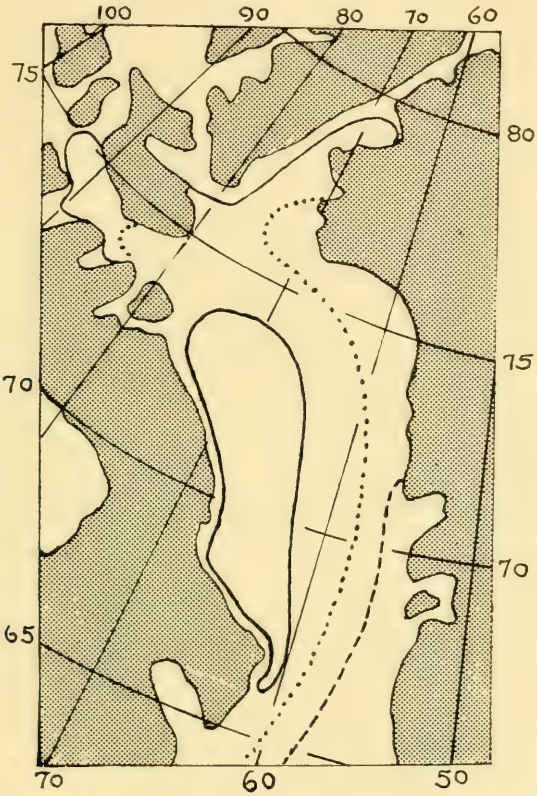
[28] The Canadian Government steamer *Beothic* struggled with ice for 20 days during the summer of 1916 on its passage from Godhavn to Cape York, but strangely enough much open water was found farther north in Smith Sound.

[29] This is the "vestis" of the Danes. Its southeastern edge reaches over to Holsteinborg in severe winters.

[30] Capt. E. Falk of the steamer *Beothic*, who has made several summer cruises into Baffin Bay during recent years, states to me that Davis Strait never freezes all the way across, but Baffin Bay does in severely cold winters. North of the seventy-fifth parallel, except for North Water and off the entrance to Jones and Lancaster Sounds, the bay freezes solidly every winter from the beginning of December to the first of June.

[31] According to a statement of Capt. E. Falk, master Canadian Government steamer *Beothic*.

Water with a warm current from the Atlantic, which, diving beneath the cold water of Davis Strait, is thought to emerge on the surface to melt the ice from the head of Baffin Bay (see Nielsen, 1928, p. 221). There is no definite evidence of such a phenomenon contained in the observations of either Nielson (1928) or Annually (1929, pp. 87–95). It seems more likely, however, that instead of a warm northward inflow, this persistent polynya in Baffin Bay is maintained by a set



PACK ICE AREAS IN BAFFIN BAY

FIGURE 22.—The dashed line represents the normal maximum limit to which pack ice extends at the end of a northern winter. The dotted line is the average July boundary of the main body of the pack. The solid line represents the normal minimum area and position to which the Baffin Bay pack ice shrinks—usually in September. Note the narrow shore lead along the Baffin Land side which isolates the pack and may be the reason for designating this area of ice as "the middle pack."

in the opposite direction. The fast ice in Smith Sound is so strong that it resists the current, but that formed just to the south is swept away, leaving open water behind it. This explanation is supported, moreover, by the recorded drifts of several ships and ice floes. Several observers in the vicinity of Etah have described looking southwestward across the zone of fast ice over the open sea. The break up of the fast ice in Smith Sound during June and July temporarily chokes North Water, but eventually the latter clears, and its area is greatest in late summer. The ice in Kane Basin, if it breaks loose

at all, does so in August, but the predominant circulation soon carries it across North Water. The immobility of the fast ice in the northern tributaries of Baffin Bay, often foreshadows a secondary maximum of pack ice to the head waters some times in late summer.

The ice cover of Baffin Bay varies greatly in size from year to year, or over a group of years.[32] During some summers the central portions are well open to navigation and reasonably safe for the metal hulls of ordinary ships of commerce. During these years the only pack is the west ice, shrunk to a narrow belt close to the shore (except for the normal shore lead), from Cape Kater to Cape Mercy. Such conditions were found in the summer of 1928 by the *Marion* expedition when the 36-mile wide pack off Cape Dier occupied a total area of only 18,000 square miles. Nevertheless the ice was heavy and thick enough to prevent the passage of the *Marion* to the coast.[33] At this time the pack consisted of glaçons 5 to 10 feet in diameter and

OPEN PACK ICE IN SUMMER—WEST SIDE OF DAVIS STRAIT

FIGURE 23.—The open summer condition of the pack off Cape Dier, Baffin Land, as found by the *Marion* expedition in August, 1928. Note the small pool of water on the glaçon in the foreground, formed by the melting of ice. Fresh water is thus always to be found during summer, even great distances at sea in the polar regions. (Official photograph, *Marion* expedition.)

of larger glaçons up to 50 to 75 feet across. The fact that the ice was quite thick, rising 2 and 3 feet above the surface and extending down 8 to 10 feet, testifies to a much more northern source. The outer edge of the pack was fairly open, but 15 miles inside, there was very little open water in which to navigate. The southern edge was met in the offing of Cumberland Gulf, latitude 64° 36′ N., longitude 59° 10′ W.

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[32] Munn (1923, p. 65) comments on the annual variations of the middle ice in Baffin Bay, where there was a very small amount in the summers of 1920 and 1921, and practically none in 1922. The three years' deficiency, Munn suggests, may have been due to an ice jam in some of the ice-choked entering sounds, viz, Jones Sound, Smith Sound, and Lancaster Sound. He claims that when such jams break away a heavy and extensive "middle pack" may be expected in Baffin Bay, and to a less extent to the southward. He is also of the opinion that a smooth, quiet summer allows the middle ice to spread farther abroad, causing a surface layer of cold water and favoring the formation of more ice freezing the following winter.

[33] The Danish Meteorological Institute (Annually, 1928, p. 3) states that pack-ice conditions in Davis Strait and Baffin Bay the summer of 1928 were favorable.

The rate of wastage of pack ice in the Arctic during the summer is well shown by the records of the *Godthaab* and the *Marion* expeditions for the summer of 1928. On July 3, the west ice of Baffin Bay extended halfway across to the Greenland coast; on August 15 it occupied only one-fourth of the strait; and on September 15 it had withdrawn from the Baffin Land coast as far as Cape Broughton, a promontory 90 miles north of Cape Dier. Again the average rate of retreat of the southern bounds of the west ice in Baffin Bay during the summer of 1928 was approximately 1 mile per day. In general the shrinkage of pack ice in the far north is a phenomenon which accelerates to a certain period during the summer, after which it is gradually retarded, until freezing begins.

The circulation of the waters of Baffin Bay and the movement of the pack is known only in a general way. The first systematic oceanographic survey of the bay was made by the *Godthaab* expedition which took a large number of observations there in the summer of 1928.[34] These data were published in the Hydrographic Bulletin (Annually, 1929), and I have used them to construct a map of the prevailing circulation as shown by Figures 91 and 92, pages 139–140. In addition to this dynamic topographic map of Baffin Bay, some of the best available information is contained in the drifts of ships caught by the ice in various parts of the bay. These follow:

Ship	Distance	Drift from		Drift to	
		Latitude	Longitude	Latitude	Longitude
	Miles	° ′	° ′	° ′	° ′
North Star (Saunders)	150	75 40	60 50	71 30	71 05
Advance (DeHaven)	900	75 30	93 00	67 00	60 30
Enterprise and Investigator	290	74 00	92 30	72 50	73 00
Resolute (Kellett)	1,020	74 41	99 30	64 30	62 30
Rescue (Griffin)	900	75 30	93 00	67 00	60 30
Fox (McClintock)	1,194	75 15	62 16	63 47	56 36
Polaris Party (Tyson)	1,700	78 45	72 00	55 00	52 00
Greely's Boat (Greely)	65	79 40	72 00	78 45	74 00

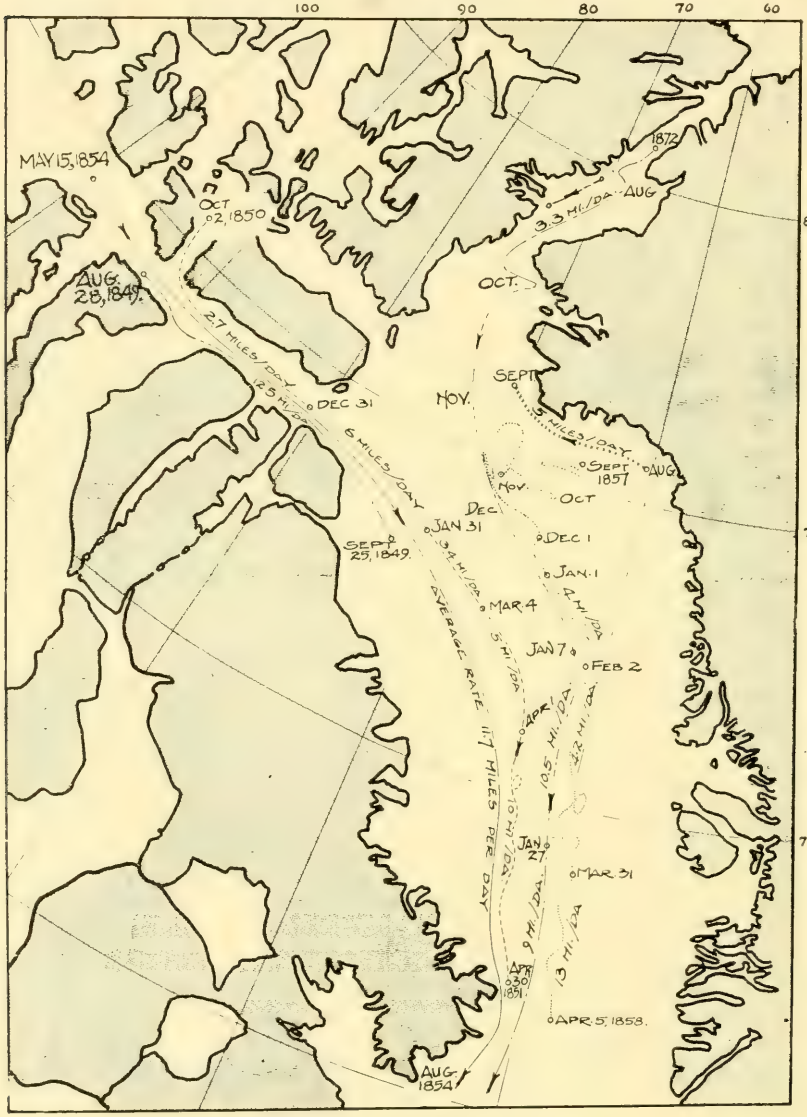
Icebergs have been observed to drift northward along the east side of Baffin Bay. On the other hand the *Marion* expedition mapped a southerly current along its southwest side with a rate of 7 miles per day. (See fig. 96, p. 148.) Thus it seems well established that a general cyclonic circulation prevails, so that the western (iciest) zone evacuates through Davis Strait, while a compensating indraft follows northward along the Greenland side.[35]

What proportions of the pack that moves out through Davis Strait into the North Atlantic originate in Baffin Bay itself is an interesting question. Consideration of the above table and other available data throws some light on this problem. The first of the pack that emerges through Davis Strait, on the resumption of

[34] Riis-Cartensen (1929) gives a preliminary account of the expedition, the published reports of which have not yet appeared.

[35] Mecking (1906, supplementary map) has drawn a southbound current and ice band parallel to the eastern shore of Baffin Bay, its axis closely coinciding with the 600-meter isobath. Such a representation is based largely on the drift of the *Fox* in 1858, and to a less extent on the behavior of the middle ice. This feature of the circulation is, however, not confirmed by the recent oceanographic observations of the *Godthaab* expedition nor is Mecking's southerly current supported by the general laws for dynamic gradient currents in the Northern Hemisphere.

freezing, may be credited to ice that has entered the bay from more remote sources during the preceding summer to mix there with the local production. Heavy masses of Baffin Bay origin follow, con-



RECORDED DRIFTS IN BAFFIN BAY

FIGURE 24.—The drifts of ships beset in ice and also floe parties, such as that of the *Polaris*, throw considerable light on the general movement of the water and ice. — - — - — -, the *Polaris* floe party. xxxxxxxx, the *North Star*. . . . . ., the *Fox*. - - - - - - - - - -, the *Advance* and the *Rescue*. ————, the *Resolute*. — - — - — - -, the *Enterprise* and the *Investigator*. (See Table, p. 47.)

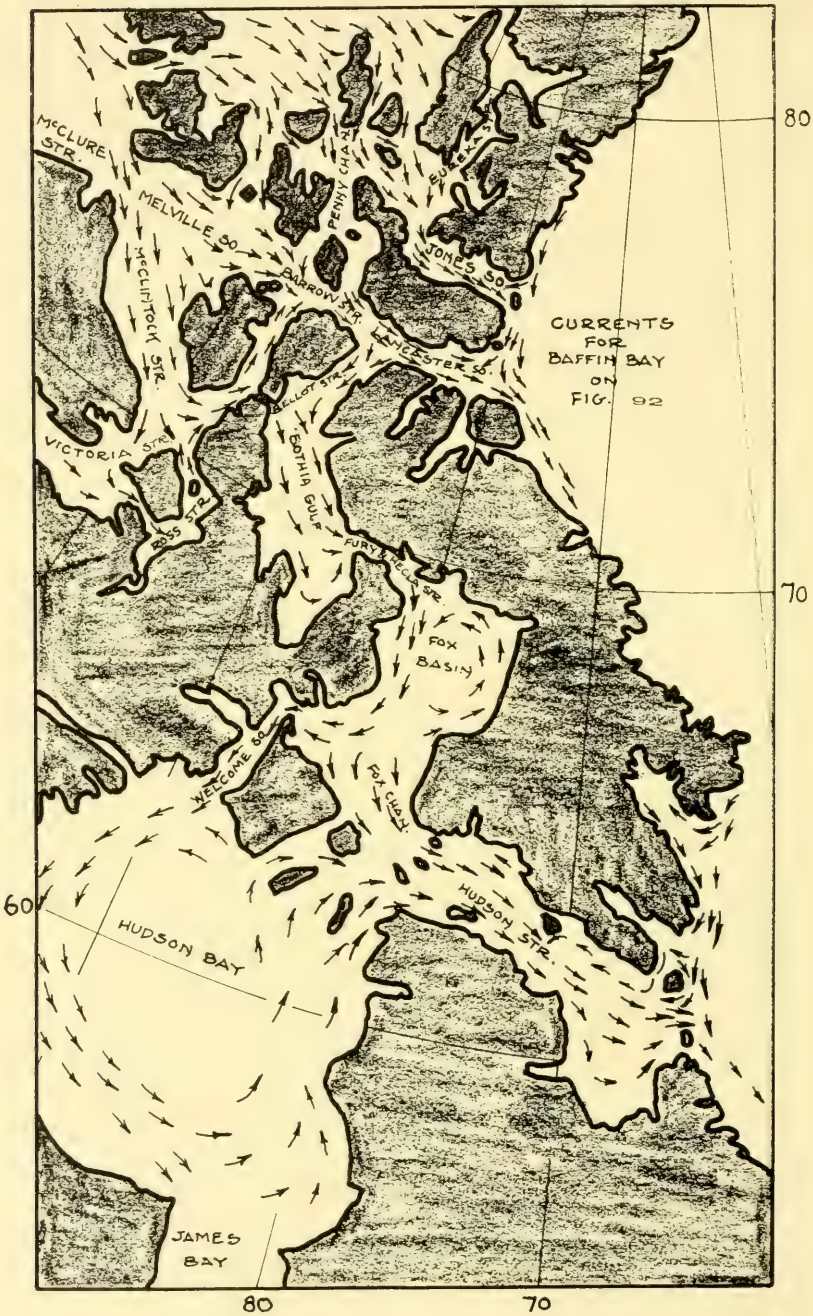
stituting the bulk of the pack that characterizes at that season. The ice of late summer and fall comes mostly from regions quite remote, Baffin Bay acting as a cachment basin with the only escape through the lower end.

The drainage area of sea ice to Baffin Bay extends hundreds of miles into the tributary sounds and straits of the American Archipelago. One of the largest streams of pack is discharged into Baffin Bay through Barrow Strait and Lancaster Sound. The drift of the two ice-beset ships *Advance* and *Rescue* of the U. S. Grinnell expedition in 1850 (see table on p. 47) shows the general direction of their courses. Kane (1854, p. 522) states that the rate of drift of these ships with the pack was approximately 2.5 miles per day during October in the west end of Lancaster Sound. It increased to 3 miles per day during December in the mouth of the sound and attained a maximum rate of 5 miles per day off the northeast coast of Baffin Land in January. The British ship *Resolute* beset in the same waterway was carried halfway across the archipelago to Davis Strait in one season. Mecking (1906, p. 27), after examining the records of many of the searchers for Franklin shows that much ice must be carried through the archipelago into Baffin Bay and Davis Strait, and Figure 24, page 50, shows the circulation of the water in the Arctic Archipelago as deduced by Mecking. We conclude in view of the foregoing data that not more than two-thirds of the pack that drifts out of Baffin Bay is actually formed within the latter.

HUDSON BAY PACK

Ice begins to form in Hudson Bay during October and by the end of the month most of the harbors are frozen. The bay itself remains comparatively free from ice during winter except for a 5 to 6 mile wide fringe. According to Lowe (1906, p. 293) fast ice continues to make even up to the 1st of June, but when it begins to break up it does so rapidly, sometimes early in July. During a boisterous winter the ice is liable to raft, in which condition its melting time is much lengthened. Aerial observations of ice conditions in Hudson Strait have been reported by McLean (1929, pp. 12–13). The appearance and disappearance of the ice is from west to east. It arrives at the western end of the strait in November and two weeks later is found at the ocean entrance. February records only 15 per cent of open water, the congestion remaining until the month of May when a noticeable decrease is observed. The middle of July normally records 90 per cent of open water and a navigable Hudson Strait. Hudson Strait is deemed safe for navigation during normal years from the latter part of July or first of August until the latter part of October. Navigation of this region is an important commercial problem for Canada, the principal difficulty lying in the blocked condition of the eastern end of Hudson Strait (see McLean, 1929). Congestion there during spring and early summer is caused by ice from Hudson Bay and Fox Channel mixing with that from Davis Strait. Not only does the Davis Strait pack cross the mouth of Hudson Strait, but it is also carried by the current in along the north side for a distance of 120 miles or more before it recurves to pass out parallel to the opposite shore. Under such conditions it is very difficult to distinguish between the arctic ice and the heavy floes of local derivation. The thickness of Hudson Strait and Fox Channel glaçons may vary from 7 to 19 feet.

Hudson Bay and Fox Channel, with their wide shallow areas, have often been described as ideal regions supplying the main stream of pack ice which moves southward into the western Atlantic.



THE CURRENTS IN THE ARCTIC ARCHIPELAGO

FIGURE 25.—The basis for the currents indicated by the arrows has been obtained from a number of observations on the drift of floating objects, such as ships, wreckage, buoys, ice, etc. The arrows show quite conclusively that the general movement of the water and the ice is from the northwest out into Baffin Bay and Davis Strait. (The map is taken from Mecking, 1906.)

Besides the local ice, additional masses of pack from more northern sources are discharged into Fox Basin through Fury and Hecla Strait. Mecking (1906, supplementary map) (fig. 25, p. 50) shows that the pack ice from the Arctic Ocean enters the northern end of the Gulf of Boothia and amasses in the lower end of the latter. The congestion is partly relieved by the escape through Fury and Hecla Strait, from whence the pack continues southward through Welcome Sound and Fox Channel. The escape of ice from all of these cachments is greatly hindered, nevertheless, by obstructing islands and narrow channels.

The famous ice-choked condition of Hudson Strait itself may be due not wholly to the ice from Hudson Bay and Fox Channel but partly to the great floes of Baffin Bay ice, which from our knowledge of the currents are certainly brought to this locality. Hudson Bay itself, due to the fact that its fast ice rapidly dissipates, being only 2 to 3 feet thick, does not contribute such large quantities of pack to the Atlantic as it first might appear. The major part of the pack which hampers Hudson Strait during the spring, moreover, consists largely of the Baffin Bay variety, augmented by smaller contributions of fast ice which have formed in Fox Channel (see McLean, 1929, p. 13) and along the sides of Hudson Strait itself.

It is interesting to speculate in what proportions the pack drifting out through the lower end of Baffin Bay and the ice setting out of Hudson Strait contribute to the total mass that drifts southward to the Grand Bank. Mecking (1906, p. 106) shows that the most of the pack to Newfoundland comes from Baffin Bay, it having formed there or brought to the bay through the waterways of the Arctic Archipelago. Munn, on the other hand, stresses the sea-ice discharge through Hudson Strait.[36] If the pack to Labrador be divided according to its sources—(*a*) Baffin Bay ice, (*b*) Arctic ice via Baffin Bay, and (*c*) pack ice through Hudson Strait, we believe the following respective weights are representative: 60, 30, and 10.

PACK ICE ALONG COASTS OF LABRADOR AND NEWFOUNDLAND

Ice appears at the mouth of Fox Channel and in Hudson Strait in October and November, the time varying somewhat from year to year, depending on meteorological and oceanographical conditions. The pack ice out of Hudson Strait, and the first of the glaçons and floes which have begun to swell southward from Davis Strait join off Cape Chidley, the northern extremity of Labrador. The combined packs, first in narrow strings and strips, and then in a much broader, heavier stream, reach the northern Labrador shelf early in November. December witnesses the advance along the coast and its arrival off Newfoundland in January. Fast ice during this period makes inside the headlands and harbors, the freezing time for the northern section being November, and for the southern estuaries December. Newfoundland harbors freeze in January but the ice is seldom very heavy and readily breaks up in April.[37] Sea ice, it is said, will make in open water on cold calm nights in the

[36] The drift of a tool box incased in ice from the inner waters of Hudson Strait out and down the Labrador coast to Nain proves ice for Newfoundland comes from this outlet.

[37] The harbor of St. Johns is often blockaded by the northern pack during the months of February and March at the time when the sealing steamers wish to depart. Exit is sometimes only accomplished by means of much cutting through the sheets, and often blasting when the pack is especially heavy.

latitude of St. Johns, Newfoundland, 1 to 2 inches in thickness, and out from the coast for a distance of several miles. Rodman (1890, p. 26) has published a table showing the approximate dates of appearance and disappearance of ice along the Labrador and Newfoundland coasts.

On several previous occasions in discussing certain regions we have called attention to the important influence which bathymetrical conditions have on ice distribution. The Labrador shelf is no exception to the rule, providing a high road, so to speak, along which the pack may easily advance to lower latitudes. The bathymetrical map of Davis Strait shows that the Labrador shelf is much wider than that along the other coasts of this region. It maintains an average width of 80 miles, as determined by the 500-fathom isobath, from Cape Chidley, Labrador, southward to Hamilton Inlet, thence to the latitude of Cape Race it spreads out very wide; for example, off St. Johns it measures nearly 280 miles. The breadth and general out-

THE OFFING OF THE LABRADOR COAST IN JUNE

FIGURE 26.—The procession of pack ice which is continually being borne southward along the Labrador coast for seven months of the year by the cold current. This coastal belt of pack ice is claimed to play an important rôle in the southward distribution of the icebergs, they being fended off the coast and kept out in the cold current. (Photograph by E. M. Kindle.)

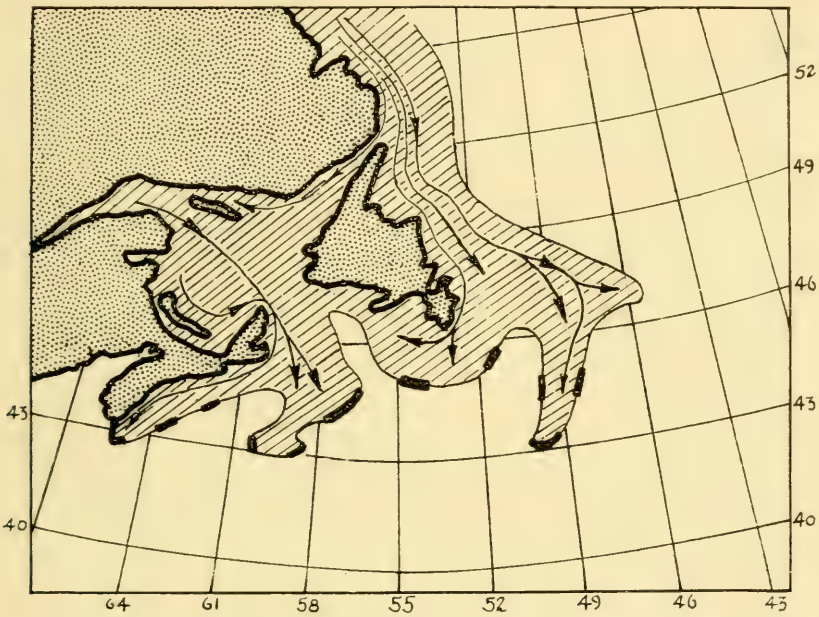
line of the east North American pack along this coastal stretch is largely a reflection of the depths. In years of abundant pack, the outer edge of the field off St. Johns has been recorded a hundred or two hundred miles from the coast.

As summer advances, the pack melts back toward its northern roots uncovering first the Newfoundland and then the Labrador coast lines. The Strait of Belle Isle is usually open to navigation from July to December, the first of the trans-Atlantic steamers entering June 15 to July 1 and the last passing out the first week in December. The Labrador coast is often free of pack ice, at least for navigation, during July, while in other summers the coast has been continually hampered.[38]

[38] An excellent example of the rate of dissipation of the Labrador pack is afforded by the fact that the *Godthaab* expedition in early June, 1928, found a field of pack ice extending along a large part of the Labrador shelf of 18,000 square miles area, but six weeks later the *Marion* expedition found these waters clear and all ice disappeared. Pack ice in the western North Atlantic was markedly below normal the year of 1928.

PACK ICE ON THE GRAND BANK

The pack reaches the northern part of the Grand Bank late in January or early in February, where the water, still cold as a result of the preceding winter's chilling, keeps it from melting rapidly. The Newfoundland Banks (Grand, Green, and St. Pierre), aggregating some 60,000 square miles, are the submerged continuation of the North American Continent, which slopes here southwestward far out on the Atlantic sea floor. Off southern Newfoundland the pack tends to part as the current meets the northern buttress of the Grand Bank, around which it sweeps. The inshore arm no longer flanked by the coast line moves southwestward past Cape Race in the submarine ravine known as the Gulley, but the larger, heavier tongue



THE DISTRIBUTION OF PACK ICE SOUTH OF NEWFOUNDLAND

FIGURE 27.—The limits to which the main body of pack ice has been recorded south of Newfoundland is shown above. The short heavy lines represent the positions at which fields of pack ice have been sighted during the height of the ice season. (Figure after Huntsman, 1930.)

follows southward along the eastern side of the Grand Bank. The western fields in years of great abundance may block the bays and harbors along the south shore of Newfoundland as far west as the Miquelon Islands[39] and send scattered floes even out to the edge of the Atlantic slope. The field which drifts down along the eastern side of the bank is interesting, not only on account of its intimate association with the icebergs but also because it attains the southernmost point to which sea ice from high latitudes drifts in the Northern

[39] The waters of southern Newfoundland have often been described as dominated by a current which sets westerly from Cape Race along the coast and which rounds Cape Ray entering the gulf. Much data, however, such as current meter measurements, the distribution of salinity and temperature, the results of hundreds of drift bottles, the general configuration of the bottom, and the tidal movements, show that the current and ice after rounding Cape Race do not set any farther west than the Miquelon Islands.

Hemisphere. In heavy ice years the pack lengthens out in long strips (see fig. 5, p. 16), and chains paralleling the edge of the bank, i. e., the direction and flow of the current, as far south as the "Tail" on the forty-third parallel. The farther southward the pack drifts the more open it becomes and likewise the shorter time it lasts. South of the forty-fourth parallel it is very patchy and dismembered, surviving only a day or two. The latter part of March and the first of April witness the deepest southern invasions.[40]

Scattered floes from the ice tongues are continually being broken away by the prevailing westerly gales and driven across the continental edge into deep water, where the ocean swell and warm surroundings rapidly melt them away. A ship may report sighting patches of pack in the morning, while another vessel passing the locality in the afternoon may see no signs of it at all. It is a well-established fact that pack ice rarely, if ever, extends westward around the Tail of the Grand Bank. Although the cold current might tend to carry the ice in such a direction the strong westerly winds prevailing at this time of the year are dominant. A floe of pack ice was reported to the ice patrol in March, 1924, as 120 miles southeast of the Tail of the Grand Bank.

The fields on the Grand Bank reach a maximum during April, after which they recede, and by the latter part of the month or the first of May extend no farther south than the northeastern part of the Grand Bank. Under favorable conditions small fields of pack may be sighted occasionally along the northern slopes of the Grand Bank throughout May, but finally summer temperatures cause its complete disappearance. The distribution of pack ice tends to follow the primary circulation of the water, which over the Newfoundland Banks progresses as a number of vortices, one spilling over into the other. The winds always exert a great effect, tending to mask that of the currents. A characteristic embayment in the pack ice on the Grand Bank is nearly always to be observed over the southwestern slope, as shown on Figure 27, page 53, where unmistakably warmer water floods in from off-shore.

Although in the open Atlantic the pack tends to scatter and many floes to drift away from the main fields, the ice, nevertheless, retains considerable strength and is still an imposing spectacle. On the Grand Bank, even as far south as the forty-fifth parallel, ships may easily become imprisoned, with no water from the masthead as far as the eye can see.

If we examine a few of the glaçons in more detail, we find that all exhibit a more or less tabular shape, with some of the older pieces hummocked in round uneven contours. The glaçons on the outer edge of the pack, and scattered here and there in the open water, show evidences of the greatest deteriorations. Melting, however, of all the ice progresses much faster at the water line than above or below, resulting in characteristic tabular and hourglass shapes. The thinner, smaller portion is always uppermost, not only on account of equilibrium but also because the portion exposed to the air and to the sun in the cold water of spring melts faster than the part below water. The outward sloping form of the submerged under-

[40] According to the Deutsche Seewarte the most southerly penetration of pack ice was in April, 1887, May, 1885, and June, 1882 and 1883, when it was sighted on the fortieth parallel near longitude 49°.

body easily explains its most serious threat to the propellers of ships. In advanced stages of melting the glaçons show many gracefully rounded erosions, and continual licking of warm waves sculptures the ice into fantastic shapes.

GULF OF ST. LAWRENCE PACK

Pack ice from along the coast of Labrador enters through the Strait of Belle Isle into the Gulf of St. Lawrence. How much enters is problematical but undoubtedly the indraft is greatest with easterly winds when a continuous flood current has been observed for the period of a month. Westerly winds on the other hand tend to drive the ice off into the Atlantic and to set up an outflow of the water.

PACK ICE SPREADS OVER GRAND BANK SOUTH OF NEWFOUNDLAND

FIGURE 28.—Pack ice is carried over a thousand miles southward out of Davis Strait and into the North Atlantic every spring. The duration of pack ice so far south is a matter of a few weeks only. This photograph was taken April 9, 1921, in latitude 45° 50′ N., longitude 49° 20′ W. A seal can be seen on one of the floes. (Official photograph, international ice patrol.)

The normal circulation in the Strait of Belle Isle is an indraft along the Quebec shore, and an opposite set along the Newfoundland side. From December to July, however, pack ice is liable to be carried into the gulf, the largest contributions tending to hug the Labrador coast on the northern side of the strait. Huntsman (1930, p. 6) relates an occurrence in June, 1897. So much of the pack was brought through the strait by favorable winds that fishing was interfered with for a distance of 150 miles around, although the whole gulf had previously been open. If this can happen in summer, a natural query is, how much greater quantities must enter during winter, when conditions are probably more favorable?

River ice, gulf ice, and Davis Strait ice are mixed to form the gulf covering. But in what proportions these mix there, is not

known. According to Huntsman (1930, p. 7) the frigid character of the gulf is almost wholly due to the Davis Strait ice, either directly as it drifts in or indirectly as it cools the inflowing water. The fact that the contributions from Davis Strait through Belle Isle bestow an icy character on the Gulf of St. Lawrence is clearly demonstrated by comparing the conditions there with those of Hudson Bay. Although the latter is much farther north it remains comparatively open except around it shores while the Gulf of St. Lawrence is ice congested. The Strait of Belle Isle is ordinarily open to navigation from July to December. The summer route through the strait is much traveled because it provides a short ocean journey to Europe. The mileage from Montreal to Liverpool, including two days on inland waters, is 2,785, against 3,100 from New York.

Some of the patches of pack ice reported earliest in the season in the western North Atlantic, sighted on the northern part of the Nova Scotian shelf, have drifted out of the St. Lawrence through Cabot Strait. The main body of this pack moves out past Cape North and Scatari Island, on the Cape Breton side, and spreads southerly toward Sable Island. (See Huntsman, 1930, p. 7.) Another branch, consisting mostly of sludge moves southwesterly along the Nova Scotia coast, even as far as Halifax. But since its presence is mostly due to favorable winds, its existence is brief. Glaçons and sludge in very small quantities have been known to drift at rare intervals southward past Cape Sable, but such ice is rapidly melted and, according to Bigelow (1927, p. 698), never drifts into the Gulf of Maine. Strings of the St. Lawrence pack are often blown considerable distances offshore, sometimes reaching the vicinity of Sable Island (as shown on fig. 27, p. 53), or even surrounding the island, but very seldom is any of this ice considered a menace to navigation. The ice patrol usually advises trans-Atlantic ships for Halifax to select a course south of Sable Island, whereby they will avoid all dangerous ice.[41] Not only may the St. Lawrence pack be the first to drift out into the Atlantic during spring, but in the face of approaching summer it is usually the latest to disappear from the latitudes south of Newfoundland. Its persistence is partly due to the temperature of the waters of the Gulf of St. Lawrence and of the inflow from the latter. The St. Lawrence pack may spread out from Cabot Strait occasionally, over an arc from St. Pierre to the Cape Breton coast in February to April, then gradually shrink to the mouth of Cabot Strait during May. The position and extent of the St. Lawrence pack attracts attention from the latter part of April until the middle of May, due to the large number of steamships which are attempting to force passage through it. The field may be described as follows: (*a*) An outer zone consisting of loose sludge and glaçons broken by numerous leads of open water and bounded offshore by an arc from Miquelon Island to Cape Canso; (*b*) an inshore zone of heavy pack ice without leads, and with its outer edge following a convex curve from Cape Ray to Scartari Island; (*c*) the innermost zone, consisting of heavy rafted ice packed tightly in an effectual barrier, Cape

[41] Small quantities of pack ice of local origin form in certain sheltered areas along the eastern coast of the United States as far south as New York during unusually severe winters; e. g., in Cape Cod Bay, in Vineyard Sound, and in Long Island Sound. But the extent of these packs is very limited and their existence brief.

Ray to Cape North. The ice is usually tightest on the Cape Breton side of Cabot Strait, and loosest on the Newfoundland side. In light ice years open water often extends northward from Cape Ray to the Bay of Islands.

The St. Lawrence pack recedes in much the same manner that it advanced, the lower gulf being blockaded longest by the ice floes. After the river ice is broken up by the Canadian Government ice breakers, an ice patrol of the gulf is established in order to furnish the best information possible to entering ships. The ice conditions of the Gulf of St. Lawrence are of great economic importance to Canada, practically all of her overseas trade passing through this waterway on its course to and from the river ports of Montreal and Quebec. The gulf is usually safe for navigation by the first half of May, and in some years even as early as the latter part of April. Navigation to Montreal and Quebec normally closes in the month of December.

Annual Variations in the Limits of Pack Ice in the Western North Atlantic

The pack ice which drifts southward along Labrador blockading the coast and often encircling Newfoundland is subject to great annual variations. In some years the amount may be so small as to be almost negligible while during others the fields are extensive. Records for the past 350 years report important variations in the annual limits of pack ice in the western North Atlantic. Martin Frobisher, searching for a northwest passage to India in 1576, encountered great floes and huge icebergs off southern Baffin Land, but in 1588 John Davis landed without difficulty at several points along this normally icebound coast. Kane in 1853 found ice conditions very favorable and thus was able to sail farther north between Greenland and Ellesmere Land than any previous explorer, but McClintock, four summers later, in the *Fox* following the usual track across Melville Bay was nipped and held tightly in the pack until released the next spring. The annual abundance of pack has also been closely followed by the Newfoundland sealers who realize that a scarcity of ice usually means a poor catch and small profits. Trans-Atlantic ships have for years been reporting ice sighted to the hydrographic offices of their respective countries in the hopes of greater common safety. The seventh international geographical congress held at Berlin designated the Danish Meteorological Institute at Copenhagen as the official repository of reports on ice conditions in the Arctic regions. That institution publishes annually a digested report on ice conditions, together with maps showing the pack-ice limits by months. The Canadian signal service, from its posts in Labrador and Newfoundland is keeping shipping advised daily of ice conditions past these stations. The international ice patrol has also recorded the boundaries of the pack south of Newfoundland since 1913.

The variations in the limits of the pack ice off Newfoundland have been studied by Meinardus (1906); Mecking (1906 and 1907); Schott (1904 and 1904a); and Smith (1925 and 1926a). Meinardus (1905), investigated what relationship prevailed between the annual amounts of pack ice in the western North Atlantic and certain contemporary

changes in the intensity of the North Atlantic atmospheric circulation, coming to the conclusion that a weak circulation of air from August to February is followed by a relatively small amount of pack ice off Newfoundland during the succeeding spring. After a strong circulation during the fall and early winter months more pack than usual spreads over the Grand Bank. His studies are especially important because they are based on a series of years, running back to 1860. The scale is +2 for a year of much ice, and −2 for little ice, a closer comparison probably being impossible due to the nature of the material for the early years.

Mecking (1906 and 1907) has published two important papers as a result of a careful study of pack-ice conditions in the northwestern North Atlantic, and an investigation into the possible causes of the annual variations for the period of 1880 to 1900. His data are obtained from the following sources: United States Army Signal Service, United States Hydrographic Office, United States Weather Bureau, and Deutsche Seewarte.

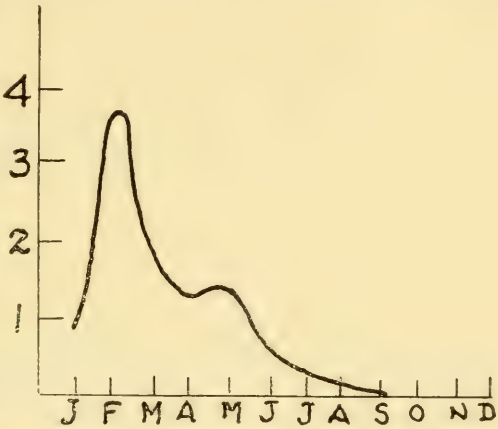
The records of these offices consist of the reports of ice sighted by trans-Atlantic ships on their regular voyages through the ice regions off Newfoundland. The adoption of prescribed lane routes across the Atlantic in 1875, their modification in 1898, and the present method of seasonally shifting the tracks whenever ice conditions are a serious menace, are all modifying factors which must be given due consideration in arriving at an accurate ice record over a period of many years. The fact that many reports often refer to the same field or floe may result in duplication and so caution is needed for a correct compilation. Mecking has conclusively shown that the factor chiefly controlling the variations in the limits of the spring pack ice in the northwestern North Atlantic is the barometric gradient during the previous winter across the ice stream in the vicinity of the Labrador coastal shelf. The assumption is that favorable winds and currents during the colder months of the year over Labrador will drive more pack ice than normal past Newfoundland in the following spring. The agreement between the values of the ice curves and the pressure gradients on his graphs is close. The spring of 1887 was, however, an exception when a great quantity of pack appeared off Newfoundland, although the pressure gradient had averaged weak during the preceding winter. This inconsistency, Mecking thinks, was due to the scarcity of icebergs which normally tend to break up the pack ice, allowing it to drift freely. Also the year of 1889 was peculiar in that practically no pack ice drifted south of Newfoundland during the spring despite a favorable pressure gradient. Mecking attributes the inconsistency to the extremely warm summer of 1888 which melted so much ice as to produce a deficiency in the following spring. And in this respect this is the only case recorded when the temperature in one summer was noticed in the crop of pack ice the succeeding spring.

Schott (1904, p. 305), with the aid of ship reports contained in the files of the Deutsche Seewarte, reviewed the period 1880–1891, comparing each year of the series with regard to quantity of pack ice off Newfoundland. He agrees with Mecking that 1889 was an unusual year, but he points out that the pack appeared in September, to remain for the balance of the year.

Mecking (1907, p. 11) found that the pack ice off Newfoundland in normal years reaches its maximum in February and then diminishes to a secondary, much lower maximum in May. The monthly percentages during a normal year are:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
9	37	18	13	14	5	2	1	0	0	0	0

The chief maximum results from the arrival off Newfoundland of the accumulation of ice from Davis Strait in February. And by calculating the rate of drift and the distance, he concludes that the winds most effective in its transport are those of December. A year characterized by winds stronger than normal will not only advance the date of this maximum but will also bring a greater abundance of ice; weaker winds not only postpone the maximum but drive down less ice. The date of the maximum may vary from February to May, or to even June in some years. The second maximum he believes represents the ice which became entangled among the relatively slow moving body of bergs, which may delay the pack as much as two or three months.



SEASONAL VARIATION IN PACK ICE IN THE WESTERN NORTH ATLANTIC

FIGURE 29.—A graph representing the relative volume of pack ice by months normally drifting south of Newfoundland. This graph is based upon ice sighted by trans-Atlantic shipping. (Compiled by Mecking, 1907.)

In the course of an investigation on the annual variation of icebergs, conducted by the international ice patrol, data were compiled on the monthly record of pack ice. No unmistakable evidence of a double maximum of pack ice appears in these records, and in view of the dispersal that takes place, especially [42] in the case of the icebergs, in the course of a 1,500-mile journey, it seems unlikely that such a secondary maximum should develop.

My own study of the meteorological and ice conditions for the period 1880 to 1926 (Smith 1925, p. 229 and 1927, p. 31), was based on the data on file at the British Meteorological Office and on the ice records collected by Mecking (1907), by Schott (1904), by the United States Hydrographic Office, and subsequent to 1913 by the international ice patrol. The period investigated embraces 47 years, a series of sufficient length to permit mathematical correlations and

[42] The steamer traffic to the St. Lawrence beginning in May each year follows the tracks past Cape Race which are practically deserted all winter. Concurrently, the ice records of the Hydrographic Office for this region show a marked increase. It is sometimes very difficult to determine whether or not the reports have been duplicated, and if so, how many times; thus it is very easy to record too much ice. Is it merely a coincidence that the date of opening this region, May, is also the time of the second ice maximum?

thus to avoid any liability to bias. The correlation between the winter atmospheric gradient, Ivigtut to Belle Isle, and the spring crop of pack ice south of Newfoundland was found to be +0.86. (See pp. 180–189.)

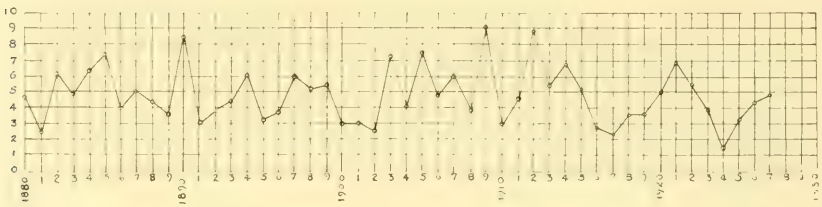
Brooks and Quennell (1928, p. 33) besides investigating the effect of Arctic and Greenland Sea ice on European weather also studied the effect of varying masses of pack ice and cold water introduced off Newfoundland. They found:

1. Much pack ice off Newfoundland in the spring, April to June, tends to occur with low atmospheric pressure for the same period at Iceland and high pressure at the Azores.

2. Much pack ice off Newfoundland in the spring tends to be followed nine months later by high pressure over northern Norway and low pressure over southern England.

3. Much pack ice off Newfoundland in the spring tends to be followed 15 months later by high pressure at the same places as (2).

Finally, the effect of ice off Newfoundland on the pressure over western Europe is generally similar to that of the polar cap ice and of the Greenland Sea pack, but as might be supposed, is much less pronounced. The correlation coefficients between ice off Newfoundland, April to June, and pressures at Vardo, Valencia, and Berlin, June to March, of 30, 20, and 31, respectively, are not high, yet may have some small value for forecasting European weather.



PACK-ICE GRAPH FOR THE WESTERN NORTH ATLANTIC

FIGURE 30.—A graph representing the relative amounts of pack ice south of Newfoundland by years, 1880–1927. The data upon which this graph is based was taken from Mecking (1906) for the years 1880–1900, and since then from the records and the researches of the international ice patrol.

LAND ICE

Glacier ice, formed from precipitation on land, is of great importance as the source of icebergs. Under the present distribution of temperature and snowfall over the earth, the permanently ice-decked lands lie mostly within the polar regions. The greatest single ice sheet in the Northern Hemisphere is that which overlies Greenland, in area equal to all lands east of the Mississippi River and south of the St. Lawrence River. Greenland is the principal source of the icebergs that are found drifting in the North Atlantic and in its tributary seas.

The treatment of icebergs must necessarily include their general distribution in time and place, their form, size, color, markings, volumes of flotation, manner of disintegration, etc., all depending to a great degree upon conditions which existed long before the iceberg was born. Chamberlin, Drygalski, Koch, Priestley, de Quervain, Hobbs, and many others have carried out notable observations

and studies on polar glaciology. It is mainly from this literature that the following classification is taken. Practically all the various forms in which glaciers have been observed may be classified as follows:

(*a*) *Cachment, hanging, cirque*, or *cwm*, a glacier occupying a small depression on a slope.

(*b*) *Alpine.*—The common valley glacier.

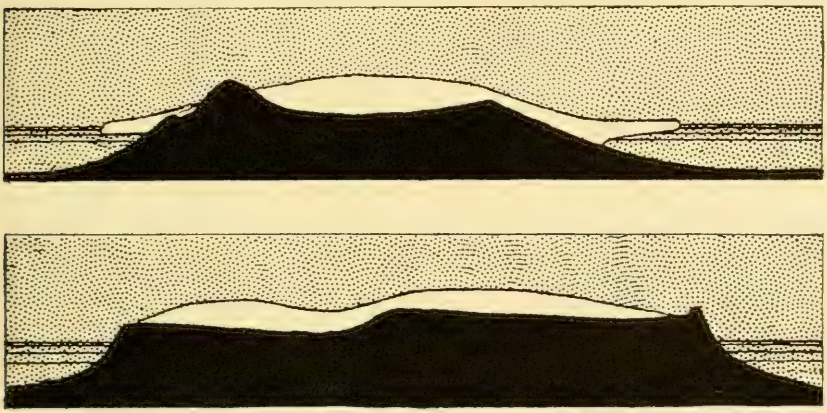
(*c*) *Plateau or highland.*—Spreading from one or more cachment basins over a level plateau.

(*d*) *Piedmont.*—A coalescence of ice masses in cachment basins, or of glaciers at the foot of a descent.

(*e*) *Inland ice.*—All land forms hidden.

(*f*) *Ice-foot or snow-drift glacier.*—Wrapped about foot of mountains.

(*g*) *Shelf ice.*—Accumulation of snow on fast ice of protected coastal shelf.



A COMPARISON BETWEEN THE ANTARCTIC AND THE ARCTIC ICE SHEETS

FIGURE 31.—The profile of Antarctica, above, when compared with that of Greenland, below, reveals a striking dissimilarity in the general form of the ice sheets. The former with its marginal overflow, causes the ice to calve many huge tabular icebergs. In the north, however, the ice edge characteristically ends on land, and glaciers plowing across the uneven foreland produce irregular, picturesque-shaped icebergs. (Figure after Priestley and Koch.)

The first five forms have a wide distribution in polar regions, but the last belongs to the antarctic. The ice-foot or snow-drift glacier should not be confused with the ordinary ice foot which is so common in the north, the latter consisting of salt-water ice formed on a chilled shore line near the tide line. (See p. 23.) The ice-foot glacier is seldom found in the north. Differences in certain underlying factors specific to the region develop corresponding differences in the features of the ice. As an example of one of these agencies we can point to the low mean annual air temperature which prevails in the Antarctic. The warmth of the arctic summer has no parallel in the far south and mainly because of this thermal difference the ice sheets of the north polar regions are unlike those of the southern. The margin of Antarctica's cap, overflowing its land support, is free to spread over the sea until fracture detaches huge strips, sometimes including 10 to 20 miles of its front. In Greenland, by contrast, the edge of the inland ice ends on land, and glacier tongues are deformed as they plough across the uneven foreland so that their

icebergs are irregular in shape. The box-shaped berg is, therefore, in general, characteristic of the Antarctic, as the pinnacled, picturesque type is of the north.

We have called attention to the fact that Greenland is the only land of continental size in the Northern Hemisphere which supports an ice sheet.[43] At first thought it may seem surprising that other extensive land areas, some of which lie much nearer the pole than does this seat of glaciation, remain nevertheless quite bare. Thus the northern sections both of Greenland and of Ellesmere Land are destitute of an icy covering. The white colored areas on Figure 11, page 20, indicate regions of glaciation. Similarly the American Arctic Archipelago is for the most part ice free, notwithstanding its frigid climate. Labrador, with its low mean summer temperature of 6.9° C. (44.5° F.), and its position in the summer path of low pressures, might also be expected to exhibit an ice covering, but actually shows only a few small cirque glaciers in the Torngat Mountains. High latitude, obviously, is not the only glacial requirement; there are several other fundamental climatic factors involved such as precipitation, elevation, distribution of land and water, prevailing winds, and ocean currents. Also, from a topographical standpoint, a region must not be exposed to prevailing winds of a velocity that the snow is blown away before it has had time to accumulate and build an ice sheet. Another glacial problem awaiting solution, in the case of the ice caps of Antarctica and of Greenland is, how can they be continually renewed when the ice itself tends to create a cushion of high atmospheric pressure, thereby tending to decrease the precipitation and so to lessen its own source of replenishment? Several theories have been advanced by Simpson, Hobbs, Meinardus, and others, but as yet no observations or conclusive evidence has been collected.

The snow and névé material, as they gradually accumulate, form a nucleus, and increasing in mass and thickness until finally the topographical features of the hinterland may be entirely obliterated, while the force of gravity causes the edges of the ice sheet to creep forward and outward along the paths of least resistance. This, in brief, is the history of the present 700,000 square mile Greenland Dome, and also of other similar areas of glaciation on the earth.

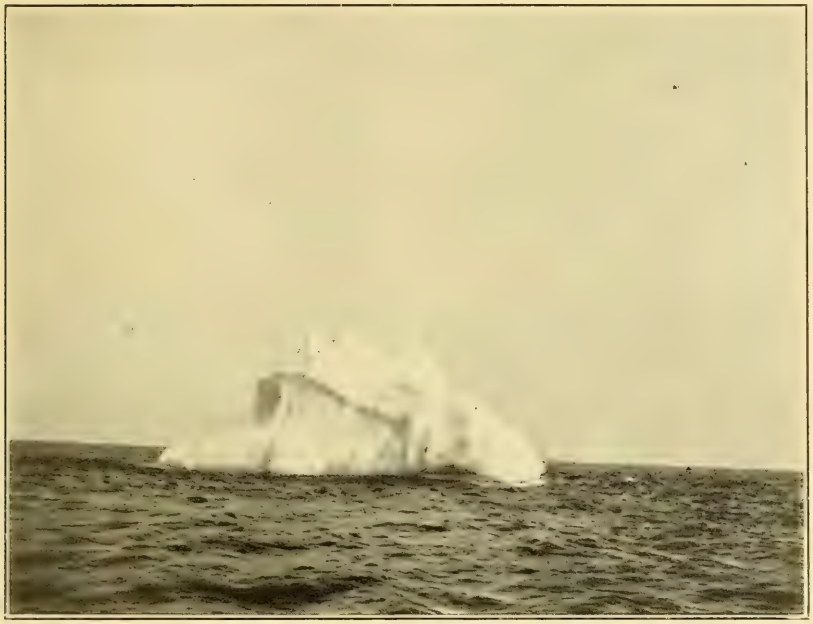
Greenland contains 90 per cent of the land ice of the north polar regions with the remaining 10 per cent lying largely around the shores of Baffin Bay. Smaller isolated areas are found on Prince Patrick Island and Melville Island, in the direction of the Beaufort Sea. Ice covers the Eurasian polar sector in Spitsbergen, Franz Josef Land, Novaya Zemlya, Nicholas II Land, the New Siberian Islands, and the DeLong group. The ice sheets of Iceland and Norway, the only other glaciated lands within the Arctic circle, are confined to the mountain plateaus, never reaching the sea.

Glaciation in Arctic Eurasia

In discussing the distribution of land ice, Spitsbergen deserves special mention as possessing two types of cover. Its northwestern part has numerous alpine glaciers separated by ridges and peaks;

[43] Stefansson (1922, p. 13) in remarking on the proportion of northern lands that are glaciated, adds that most people visualize the far north as completely covered with glaciers. He points out, however, that Greenland is the only land that really is extremely glaciated, and the total annual snowfall of Ellesmere Land, the second largest island in the polar regions, is barely one-tenth of that of St. Louis, Mo.

while North East Land is almost completely ice covered (see Hoel, 1929). The ice tongues in Spitsbergen move very slowly, finally debouching into the bays and fjords with glacier front walls not higher than 50 feet, and this slow rate combined with the low front yields only rather small icebergs. There are practically no glaciers in northwestern Spitsbergen which produce icebergs. The most productive berg glacier known in Spitsbergen discharges at the head of Storfjord and is called Negri, but nothing is known regarding the number of bergs that are calved. There are several glaciers on the eastern coast of North East Land which are productive; and King James Glacier on the eastern extremity of Edge Island contributes a number of bergs



A SPITSBERGEN ICEBERG

FIGURE 32.—An iceberg calved from a Spitsbergen glacier, sighted May 31, 1929, on the northward side of Bear Island. This is said to be one of the largest bergs sighted in the Spitsbergen or North East Land regions in several years. Its dimensions of 52 feet high by 150 feet long place it as only about one-fifth the size of the largest bergs found in the northwestern Atlantic waters. (Photograph by Capt. Thor Iversen.)

annually. Very little is known regarding the number of bergs that are released each year from Spitsbergen waters because they are far removed from the paths of navigation and therefore do not elicit especial interest. Capt. Thor Iversen, of the Norges Fiskeries, who has spent several summers in Spitsbergen waters, writes me that the small bergs typical of the region drift southwestward, and normally can be found in fair numbers from May until October around Bear Island.[44] The hydrographical observations of the Norwegian

[44] Captain Iversen says in a letter that the summer of 1929 was quite unique in that bergs and pack ice were much more plentiful than common. It is seldom for bergs to be sighted from the northwest coast of the Scandinavian Peninsula, but several strayed there in 1929. The drift of such ice south of Bear Island and along the Murmansk coast must have been due to the unusual winds and currents that were experienced that year. It is also Captain Iversen's observation and opinion that many bergs are detained temporarily in their drift by the ridge which runs between Bear Island and Edge Island and Hope Island.

Fisheries Department are now being prepared for publication, and when completed they will materially increase our knowledge of the movement of the pack ice and icebergs in Spitsbergen and adjacent waters.

Franz Josef Land (recently renamed Fridtjof Nansen Land), lying at the extreme end of the North Atlantic atmospheric depression, is wholly glaciated and several of its tidewater glaciers produce icebergs stated to reach a height of 60 feet, but the number of bergs is small. The northern half of Novaya Zemlya is also ice decked, with high and wide ice fronts reaching the sea at many points along

A TIDEWATER GLACIER OF SIBERIA

FIGURE 33.—The only iceberg-producing glacier in the eastern Eurasian sector. This glacier is located on the north side of Bennett Island, north of the New Siberian Islands, and discharges a few small icebergs. (Photograph taken from the Russian hydrographical expedition, 1910–1915.)

the coast, though its bergs are usually small. Nothing is known regarding the iceberg productivity of the west coast of Nicholas II Land. Its east coast, however, is known to have glaciers which produce icebergs, two or three score having been observed along the shore, even southward to Chelyuskin Strait in 1913 by the Russian hydrographical expedition.

Nicholas II Land, therefore, according to our present limited knowledge, ranks higher than Spitsbergen and Franz Josef Land as regards iceberg productivity. Bennett Island, made famous by the loss of Baron Toll, is glaciated in its southern half where along the coast few glaciers reach the sea. There is one ice tongue somewhat larger than the others, which probably accounts for the few

small insignificant bergs observed near Bennett Island by the Russians in the summer of 1912. The New Siberian Islands, while not ice covered, are the site of several cliffs of fossil ice which is, of course, without movement. The DeLong Islands and Wrangel Island, whose glaciers produce no bergs, completes the list of land-ice areas in Eurasian polar regions.

With the exception of the one small glacier on Bennett Island, not a single iceberg is produced along the Eurasian coast east of Cape Chelyuskin.[45] From our slight knowledge of the currents, the Franz Josef Land bergs should be carried westerly toward Spitsbergen, and the Spitsbergen bergs should, in turn, contribute to the total quantity of ice feeding into the east Greenland current. On the other hand the comparatively weak character of the circulation, especially in the eastern part of this area, argues that neither Franz Josef Land or Spitsbergen bergs are carried far. Many of these bergs, therefore, remain in the fjords and strand along the coasts. Compared to the great volume of glacial ice discharged from east Greenland, the possible contributions from Spitsbergen are minute.

The total area of tidewater glaciation of the northeastern sector of the North Atlantic is approximately 4 per cent of that of Greenland. On the basis of a productivity of 15,000 bergs for Greenland the northeastern sector contributes a total of 600 bergs annually.

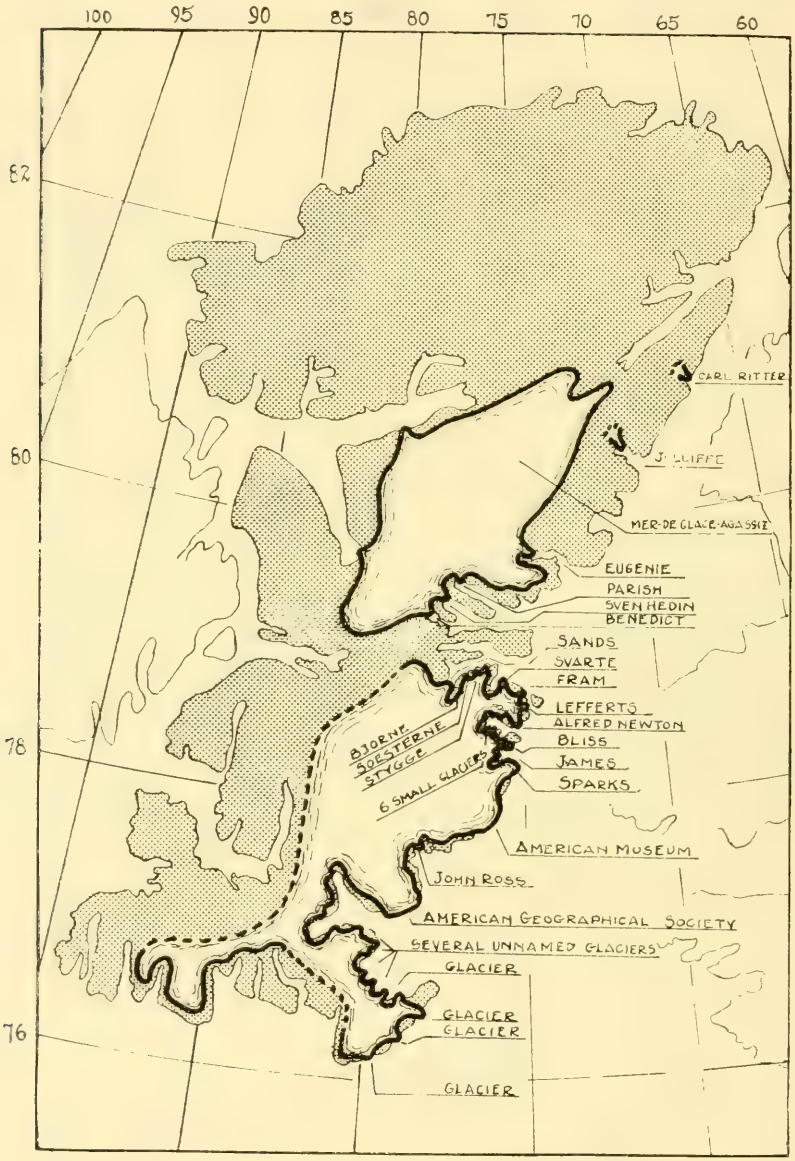
Glaciation in Ellesmere Land

One of the previously mentioned factors which determines the extent of glaciation; the altitude; is clearly revealed on the American side of Baffin Bay and Davis Strait. Ellesmere Land, deeply intrenched in the polar regions, presents only its east and south coasts for cursory view. From the decks of their ships, often kept away by menacing floes and fields, Ross, Kane, Hall, and others have recorded the general coastal picture. Due to its natural inaccessibility, the interior of Ellesmere Land has been seen by few explorers, among whom are Greely, 1881–1884; Peary, 1898–1902; Sverdrup, 1898–1902; MacMillan, 1917; and Byrd and MacMillan by airplane in 1925. The main glaciation is grouped about two reservoirs, one in the central and the other in the southeastern quarter of Ellesmere Land. According to Greely (1886, p. 273), Grant Land, the northern fourth of the island, is ice free, except for a few small glaciers in the United States Range, which are not tidewater. Grinnell Land, the subdivision next south, embraces a shield of ice called "Mer de Glace Agassiz," traced from near Archer Fjord across to Greely Fjord and thence southeastwards where the ice discharges around the shores of Princess Marie Bay. The vicinity of Bach Peninsula, and a belt to the westward, is unglaciated. The discharging points for the northern reservoir, in order north to south, are Eugenie, Parish, Sven Hedin, and Benedict Glaciers.

The southeastern quarter of the Ellesmere Land, a plateau of some thousand or more feet in elevation, is known to be glaciated from the explorations and reports of Sverdrup (1904). Beginning at the

[45] Transehe (1925, p. 396) has discussed the question whether a large proportion of the icebergs sighted around Nicholas II Land in 1913 and 1914 did not consist of bergs that had been carried there from Novaya Zemlya and Franz Josef Land by an extension of the North Cape current. The evidence being so meager causes the discussion, although stimulating, to be more or less hypothetical.

head of Beistad Fjord and proceeding out along the south side of Hayes Sound we pass Sands, Bjorne, Soesterne, Stygge, Svarte, and Fram Glaciers, the latter of which caps the point opposite Sabine



THE TIDEWATER GLACIERS OF ELLESMERE LAND

FIGURE 34.—Ellesmere Land supports two main reservoirs of inland ice from which a total of 24 sizable glaciers discharge along the eastern coast. The number of sizable icebergs calved annually is estimated to be the relatively small number of 150.

Island. The mountain peaks along this coastal foreland rise high above a dozen or more unnamed smaller valley glaciers giving the section a very beautiful alpine landscape. MacMillan (1918) (1928) states that he counted and mapped 42 tidewater glaciers between Cape

Sabine and Clarence Head. Unfortunately no map of his work has ever been published, but from various other sources (Peary, 1903) and (Sverdrup, 1904) we have been able to locate several which in order, north to south are: Lefferts, Alfred Newton, Bliss, James, Sparks, American Museum, John Ross, American Geographical Society, and several unnamed glaciers. Three unnamed glaciers lie south of Clarence Head. MacMillan (1918, p. 302) states that the whole coast line around Boger Point, latitude 77° 25′ N., is one vast piedmont glacier, this the one named in honor of the American Geographical Society (see fig. 35); it measuring about 20 miles across the front. Deeply penetrating the coast between Boger Point and Clarence Head, Constable Makinson of the Royal Canadian Mounted Police (see Annually, 1928) has recently discovered and mapped a spacious inlet (see fig. 34). Inside of Clarence Head, that is in a northwesterly direction, the coastal foreland is low and covered with several unnamed glaciers. The southern coast of Ellesmere Land, forming the northern side of the 35-mile wide Jones Sound, is more or less glaciated but few, if any, tongues extend to the sea except the Botn Glacier near South Cape. The east and south coasts of Ellesmere Land, according to the foregoing survey, contains an estimated total of about 60 tidewater glaciers.

AN ELLESMERE LAND GLACIER

FIGURE 35.—The largest known tidewater glacier in Ellesmere Land, which occasionally discharges a few icebergs of small size into Smith Sound. This glacier, named in honor of the American Geographical Society and located in latitude 77° 30′ N., has a low front 20 miles in length. Its iceberg productivity when compared with some of the glaciers of west Greenland is negligible. (Photograph by D. B. MacMillan.)

The extent of the Ellesmere Land ice caps with the number, size, and rate of movement of its glaciers is an important question. It would be very interesting to know, for example, whether or not Ellesmere Land contributes materially to the supply of North Atlantic icebergs, and also in what proportions. The fact that the eastern coast of this island is located at the northern end of the great ice stream to lower latitudes, offers any icebergs that are discharged there a direct route for 1,700 miles downstream to the Grand Bank. Little or nothing has ever been published on the iceberg productivity of Ellesmere Land. MacMillan who followed the

entire coast from Sabine Island to Clarence Head has informed me that in his opinion only one glacier in this district is sufficiently active to discharge any bergs, namely, the American Geographical Society Glacier, located at Boger Point. But even this one is quite sluggish and of low front.[46] MacMillan (1928, p. 429) also gives an estimate of 3 feet per day as the rate of an Ellesmere Land glacier. Except for one clue, we are also equally ignorant of the iceberg productivity of the glaciers from Cape Sabine northward. Sverdrup (1904, supplementary map) and Peary (1903, supplementary map), both of whom have visited the region around Buchanan and Princess Marie Bays, report the presence of a considerable number of scattered bergs. The general funnellike configuration of Smith Sound, and the consequent course of the circulation, more or less precludes the drift of bergs to this locality from points south of the seventy-eighth parallel. And, if that source is eliminated, the only other distant point of discharge from which these bergs could have come is the Humboldt Glacier on the opposite side of Kane Basin. But the probability of bergs from that quarter is also slight, for the Humboldt is penned in by sea ice for several years at a time. It appears from the foregoing that the most logical sources of the icebergs found scattered in Ellesmere Land coastal waters are the local glaciers. It would not be surprising, therefore, to learn that Ellesmere Land contributes a significant number of bergs to the North Atlantic quota. According to MacMillan (1918, p. 302) one Ellesmere Land glacier, in latitude 77° 10′, is advancing, as ice was found spreading out from the shore there, overriding Saunders Island, which on the British Admiralty Chart of 1853 is separated by 4 miles from the coast.

Glaciation in Axel Heiberg Land

The southern part of Axel Heiberg Land, called Easter Land, is ice capped on the National Geographic map of the Arctic regions (1925), but Mecking (1928, p. 227) states that glaciation is almost entirely lacking there. Our present-day knowledge of this land is very meager, but it is doubtful if any icebergs are produced. Its inaccessibility to the Labrador current, moreover, makes it of little interest in the present discussion.

Glaciation in Devon Island

The eastern third of Devon Island, rising to an altitude of 3,000 feet, is glaciated. According to Low (1906, p. 50), the glaciers are most numerous around Croker Bay on the southeastern coast, where a few discharge into Lancaster Sound, even as far west as Cumming Creek. Further west the inland ice recedes from the coast. In the papers relating to the search for Sir John Franklin, compiled by the British Government, references are made to a glacier on the northwestern extremity of the island, also to three others along the western coast. It appears, therefore, that Devon Island is ice capped in its central portions, sufficiently at least, to send out a few glaciers

[46] Lauge Koch in a letter to me says: "The Eskimos in the Cape York district informed me from observations made during bear hunting along the east coast of Ellesmere Land that opposite Cape Parry (Greenland) there is a rather large, fairly productive glacier." This piece of information agrees fairly well with MacMillan's account. Cape Parry, Greenland, being in about the same latitude as the American Geographical Society Glacier. The Eskimos, however, apparently believe the American Geographical Society Glacier to be more productive than does MacMillan.

to the western shores. But practically nothing is known regarding the productivity of these. It is safe, however, to say that its western glaciers are of little consequence as far as contributory bergs to Baffin Bay, but the berg character of the glaciers debouching into Croker Bay is not so certain. The British Admiralty Chart of the Arctic Sea dated 1853 shows a tidewater glacier on the east coast of Devon Island, just north of Philpot Island.

Glaciation in Bylot Island

The proximity of this island to Baffin Land really makes it fall in the same class as the latter, as far as glaciation is concerned. Mountain ranges 2,200 feet in altitude slope north and south, providing a central ice cap and valley glaciers. It is almost certain, however, that none of these reach sea level, thus eliminating Bylot Island as a possible source of iceberg discharge.

Glaciation in Baffin Land

Northern Baffin Land between Admiralty and Navy Board Inlets supports an ice cap but the chart of this region does not show any outlet of tidewater glaciers. Low (1906, p. 60) describes a glacier at the head of Erik Fjord, in Eclipse Sound, a mile in width. Its motion is said to be slow and despite its 100-foot high front, the few bergs calved are small. Rasmussen (1926, p. 554) has reported the presence of an ice cap in the highlands east of Albert Harbor, and according to the Eskimos, there is a still larger glaciated area back of Scott Inlet in longitude 71° W. This ice is apparently wholly inland, as no glaciers are known to extend down to sea level. Small local glaciers however, have been observed in a few places, for instance, north of Low Point and at Oliver Sound which faces southeastward toward Eclipse Sound. The ice cover of Baffin Land proper is heaviest on the eastern side of the midland ridge which runs lengthwise of the island, but even here the glaciation is deeply serrated in places, and many bare lower areas lie in the foreground. Several glaciers, according to Mecking (1928, p. 216) descend on each side of the backbone of ice on Cumberland Peninsula, but no icebergs of consequence are produced. Stefansson (1912, p. 13) says "the glaciers of Baffin Island are comparable in size to the glaciers in British Columbia." Capping the ridge between Frobisher Bay and Hudson Strait, lies the 81-mile Grinnell Glacier from which, in longitude 68° W., one barren alpine tongue descends to Hudson Strait. This is the southernmost glacier of any size on the American side of Davis Strait, Labrador having only small patches of ice in the high mountain cachements of its northern section.

Summarizing the American side of Baffin Bay and Davis Strait: none of the areas of glaciation there, viz, Ellesmere Land, Axel Heiberg Island, Prince Patrick Island, Melville Island, Devon Island, Bylot Island, Baffin Island, or Labrador are important sources of icebergs,[47] except Ellesmere Land, and little information is available as to the rate of productivity of the latter. It is estimated that 150 bergs are produced annually from the American side of Baffin Bay and Davis Strait.

[47] Mecking (1906, p. 62) comments as follows: "In Americanischen Arktischen Archipel sind zwar auch einige Gletscher gefunden, welche Eisberge abwerfen, so Z. B., Von Kane auf Cornwallis, von Sverdrup im Heureka-Sund und der Westseite des Smith-Sund, aber auch diese Quellen treten ganzlich in dem Hintergrund."

Glaciation in Greenland

Greenland, sixth largest of continents, discharges every year something like 15,000 sizable icebergs. The continent, with the American Arctic Archipelago, is so placed that it nearly blocks off polar communication on the west, but with Spitzbergen to the east it forms a connection between the Atlantic and the polar basins, 245 miles in width.

THE TOPOGRAPHY OF THE GREENLAND ICE AND A TRANSVERSE SECTION

FIGURE 36.—The ice cap is composed of two and possibly three domes, all of which obtain the greatest altitude in the south, with more gradually decending slopes facing the north. (Figure after de Quervain and Mercanton, 1920.)

The general physiography of Greenland and of its ice sheet, the points of iceberg discharge, and the approximate number of bergs produced in time and place are discussed in the following pages but other subjects not so closely related to icebergs, such as the origin of the inland ice, its temperature, structure, snow limits, mechanics of motion, etc., are omitted as leading too far afield.

Our present knowledge of the inland ice comes from the records of various short excursions toward the interior, and also from nearly a dozen complete crossings between the east and west coasts. If we exclude the northern unglaciated section, composed of sedimentary layers, practically all of the plateau on which the inland ice lies is of gneiss, sloping gently toward the northwest. Basalt outpourings of the Tertiary period are visible around the unglaciated fringe on the western side at Disko Bay and on the east coast in the vicinity of Scoresby Sound and Jameson Land. Koch (1928, p. 438) constructed longitudinal elevations of the continent, using as ordinate values the heights of the land along each coast. These profiles plainly show the general shape of the land border as it slopes from the south toward the north, but probably the most important physiographical point revealed on them is the V-shaped cut midway

between Cape Farewell and the opposite northern extremity. A topographic map of the ice sheet, moreover, shows a fairly good agreement with the coast profiles, establishing beyond reasonable doubt that Greenland, as well as its ice cap, is crossed from Disko Bay to near Angmagssalik by a great transverse depression. This continental crease is believed to be of great importance in converging the ice and in concentrating the maximum discharge of icebergs near the seventieth parallel of latitude on both coasts.

Knowledge of the topography of the ice sheet is not yet complete. Apparently there are two main domes of glaciation, one in the north-central part and the other in the south-central part with altitudes of 9,750 and 8,775 feet, respectively. A third center of doubtful altitude and position is believed to lie in the region north of Angmagssalik, but in any case it does not interfere with the major depressional feature which appears to be well established. An accurate east to west profile of the ice cap made by de Quervain shows a convexity from the thickest portions which lie slightly to the eastward of the continental median line. The ice is fed mostly in the high mountains of the east, which are themselves submerged by inland ice even to their tops. The west side of Greenland, on the other hand, is the region of dissipation, both because of the topography and of the warm climate, the mean annual temperature of the west coast of Greenland being much higher than is that of the east. In general contour, the cap is noticeably flattened in the interior, but near the edges the slope is steeper, the eastern descent being short while the western slope is longer. Thus the main mass of ice moves toward Baffin Bay and Davis Strait. Preliminary seismic soundings made on the inland ice of west Greenland (see Wegener, 1930) indicate that although the ice rises toward the interior, the continental bedrock is depressed there. The fjord systems of Disko Bay and Northeast Bay, between parallels 69 and 72, extend inland under the cap for a known distance of at least 100 miles and perhaps farther, the formative effect of this topography being clearly traceable in the configuration of the ice surface. The movement of the ice upon reaching these drainage basins is noticeably increased through the various valley systems, as many active glaciers feeding into the sea. The volume of the annual discharge from the west coast fjord system is estimated to be equivalent to that of the river system before glaciation prevailed.[48]

Brooks (1923, p. 445) using de Quervain's and other data concludes that the Greenland ice cap is not a relic of the ice age, but is fully maintaining itself at present. He shows that the snow line, situated about 50 miles in from the western edge, divides the cap into a central interior called the accumulator, and a marginal dissipator. The accumulator is estimated to receive annually through precipitation an increment of ice 36 centimeters in mean thickness. The reduction of thickness in the dissipator ranges from a maximum of 2 meters around the edge of the cap to zero at the snow line, or an average of 95 centimeters over the whole width of the 50-mile margin. The wastage takes place in two ways—75 per cent is ablation by insolation and wind and 25 per cent is attributed to calvings of the tide-

[48] Mecking (1928, p. 251) gives an excellent description of the actual appearance of the surface features of the inland ice.

water glaciers. Little definite change in the extent or thickness of the inland ice appears in historic records, but there is geological evidence in the position of erratics around Disko to indicate that glaciation was once much stronger there than it is to-day. The marginal advances and retreats of the inland ice are slow with no two localities exhibiting exactly the same characteristics. Jacobshavn Glacier, in Disko Bay for example, has, except for one interruption, shortened by about 10 miles, in 22 years, 1880–1902, while during this same period, Brother John Glacier in northwest Greenland has been reportetd as advancing. Wegener (1930) found that during the last 35 years a considerable retirement had occurred along the fronts of the Disko and Northeast Bay outlet glaciers from the inland ice, but small isolated glaciers, e. g., those of Nugsuak Peninsula, had not changed materially.

The topography of the land bordering the outer margin of the ice cap governs the general form, size, and rate of production of icebergs. The inland ice, in some places, worms its way down to the coast through deep-carved fjords while in other sections it overflows all land features directly into the sea. The outward expansion from the vast central reservoir results in marginal wastage, ranging from melting by the sun and air to the calving of huge masses, millions of tons in weight. Notwithstanding the great length of the fragmenting periphery, we on the *Marion* expedition were struck by the relatively few definite iceberg producing points. Quantitatively, it is estimated that 10 to 20 per cent of the solid annual output from the ice cap is released as sizable icebergs. Along the coast of west Greenland, for example, only 22 of the 100 and odd glaciers are known to be iceberg producing with one group of these, 6 in number, accounting for 70 per cent of the total annual output. As yet we know very little about the conditions that would transform a tidewater glacier with a small discharge into a point of great productivity, or vice versa. This is not wholly a question of mass or of rate of motion, since the Ekip-Sermia Glacier we had the opportunity of visiting in the Igdloluarsuit, Disko Bay, west Greenland, possessed both these qualifications, yet no sizable pieces of ice were calved. An important part must be played by the ice fjords themselves, because some of the best known iceberg producers are characterized by spacious, well-worn beds that mold themselves smoothly into easy, outer embayments across the coastal foreland. The boldly cut incisions along the western side of Greenland between the sixty-ninth and seventy-second parallels, the heart of the iceberg country, provide deep-draft channels from which the bergs have easy access to Davis Strait and to Baffin Bay. The production of icebergs we believe does not necessarily follow from the presence of an ice cap even that the size of Greenland's; an essential condition is topography of a sort providing favorable debouchment.

In studying the degree and rate of iceberg productivity of tidewater glaciers it is convenient to divide Greenland into six districts, as follows:

East Greenland	Northeast Foreland to Cape Farewell.
North Greenland	Independence Fjord to Cape Alexander.
Westernmost Greenland	Cape Alexander to Cape York.
Melville Bay	Cape York to Svartenhuk Peninsula.
Northeast and Disko Bays	Svartenhuk Peninsula to Egedesminde.
South Greenland	Egedesminde to Cape Farewell.

The outstanding feature of the east Greenland district is the degree to which topography interferes with the movements of the bergs. The east coast glaciers, many of which are reported as iceberg producing, are greatly modified by a submerged threshold which bottles up all but the very small bergs until they have partially or completely disintegrated. Icebergs which are discharged from the deeper fjords of east Greenland are also often blockaded by pack ice which prevents their escape during a large part of the year. Lastly the wind (Greenland lies on the northwestern side of the great "Icelandic minimum") tends to blow the pack ice and to hold the bergs on shore.

The north Greenland glaciers average large, but dammed up by pack ice and frozen solid by "sikussak" (p. 24) no large bergs can calve. Most of the ice streams are backed by a flat, slow, spreading ice sheet which naturally has little tendency toward movement. The cap itself ends along miles of high precipitous walls, but where the glaciers extend out, they do so in broad, flat tongues the floating ends of which merge into the fast ice. No icebergs are ever formed under such conditions.

The westernmost district is characterized by a great number of small glaciers which have little movement. This sluggishness is due not so much to blockades from the sea, as lack of a head from the inland ice reservoirs.

Melville Bay glaciers are numerous and the ice fronts the sea in greater proportion than in any other region. A few of the glaciers exhibit a high rate of productivity but the littoral is beset by many off-lying skerries, and many years have been recorded when fast ice completely barred egress to the sea.

The Northeast Bay and the Disko Bay district is characterized as the most productive region of icebergs in Greenland. It is likewise known for three distinctive types of ice fjords:

(*a*) Glaciers which slope steeply down to the fjord. Example: Sermilik Glacier.

(*b*) Glaciers which flow down on to a foot forming a short step on the foreshore. Example: Great Karajak Glacier.

(*c*) Glaciers having a continuous easy descent from the inland ice to the sea. Example: Jacobshavn Glacier.

The bottom of the Jacobshavn Valley and upper fjord is featured by an easy gradual slope, an important condition we believe, which favors the production of large icebergs. The noted oscillation of the glacier, often stretching far out before calving, is another indication of the lack of disorderly fracturing. A glacier flowing down a steep and uneven declivity, on the other hand, suffers many interruptions to the stream flow, and only small pieces of ice float away. The many small-size fjords of southern Greenland usually have one or more glaciers discharging at their head. The narrowness and shallowness of most of this class, however, forms only small insignificant bergs, scarcely larger than growlers that seldom escape from the fjord.

EAST GREENLAND GLACIERS

Owing to the inaccessibility of the east coast of Greenland, our knowledge of its iceberg-discharging glaciers and of their productivity is less complete than for the west coast. First, many of the

bergs are held near their sources by intricate coastal cachments, and secondly, those that do escape seldom, if ever, drift so deeply into lower latitudes that they attain the more populated tracks of ships. There are also some basic differences in the ice line of the two coasts. The inland ice crowds closer to the sea along the southern half of the east coast than it does along the west; on the other hand, the east coast north of the seventieth parallel exhibits a wider land fringe than does the other side in the same latitude.

The largest and most productive glacier in east Greenland is said to issue from Kangerdlugssuak Fjord, near latitude 68° N., but the size and the number of icebergs it produces is unknown. Garde (1889, p. 228) lists the six most productive glaciers south of parallel 66 north, i. e., Angmagssalik, on the east coast as follows: Sermilik, Ikerssuak, Pikiutdlek, Igdlutarssuk, Tingmiarmiut, and Anoritok. Here again there are no data on the annual volume of discharge or the number of icebergs. All the glaciers between Kangerdlugssuak Glacier and Germania Land, a distance of over 500 miles, according to Kayser (1928, p. 414) terminate at the head of deep fjords. Many glaciers in Scoresby Sound produce massive box-shaped icebergs, but the shallow threshold across the fjord mouths imprison many and few escape to sea. North of Hudson Land and the seventy-fourth parallel the rate of marginal discharge of the galcial ice decreases rapidly partly on account of the slower movement of the inland ice and partly on account of the sea ice, sealing the glacier front. The interiors of some of the larger fjords, however, protected from the direct force of the pack, open up regularly every summer, and many icebergs break away from the glacier fronts. There is, nevertheless, only one glacier of this character in the northeast sector which rivals the production of the greater glaciers of the west coast, namely, Storstromen Glacier, in Dove Bay. The most active berg glaciers are well distributed along the coast from Cape Billie, latitude 62° 10′, to Scoresby Sound, in 70°, while northward the productivity markedly decreases. Apparently there is little difference in the total annual volume of discharge between the east and west coasts (7,500 bergs from the west coast), but the closer blockade of sea ice in the east greatly diminishes the berg supply to the Atlantic.

Drift and Distribution of East Greenland Bergs

The general drift of the icebergs from their source is southwestward along the coast to Cape Farewell, and the bergs may drift as far northward around the latter as Godthaab, just as the pack ice does. Those which remain out in the current along the continental edge travel the fastest, but vary in speed with the week-to-week or even the day-to-day pulsations of the current.[49] East Greenland bergs gather in greatest numbers during the summer; off Cape Farewell several hundred having been reported in sight of a ship at one time. The pack ice tends to hold them off the coast, but the effect of the earth rotation being in the opposite direction keeps them from spreading out to the North Atlantic. The van of bergs arrives at Cape

[49] Nielsen (1928, p. 226) states that the polar current on the continental side of the Greenland Sea is 10 to 14 miles per day, while closer in to the coast it is only half as great. Summer velocities are greater than winter ones. In autumn off Angmagssalik a speed of 5 to 16 miles per day has been recorded.

Farewell in April, where they are plentiful until August. They then decrease rapidly in numbers with autumn, and winter sees these waters more or less free. The only deviation from a course generally parallel to the coast is a string of bergs that are caught in the east Iceland current, to be carried to the vicinity of northern Iceland and possibly farther southeastward. They come mostly from the coastal glaciers north of Scoresby Sound, but a few of them may be from Spitsbergen. The number annually borne east of Iceland, however, is very small, partly because few are produced either in Spitsbergen or in northeast Greenland.

There are records of occasional bergs, "erratics" that wander from the better recognized paths of travel to be carried hither and thither in irregular tracks. Being relatively large and massive the processes of melting and erosion often fail to affect their destruction until they have completed long journeys. The greatest distance that east Greenland bergs have been reported off the coast by the Danish Meteorological Institute is 240 miles southeast of Cape Farewell. Probably several of the reports of ice sighted in the vicinity of the British Isles, or the Faroe Islands, rare phenomena but nevertheless authentic, refer to bergs that have drifted from northeast Greenland via the East Iceland current.

The 18 icebergs observed along the Greenland coast in the summer of 1928 by the *Marion* expedition between Cape Farewell and Disko Bay can be assumed, because of their position with regard to the current to have come from east Greenland and none to which that source could most reasonably be ascribed were found northward of Godthaab. The east Greenland bergs were distributed as follows: 2 lay about 20 miles southwest of Cape Farewell; 7 off Arsuk Fjord; 8 off Fiskernæs; and the northernmost one off Godthaab. Comparing this distribution with the track of the *Marion* (fig. 1), it will be noted that a few bergs were sighted at each point where the course approached the coast, but none offshore. The coast sectors, south of Godthaab, not visited probably contained icebergs also, so that a total of 50 bergs is probably a conservative estimate of the number of such bergs then present along that part of the coast. The total absence of any bergs more than 30 miles out from the coast was striking, as typical of their on-shore tendency.

Our knowledge regarding the probable movements of such bergs after passing the meridian of Cape Farewell has been placed on a far more certain basis as a result of the current survey which we carried out on the *Marion* expedition during the summer of 1928. According to Figures 95 and 96, pp. 147–148, bergs less than 30 miles off Cape Farewell will be carried northwestward, parallel to the coast, at a rate of about 14 miles per day. Their rate of travel would constantly increase until, off Cape Desolation, they are being borne northward along the coast at 22 miles per day but with the increase in the velocity the width of the current decreases somewhat. Bergs on the outer edge of the current will, of course, not move as rapidly as those nearest its axis and off Cape Desolation. The ice receives its first opportunity to leave the coastal belt off Cape Desolation where a branch turns off to the left, developing into a broad, tortuous drift and losing speed proportionally. Icebergs so deflected will move slowly to the westward along these paths at only 4 to 6 miles per day. The chart shows that any ice which holds to the coast continues northward in

a current band about 25 miles in width at an average rate of 11 miles per day. Off Fiskernæs the water on the outer side of the current again diverges in a narrow band about 15 miles in width and with a velocity of 6 miles per day finally reaching across to the American side of Davis Strait. The *Marion* found the current to run at 13 miles per day just off Godthaab but a short distance farther north there was less dynamic tendency toward movement, suggesting a general slackening and expansion near the latitude of Sukkertoppen.[50] The northern terminus of the ice and current at Sukkertoppen is probably due somewhat to the broadening of the continental shelf of Little Hellefiske Bank, which tends to scatter the current and hold back the ice. Our hydrographical survey indicates that icebergs in the current will be carried northward along the west Greenland coast at the average rate of 15 miles per day, or in one month travel from Cape Farewell to Sukkertoppen.

According to these calculations icebergs travel much faster along the west Greenland coast than they do along the Labrador side or, according to previous scanty information, than they do along the east Greenland coast. At the same time it should be remarked the current along west Greenland is narrow, and unless a berg remains within its bounds it will not follow this drift. The boundary between the main current and the main body of water to the west of the latter is sharply defined, and since it is unlikely that a berg will keep within such limits because the disturbing factors, such as the effect of sea ice, of coastal promonotories, of cachments, of gales, etc., we may assume that many east Greenland bergs are carried inshore, to strand, while many others scatter out from the outer side of the current. In the latter case they come into a sort of dead water where they will disintegrate eventually without having made any material progress one way or the other. Only at two points, at Cape Desolation and near Fiskernæs, is there any branching of the ice, and even there such a tendency is slight. Bergs which follow such offshore dispersals lose speed very rapidly and wind along tortuous trails.

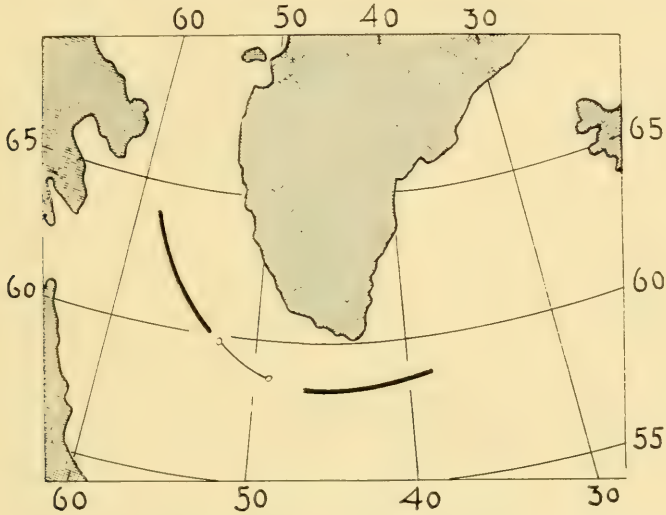
It should be noted that the positions of bergs as sighted by the *Marion* expedition along west Greenland are all practically within the bounds of the current, and places where our dymanic map shows no current, those areas were for the most part berg free.

The conditions under which one particular berg was sighted by the *Marion* off Godthaab afford instructive evidence as to the behavior and distribution of east Greenland bergs, supporting the conclusions first reached from the dynamic current map. As we stood out into Davis Strait from Godthaab Fjord, a medium-sized berg was sighted dead ahead, distance 20 miles from the coast. We noticed on approaching that the course had to be altered continuously to the left in order to counteract an apparent set of the ship northward. The depth of 60 fathoms at that spot indicated that the berg was either grounded or at least that it extended downward very nearly to the bottom so that it was difficult to escape the conclusion that a surface

[50] In constructing a dynamic topographic map the motion on the chosen surface (in the case of fig. 95), the sea surface is compared with that on a plane where the water is at rest. If the water, therefore, surface to bottom, is moving at the same or nearly the same velocity, no basis of comparison is possible and the dynamic topographic map will show no isobaths where in fact there is a current. There is of course the possibility such may have been the condition in the vicinity of Sukkertoppen the first week in August, 1928.

current was setting strongly toward the north. This berg, therefore, and the 17 others seen to the southward, could not have drifted there from the northern or western sides of Davis Strait but must have come from the east coast of Greenland via Cape Farewell.

We occasionally read statements such as the following translated from the Deutsche Seewarte Segelhandbuch (1910, p. 296): "The ice girdle along the west coast of Greenland spreads out northwestward crossing over Davis Strait and then sets south along the Labrador coast following the path of the current toward the Newfoundland Banks." Johnston (1915, p. 40) says that many of the icebergs sighted east of Flemish Cap in the North Atlantic are east Greenland bergs carried there by one of the branches of the cold current. In discussing the behavior of pack ice it has already been shown that there is little likelihood of any east Greenland ice reaching over to



THE WESTERNMOST LIMITS OF EAST GREENLAND ICEBERGS

FIGURE 37.—The most westerly positions in which icebergs from the glaciers of east Greenland have been sighted around Cape Farewell. The heavy line marks the position of bergs in July, 1922, and the slender line, bergs in June, 1917. (From records over a long period kept by the Danish Meteorological Institute.)

American waters. The main obstacle to such a journey is not so much lack of transportation as inability to survive long enough in the relatively warm off-shore waters. Icebergs being of large bulk and mass are, however, able to withstand the process of melting for a much longer time than is sea ice and therefore are occasionally found in places very remote from their sources. Thus the files of the Danish Meteorological Institute show, such as April, 1913, a berg was sighted about 200 miles west-southwest of Ivigtut, Greenland, a position about halfway across Davis Strait. Again in June and July, 1917, bergs were sighted in latitude 59° 30′, longitude 51° 00′, and latitude 59° 30′, longitude 52° 00′. These positions when plotted on the dynamic current map, Figure 95, directly coincide with one of the southwestern branches and while the extraordinary long journeys accomplished by icebergs in rare instances forbids any positive

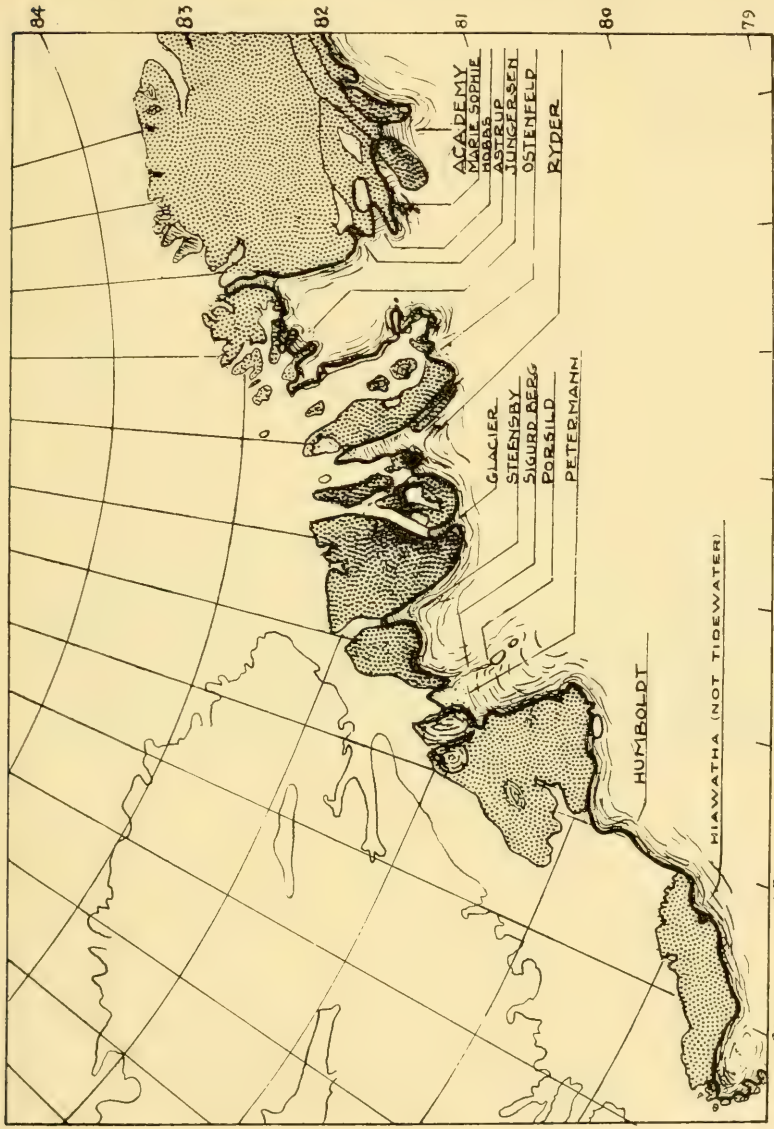
denial, the current maps of the Labrador Sea now made available as a result of the *Marion* expedition, permit us to state definitely that for an east Greenland berg to cross Davis Strait is certainly an exceptional occurrence. The westerly drifts, it will be seen, progress in broad tortuous bands at the very sluggish rate of approximately 5 to 6 miles per day. In other words, it would take four months to cross Davis Strait opposite Ivigtut and two months opposite Godthaab. If the first of the season's bergs which arrive at Cape Farewell in April were to remain in the axis of these currents, some of them might reach Labrador, near Nain, or opposite Hudson Strait by the middle of August; if following the northern route, by the first of July. But the likelihood of such "strays" keeping always in the currents is so small that probably they would be "lost" somewhere in transit. Furthermore, the surface water in the central part of Davis Strait is then so warm that any berg coming into it would soon melt. As for east Greenland bergs penetrating directly south of Cape Farewell into the Atlantic, no such distribution is supported by our present knowledge of current or of winds. The temperature of the oceanic water, furthermore, through which bergs would have to travel for a distance of more than 900 miles in order to appear off Flemish Cap precludes any such behavior of the ice.

NORTH GREENLAND GLACIERS

Considering how large and numerous are the north Greenland glaciers, their production of icebergs is insignificant. Probably the greatest barriers to the discharge and distribution of bergs for them is the heavy pack ice which tightly seals these glacier fronts for years at a time. This ice blockade may run in age all the way from sikussak, to paleocrystic forms tightly held in the regions on the northwest coast; or simply to old sea ice that has been lodged in the fjords and across the entrances. The constant pressure of sea ice against the glacier fronts, the low mean annual temperature, the shallow gradient of the ice, and the amount of precipitation combine on many of the ice tongues to develop a floating end which while common in the Antarctic is not found elsewhere in the Arctic.

There are 13 glaciers in this district, as follows, from east to west: Academy, Marie Sophie, Astrup, Hobbs, Jungerssen, Ostenfeldt, Ryder, Steensby, Sigurd Berg, Porsild, Petermann, Humboldt, and Hiawatha. The glaciers of Independence Fjord—Academy, Marie Sophie, Astrup, Hobbs, and Steensby—detach a considerable number of icebergs which lie packed and cemented together perhaps for several years and then, once in 15 or 20 summers or so, break loose. Jungerssen Glacier, in Nordenskiold Fjord, like Academy is featured by a stream of coalesced icebergs. A few short alpine glaciers face the polar sea in Peary Land. Practically all of them are sealed by heavy pack ice but Koch (1928, p. 326) believes that one small glacier just east of Cape Cannon calves a few bergs. Discharge is assisted by occasional changes in the thrust of the sea ice, which for this particular place is sometimes athwart the glacier end. The rate of discharge is, of course, very meager and our interest lies mainly in the fact that this is the northernmost iceberg-producing glacier in the world. The four north Greenland glaciers with their floating ends are: Petermann, Steensby, Ryder and Ostenfeldt. Petermann,

measuring over 100 miles from the front, back to the 2,200-foot contour of the inland ice, is the longest glacier in the Northern Hemisphere. Its outer front in some places is only a few meters in height, or in other places it merges so gradually into the sea ice that the bound-



THE TIDEWATER GLACIERS OF NORTH GREENLAND

FIGURE 38.—There is a total of 13 sizable glaciers in this section, of which only 1, the Humboldt Glacier, is a significant producer of icebergs. It is estimated that this portion of north Greenland calves annually about 150 icebergs.

ary can not be detected. It is estimated that the length of the floating end of Petermann Glacier is about 20 miles. The ascent of the glacier is so slight in the great distance between its outer end and the inland ice at the fjord head, that early explorers were led to believe that a sound divided this northern section from the mainland.

Though enormous in size, these glaciers with floating ends produce no icebergs. Humboldt Glacier with its 60-mile wall of ice fronting on Kane Basin is the largest glacier in the Northern Hemisphere; in fact for many years after its discovery, it was believed actually to be the wall of the inland ice itself. This glacier is only productive over a short breadth of its northern side, where Koch in 1923 observed a collection of icebergs, probably a thousand in number, strewn out for a distance of 3 miles. They were so closely packed against the front in some places that it was difficult to determine where the glacier ended and the bergs began, and also they were apparently aground, suffering continual melting and disintegration during the summer. Throughout the year fast ice completely covers the glacier wall and stretches forward to a line from Cape Forbes

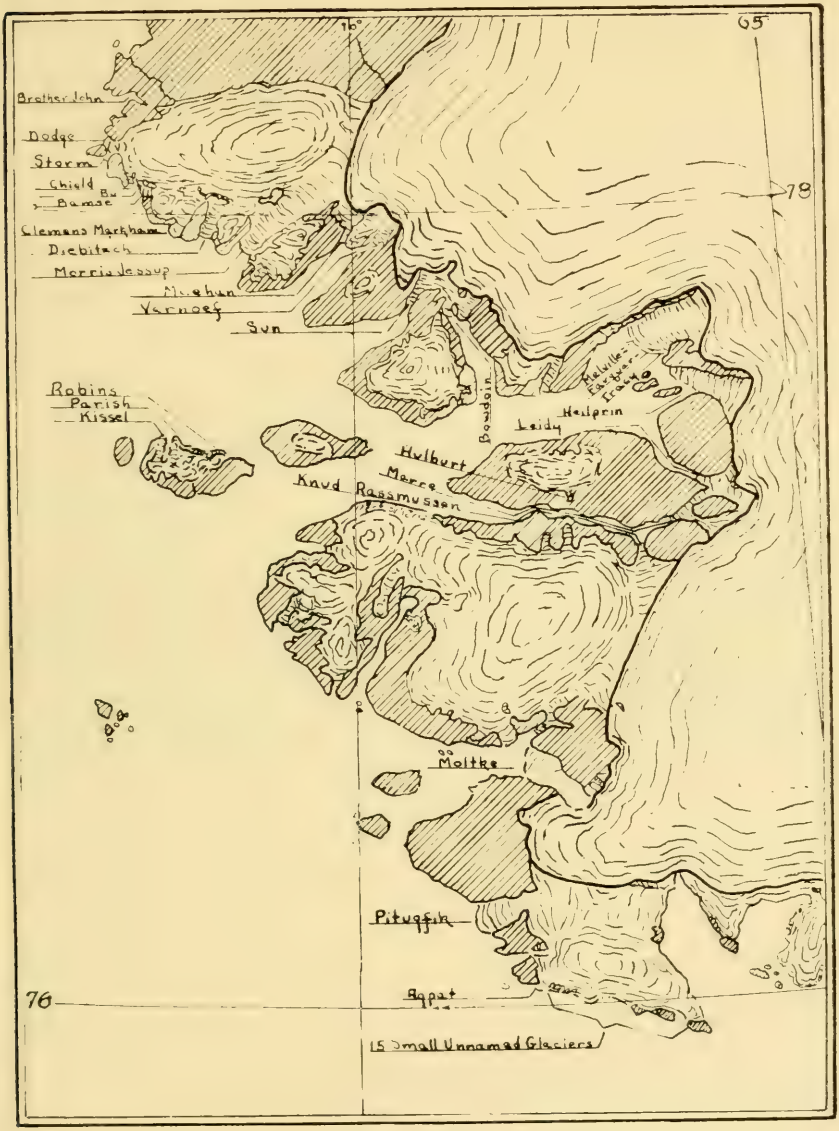
THE LARGEST GLACIER IN THE ARCTIC

FIGURE 39.—A section of the largest tidewater glacier in the north; Humboldt Glacier, With its 60-mile wide front of ice facing the waters of Kane Basin, northwest Greenland, early explorers first mistook it for the inland ice itself. This photograph shows a group of several hundred icebergs that are penned against the front of the glacier by the fast ice. Once in 20 or 25 years the fast ice is said to break up, allowing many of the bergs to move away in the currents. (Photograph by L. Koch, 1928.)

to Dallas Bay, but occasionally, perhaps once every five years or so, the sea ice more or less completely breaks up in Kane Basin, allowing many of the bergs which are not permanently grounded to drift abroad. Although the Humboldt is the largest glacier in Greenland, only a few of its bergs are believed ever to reach Baffin Bay, a conclusion which is supported by the fact that very few icebergs have ever been found adrift north of the seventy-eighth parallel. Hiawatha Glacier, the last one included in the north Greenland group, is merely a small tongue extending out from the inland ice into Inglefield Land, its end not even approaching close to the sea. In the north Greenland district we estimate that on an average 150 bergs are produced annually, most of them calved from one glacier—the Humboldt.

The Glaciers, Cape Alexander to Cape York

The glaciers situated between Cape Alexander and Cape York are so far north that the most of them are blocked by sea ice but



THE TIDEWATER GLACIERS FROM CAPE ALEXANDER TO CAPE YORK

FIGURE 40.—There is a total of 54 tidewater glaciers in this section of which only 25 are known as occasional iceberg producers. Most of the glaciers are sluggish and small. The important iceberg-discharging glaciers numbering only five are: Morris Jessup, Bowdoin, Melville-Tracy-Farquar, Moltke, and Pituqsik. The annual production is estimated to be 300 icebergs.

in the vicinity of Wolstenholme Sound this is not the case except during the winter months. A grand total of 161 glaciers, large and small, extend out from the ice cap along this highly irregular coast

line. But of these, 107—two-thirds of the total—do not even reach the sea, while the remaining 54 are at present classed as tidewater glaciers.[51] About 29 of these are apparently without motion, leaving 25 that may on occasions produce icebergs. The glaciers met from Cape Alexander to Cape York, are in order: Brother John, Dodge, Storm, Chield, Bu, Banse, Markham, Diebitsch, Morris Jessup, Meehan, Verhoeff, Sun, Bowdoin, Hubbard, Hart, Sharp, Melville-Tracy-Farqhar, Heilprin, Marre, Liedy, Knut Rasmussen, Moltke, Pitufig, and Nos. 1 to 16, Conical Rock to Cape York. Five of the foregoing, only 3 per cent of the grand total, account, furthermore, for the discharge of 80 per cent of all the bergs in northwest Greenland, namely: Morris Jessup, Bowdoin, Melville-Tracy-Farqhar, Moltke, Pitufig. The Eskimos, in commenting on the productivity of the glaciers in this district, Cape Alexander to Cape York, state that nearly all are stationary, hence little productive of bergs, and that condition has prevailed as long as they can recollect or learn through legends. We estimate that 250 to 300 bergs, about 5 per cent of the grand total drifting toward the Atlantic, are supplied from the glaciers of northwest Greenland.

The Glaciers, Cape York to Svartenhuk Peninsula

The west Greenland foreland is by nature divided transversely into two parts, the southern half, Cape Farewell to Disko Bay characterized, except for the Frederickshaab ice lobe, by a relatively wide land fringe, while the northern end, Upernivik to Cape York, is ice walled. The southern section owes its ice-free conditions largely to the decidedly higher mean temperatures which accompany the great warm-water masses from the Atlantic. As soon as we pass across the Davis Strait sill, however, into the region of icy Baffin Bay, climatic conditions abruptly change and Greenland's coast becomes extremely glaciated.

The inland ice reaches the sea in the northern section at more than 80 different places and so icy is the coast that all the principal land features are completely hidden for approximately 180 miles. The glaciers named in order from Cape York southward are: Helland, Wulff, Yugyar Nielsen, Mohn, John Ross, Gade, Döcker-Smith, Rink, Peary, King Oscar, Nordenskiold, Nansen, Dietrichson, Steenstrup, Kjaer, Hayes, Giesecke, Iodhulik, and Upernivik. Some idea of the massiveness of these ice streams can be obtained when it is stated that nine of them, viz: Steenstrup, Nansen, King Oscar, Peary, Mohn, John Ross, Wulff, Giesecke, and Upernivik are more than 5 miles wide across their fronts. Those which are especially productive of icebergs are: Steenstrup, Dietrichson, Nansen, King Oscar, Hayes, Giesecke, and Upernivik. The most productive one of them all, King Oscar, is believed to discharge as many as 500 icebergs a year.

Steenstrup glacier ranks largest with a front of 16 miles, the northern third of which is stated to be very productive. On account of the smooth contour of the marginal land, across which the glacier

[51] Koch (1922, p. 216) states that there are 16 alpine glaciers located along the slopes of Crimson Bluffs, Conical Rock to Cape York, but owing to their small size and slow rate of movement their yield is not equivalent in volume to that of more than one good-sized berg a year.

flows, its icebergs have a strikingly flat-topped, tabular appearance, averaging about 70 feet in height when floating. Nansen Glacier produces similar shaped bergs, even loftier. And Steenstrup and Nansen Glaciers are peculiar from any of the glaciers of the west coast in discharging such large blocks.[52] The coast between Hayes Glacier and Giesecke Glacier contributes very few icebergs, while the production of Giesecke and Upernivik Glaciers, it is also stated by Porsild[53] and Koch (1923, p. 53) has been unduly emphasized in early literature. Upernivik Glacier discharges a fair number of bergs, but they are mostly small and often ground in the offing of the skerries.

Melville Bay is furthermore marked with many banks which are sufficiently shallow to strand the bergs and thus detain them from drifting out of Baffin Bay. Such shoals are known to exist off Diet-

A GREENLAND ICEBERG FJORD

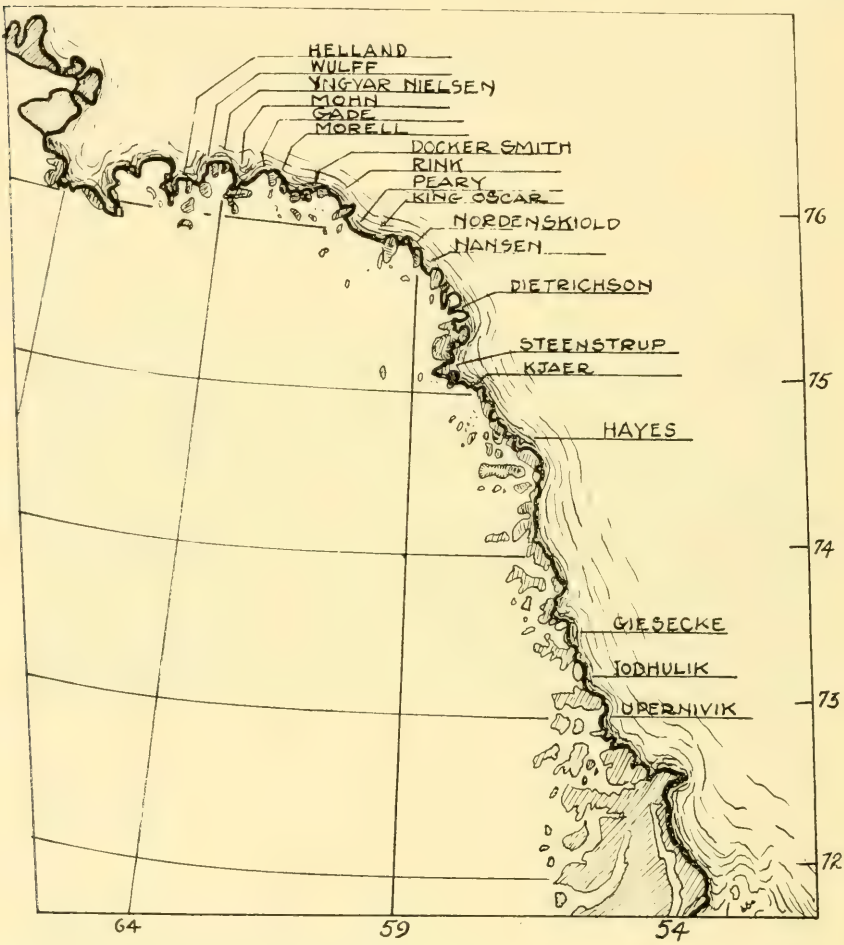
FIGURE 41.—Umanak Fjord looking east from the north shore of Nugsuak Peninsula. Umanak is one of the most important iceberg fjords on the west coast of Greenland. The fjords north of the seventieth parallel of latitude are normally covered by fast ice during the colder months, causing the icebergs to be held in the fjords until the break-up of the fast ice in June. (Photograph by A. Heim in Meddeleleser om Grönland, vol. 47.)

richson and King Oscar Glaciers because of the great number of bergs held there for protracted periods. The coast from Wilcox Head to Svartenhuk Peninsula is famous for its maze of off-lying islands, rocks and skerries, and the irregularity of the offshore ground, by tending to prevent the bergs from drifting out to sea greatly offsets the productivity of the glaciers in this section. It is estimated that the district Cape York to Svartenhuk Peninsula annually contributes about 1,500 bergs to the quota of Baffin Bay.

[52] The characteristic tabular forms often sighted by the ice patrol off Newfoundland have probably come from these two glaciers in Melville Bay. Ricketts (1930, p. 118) observed that the tabular-shaped bergs south of the forty-fourth parallel were as numerous as the more irregular-shaped bergs in 1929. As a rule the flat-topped ones are in the minority. Since 1929 was a rich berg year off Newfoundland, a possible cause may have been the release of larger than normal numbers of Melville Bay icebergs.

[53] In conversation.

The condition of the fast ice vitally affects iceberg production in Melville Bay. During a normal winter, fast ice will make out there to a line from Wilcox Head to Cape York, and it often happens that the ice does not completely break up during the next, or possibly during several summers. Thus the icebergs are free to drift out into Baffin Bay in some years from July to October, but in other years



THE TIDEWATER GLACIERS FROM CAPE YORK TO SVANTENHUK PENINSULA

FIGURE 42.—There is a total of 19 sizable tidewater glaciers in this section, out of which 8 constitute the principal producers of icebergs. They are in order north to south: Gade, King Oscar, Nansen, Dietrichson, Steenstrup, Hayes, Giesecke, and Upernivik. It is estimated that there are 1,500 sizable icebergs discharged annually from this Melville Bay region.

they may lie penned against the glaciers. Thus, Koch (1928, p. 200) in 1916 and 1917 observed many bergs jammed against the fronts of Steenstrup and Dietrichson Glaciers. In 1920, however, conditions changed, the fast ice broke up, and more than 20 square miles of densely packed icebergs were set free to drift offshore. Allowing a lag of one year for the bergs to drift from Melville Bay to New-

foundland, Koch's observations are compared with these of the ice patrol off the Grand Bank (see Smith, 1927, p. 76) in the following table:

Year	Melville Bay ice	Lag of 1 year at Newfoundland	
		Berg value	Berg count
1915	Much fast ice	2.8	54
1916	Do	2.5	38
1920	Little fast ice	6.8	746

It would be interesting to have a longer series of ice observations from Melville Bay to test this indicated relation between fast ice conditions there and icebergs a year later in the North Atlantic.

FAST ICE HOLDS ICEBERGS

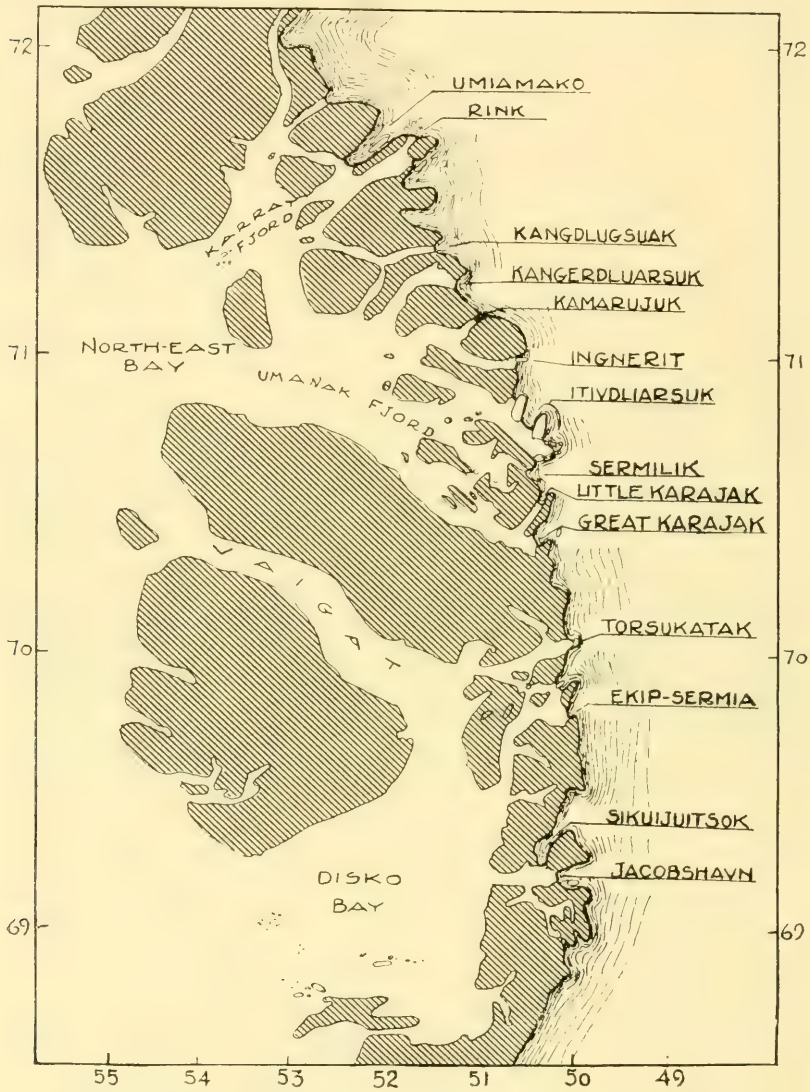
FIGURE 43.—Many icebergs are imprisoned in fast ice which may postpone their progress in the ocean currents throughout one or perhaps several winters in the far north. This photograph, which was taken about 2 miles west of Cape York, in northwest Greenland, shows several of the icebergs of Melville Bay securely held in the grip of the fast ice. Eskimos in the foreground are members of the famous Cape York tribe. (Photograph from Chamberlin, 1895.)

THE GLACIERS OF NORTHEAST BAY AND DISKO BAY

The Greenland coast, in this stretch of 160 miles, from latitude 71° 40′ to 69° 00′, includes its two deepest embayments marking the deltas of great river systems, the region pouring out over one-half of the total annual supply of North Atlantic icebergs. There are perhaps 100 glaciers, large and small, to be counted along these shores, the majority being small ones on Nugsuak Peninsula, but 14 are better known and important. These are (north to south): Umiamako, Rink, Kangdlugsuak, Kangerdluarsuk, Kamarujuk, Ingnerit, Itivdliarsuk, Sermilik, Little Karajak, Great Karajak, Torsukatak, Ekip-Sermia, Sikuijuitsok, and Jacobshavn.

Northeast Bay extends diagonally inland in two arms; into the head of the northernmost, called Karrat Fjord, Umiamako and Rink

Glaciers discharge.[54] Umanak Fjord, the southeasterly fork, receives the combined outputs of Ingnerit, Itivdliarsuk, Sermilik, and Little and Great Karajak.



THE TIDEWATER GLACIERS OF NORTHEAST BAY AND DISKO BAY

FIGURE 44.—There is a total of over 100 glaciers in this section of the west coast of Greenland, of which 12 are important iceberg producers. These 12 are estimated to discharge annually 5,400 icebergs, or more than any other coast line of similar extent in the Northern Hemisphere.

Disko Bay receives two very active ice streams, Torsukatak and Jacobshavn, the latter (according to Helland 1876, p. 108) producing

[54] Koch states in a letter that in the summer of 1928 he met more icebergs off Umiamako Glacier at the mouth of Karrat Fjord than at any other place in west Greenland so he believes this is, at present, the most productive point on the west coast.

about twice as many bergs as the former. There are two small glaciers just south of Torsukatak which flow into Igdloluarsuit, called Ekip-Sermia, but they calve only in small pieces and growlers, and therefore are termed harmless ends. Jacobshavn was formerly believed to be the greatest producer of icebergs, but in the last few years, Umiamako according to Koch, has been more productive, hence Disko Bay probably does not produce as great a quantity of icebergs as Northeast Bay.

Drygalski (1895, p. 369) states that the fjords of west Greenland freeze over in winter and that fast ice then prevents the bergs from drifting out into Davis Strait. While the inland ice itself has practically a uniform movement the release of the icebergs, on account of

SUMMER IN A WEST GREENLAND FJORD

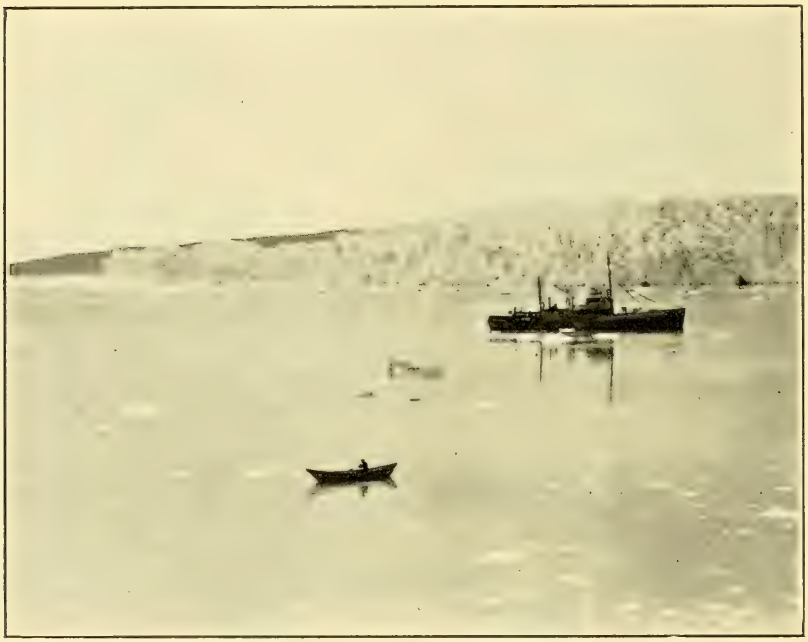
FIGURE 45.—A typical summer iceberg in a northern fjord of west Greenland. It is during this period when the fjord fast ice disappears that the bergs are free to drift out to sea. In the productive iceberg fjords several hundred icebergs are often simultaneously in view. (Photograph by A. Heim.)

winter fast ice, has a distinct seasonal period, the whole year's production being carried out of the fjords in this district during the open period July to December. In late spring, under increased solar warmth, the belt of fast ice which all winter has sealed the fjords begins to narrow and weaken. Finally the band breaks and the accumulation of ice and bergs pours out of the fjords so that the waters of Disko Bay and of Northeast Bay are choked with glacial flotsam for several weeks following such events. Drygalski (1895, p. 370) observed that in 1893 [55] initial iceberg discharge occurred at Jacobshavn about June 10; at Torsukatak 10 days later, and in Northeast Bay about July 1. During this period great swarms of

[55] Porsild (1919, p. 150) has described this phenomenon as more or less characteristic of all Disko Bay and Northeast Bay ice fjords, but in particular he identifies it with the behavior of Torsukatak Glacier.

bergs were met, so thickly infesting the coastal waters that a small boat had difficulty in navigating through the Vaigat. Upernivik district witnesses the release of the bergs about two weeks earlier than Disko Bay because, owing to its exposed position, the fast ice breaks earlier there.

The iceberg fjords of Disko Bay and of Northeast Bay and their discharging glaciers have been the object of more scientific study than have any other similar parts of Greenland. Hammer (1893), and Helland (1876), have made investigations of Torsukatak and Jacobshavn Fjords. Drygalski (1895) devoted more than two years' study to the glaciers in this district, especial attention being paid



A WEST GREENLAND GLACIER AND ITS ICE-STREWN WATERS

FIGURE 46.—Ekip-Sermia Glacier, which discharges into a tributary of the Atta Sound, west Greenland, latitude 69° 40′ N., longitude 50° 05′ W. As we approached this locality the color of the surface water took on a shade of light, milky green, due to the great amount of "rock flour" ground out along the glacier bed. A darker earth material lay suspended in the deeper layers of the fjord, leaving a muddy brown wake stirred up by the passage of the ship. Thousands of kittiwakes fed in the ice-strewn waters. This glacier, unlike Great Karajak and several others, is known as calving no ice of berg size. (Photograph by Boatswain J. B. Krestensen.)

to Great and Little Karajak, Itividliarsuk, Sermilik, Rink, and Umiamako.

Great Karajak Glacier, which debouches into the fjord of that name, exhibits a front about 3 miles in width, and rises approximately 260 feet—in the middle, 300 feet—above the surface of the fjord. The ice moves fastest at the outer end of the tongue, near its middle, where rates as high as 60 to 75 feet per day have been recorded, but 35 to 60 feet is more representative. The rates of each particular glacier, of course, vary, but some idea of the rapidity of flow of the west Greenland ice streams may be gained when they are compared with the fastest glaciers of the Alps, which travel about

U. S. COAST GUARD PATROL BOAT "MARION" IN FRONT OF A WEST GREENLAND GLACIER

FIGURE 47.—The United States Coast patrol boat, *Marion*, August 9, 1928, a few hundred yards off the glacier shown on Figure 46. The front wall was estimated to rise about 70 feet above the level of the fjord. The rough and jagged top surface of the glacier indicates the formative stresses and strains set up by the ice stream as it moves across the foreland to the fjord. Several projectiles were fired from the *Marion's* 3-inch gun causing little apparent fragmentation of the front wall. (Official photograph of the *Marion* expedition.)

SIDE VIEW OF A WEST GREENLAND ICEBERG

FIGURE 48.—A side view of the glacier shown on Figure 46, looking northward. The *Marion* can be seen far below in the left foreground. This anchorage is known as Quervainshavn, in memory of the leader of the Schweizerischen-Gronland expedition, who departed from this point to cross the Greenland ice cap in the summer of 1912. A rock cairn in the foreground just below the brow of the hill marks the encampment of the expedition. (Official photograph, *Marion* expedition.)

FRONT VIEW OF A TIDEWATER GLACIER AT A DISTANCE OF ONE-HALF MILE

FIGURE 49.—A view of about one-half mile distant from the same glacier shown on Figure 46. The heights of the front wall in this view is dwarfed by the altitude of the upland, about 700 feet, from which the glacier in a distance of a mile or more descends to the level of the fjord. Note also the manner in which the glacier spreads out fan shape after passing through the gateway of the headlands. (Photograph by Boatswain J. B. Krestensen.)

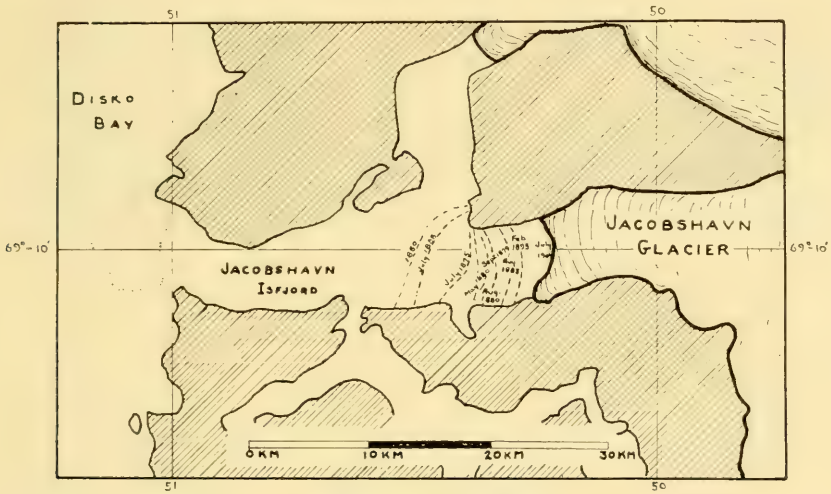
THE DISCHARGE OF THE INLAND ICE

FIGURE 50.—Ekip-Sermia Glacier near Atta Sound, west Greenland, visited by the *Marion* expedition the summer of 1928. In the right foreground lies a lateral moraine formed by the sweep of the glacier on its path to the fjord. (Photograph by Boatswain J. B. Krestensen.)

4 feet per day. Many of the great Greenland glaciers, in other words, move ten to fifteen times faster than those in the Alps. About four times a month during the summer, usually accompanying spring tides, 800 to 1,000 feet of the fjord end break off in great calvings; Drygalski (1895, p. 401) estimates that the output of Great Karajak is approximately 15.3 cubic kilometers per year, and we have estimated the annual number of sizable bergs calved as approximately 1,200. This figure is believed to be conservative, but, of course, somewhat speculative although it can be hardly over 25 per cent in error.

JACOBSHAVN FJORD AND GLACIER

Jacobshavn is one of the best known and interesting of all Greenland's glaciers. The *Marion*, in August, 1928, cruised in amongst the



JACOBSHAVN GLACIER

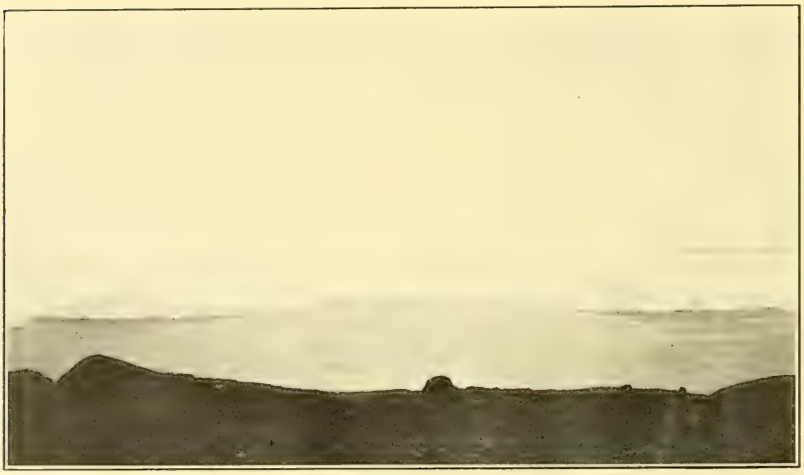
FIGURE 51.—Jacobshavn Glacier, located on the west coast of Greenland, is estimated to discharge annually 1,350 sizable icebergs into Disko Bay. It is one of the most productive iceberg glaciers in the world. The position of the front wall is shown from 1850 to 1902. Recent observations indicate a general retreat of many of the glaciers in this section of Greenland. (After Hammer and Engell.)

bergs that clogged the mouth of the fjord; as those within were joined so closely together that not even a ship's boat could pass between them. We had another view of these bergs when we landed at Jacobshavn and crossed the hills to look down from a height into the fjord itself. Jacobshavn Fjord is a more or less straight trough-like depression, possibly a rift valley, 3 to 4 miles in width and extending inland a distance of 15 miles where it meets the high front of the glacier. As far as the eye could see in the fjord icebergs were packed tightly row upon row, a majestic procession marching toward the open sea. It is estimated that on August 9, 1928, 4,000 to 6,000 bergs, 3 or 4 years supply, perhaps more, which had calved from the glacier front, lay in the Jacobshavn Fjord.

At uncertain intervals, approximately ten times a year, and without warning, so say the natives, this iceberg train moves. It starts

HUNDREDS OF BERGS JAMMED IN THE MOUTH OF JACOBSHAVN FJORD

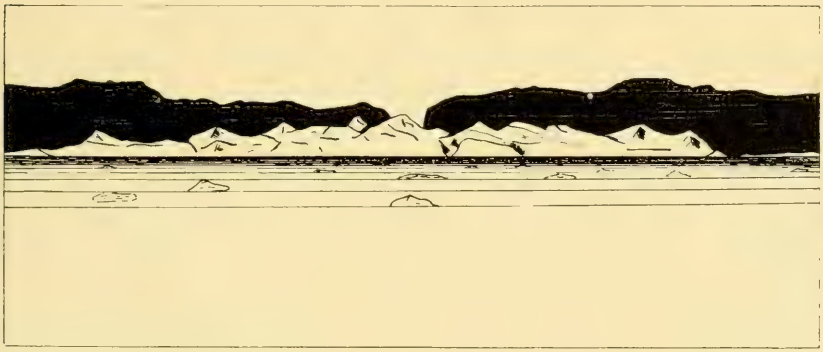
FIGURE 52.—The patrol boat *Marion* cruising in among the icebergs that clogged the mouth of Jacobshavn Fjord August 8, 1928. The bergs were packed so tightly together that it was impossible to penetrate much beyond the outermost ones. (Official photograph, *Marion* expedition.)



TRAIN OF ICEBERGS MOVING OUT OF JACOBSHAVN FJORD

FIGURE 53.—Looking southward across the 3-mile-wide Jacobshavn Fjord about 2 miles above its mouth. Jacobshavn Fjord empties into Disko Bay, latitude 69° 10′ N., longitude 50° 19′ W., on the west coast of Greenland. This train of icebergs packed closely end to end and side to side is continually passing down the fjord and out to sea. (Official photograph, *Marion* expedition.)

slowly but rapidly gaining momentum, the ice attains the incredible speed of 5 to 8 miles per hour—as fast as a fox can run, affirm the Greenlanders—to the accompaniment of a deafening roar that can be heard for miles, and that sometimes lasts for days. Sealing nets that the natives stretch under the water between the bergs are quickly torn asunder, thus in a few moments a great part of the community's investment in fishing may be swept away. Nobody can foretell the approach of the catastrophe. This phenomenon unique to the Jacobshavn Fjord is termed an "outshoot" by Porsild (1919, p. 151), who attributes it to the pent-up melt water, and which like a spring freshet, breaks the ice dam to discharge hundreds of icebergs out into Disko Bay. Jacobshaven ice fjord also differs from many of the others in that it expels icebergs even in the winter despite the barrier of shore ice unless, of course, the latter is unusually heavy and of great width. But outshoots are most liable to take place in summer and in June at spring tide, while they are rarest, of course, in midwinter.



THE MOUTH OF JACOBSHAVN FJORD

FIGURE 54.—Looking eastward toward the fault in the foreland which marks the profile of Jacobshavn Fjord. In the foreground and being pushed outward in the direction of the water are hundreds of tightly packed icebergs. (See fig. 52 for a close-up view.) This sketch was made as the *Marion* approached Jacobshavn Fjord on August 8, 1928.

The icebergs discharged from the Jacobshavn Fjord appear to be retarded at its entrance. Hammer (1893, p. 27) observing this fact believed that the temporary halt was due to the presence of a bank at the mouth of the fjord, which he likened to the stopper in a bottle. But the presence of a shoal there, however, has never been proved by soundings, for none of the depths recorded on the charts, viz, 200, 210, and 240 fathoms, are in the position of the supposed bank. The *Marion* expedition took several fathometer soundings among the outer icebergs that crowded the mouth of the fjord August 8, 1928, but the depths of 195 to 215 fathoms, considerably greater than 130 fathoms, the figure assumed by Hammer (1893), do not suggest any shoaling. The presence of a bank, nevertheless, is not entirely disproved, inasmuch as the *Marion* succeeded in penetrating only to the outer margin of the iceberg cluster, and therefore may not have got in far enough to be directly over the shallowest spot.

Jacobshaven Glacier, about 250 feet high and about 4 miles wide, is reputed to produce the largest and most picturesque bergs in the north, ranging from 130 to 325 feet in height. Helland (1876, p. 106) measured one 396 feet to its top, and Drygalski (1895, p. 381) recorded another 425 feet high, but the usual dimension of a large berg is given as 220 feet. Many of the bergs in Jacobshavn Fjord, according to reports, tower higher above the sea than does the glacier front from which they calved, all of which is considered proof of buoyancy as the cause of calving. The bottom of Jacobshavn valley and upper fjord has a relatively easy, gradual slope, two important features leading to the production of large icebergs. The glacier front during the period 1880 to 1892 suffered several oscillations, the final result of which has been a retreat of approximately 6 miles. The rate of movement is 45 to 65 feet per day; faster in the middle than on the sides, and swifter on the top surface than on the underside. Our estimate of the number of sizable bergs now emitted from the fjord mouth is 1,350 annually.

South Greenland Glaciers

The heads of nearly all the fjords in southwest Greenland are filled with ice tongues which, after the fast ice has left the fjord, calve growlers and small lumps throughout the summer. The two southwest fjords in which the *Marion* cruised, Arsuk and Godthaab, contained a few scattered growlers, and without doubt the other fjords between Cape Farewell and Disko Bay were similarly characterized. The fact that the southern fjords are so much narrower and shallower near their heads than those in Disko Bay and northward is one of the principal reasons that no sizable masses of ice are discharged.

The outstanding glacial feature of the southern section is the huge ice lobe which protrudes from the inland cap and fronts the sea for 11 miles in the vicinity of Frederickshaab. Despite its large proportions no icebergs are produced, because the end melts at the same rate as it moves forward, the two processes thus maintaining a more or less fixed state of equilibrium. With the description of Frederickshaab Glacier we complete the account of the tidewater glacier points from the northern extremity of Greenland down its west side to Cape Farewell. There are 150 to 175 sizeable glaciers in this distance, of which only about 15 per cent contribute icebergs to the North Atlantic

Rate of Productivity of Iceberg Glaciers

In the following list is an estimate of the number of sizable icebergs that are discharged annually from the so-called dangerous, tidewater glaciers. By the term "sizable" we mean bergs that if unhindered are sufficiently large to complete the long journey out of the Arctic into the western North Atlantic. The data are taken from all available sources mentioned in the text.

Region	Iceberg glaciers	Annual number of icebergs
Eastern North America:		
Ellesmere Land	American Geographical Society, and others.	150
Greenland:		
North Greenland	Humboldt	150
Cape Alexander to Cape York	Morris Jessup	300
	Bowdoin	
	Melville-Tracey-Farqhar	
	Moltke	
	Pitufig	
Cape York to Svarten Huk	Gade	1,500
	King Oscar	
	Nansen	
	Dietrichson	
	Steenstrup	
	Hayes	
	Giesecke	
	Upernivik	
Northeast Bay and Disko Bay	Umiamako	1,500
	Rink	
	Itividliarsuk	450
	Sermilik	150
	Great Karajak	1,200
	Little Karajak	
	Torsukatak	750
	Jacobshavn	1,350
Total		7,500

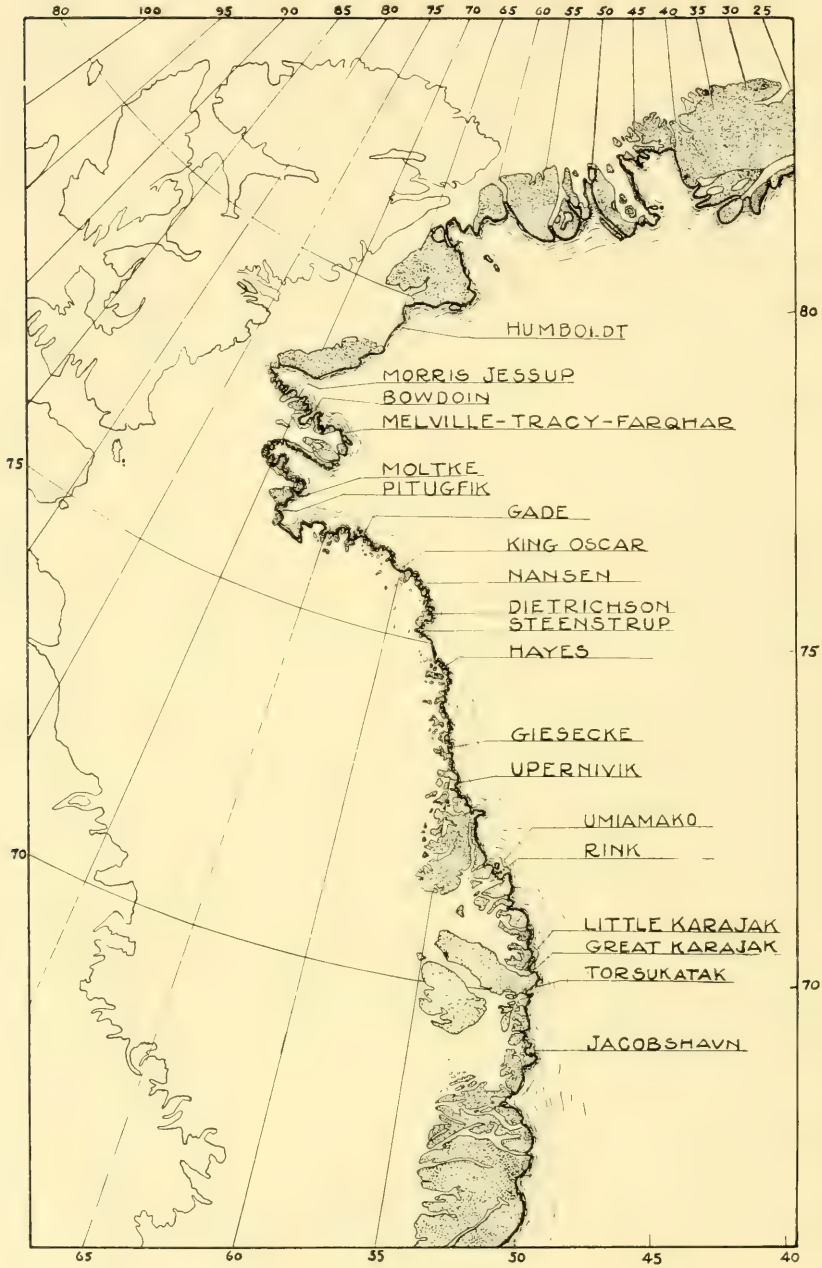
Due to the scantiness of data the foregoing estimate of 7,500 sizeable icebergs calved annually from west Greenland may be criticized as speculative. It is justified, however, even if it does no more than convey a general idea of the points of discharge and of the relative proportions in which their glaciers combine to make up the grand total. The figures of 1,350 bergs for Jacobshavn, 750 for Torsukatak, etc., are largely based on Helland's (1876, p. 108) measurement 5.8^3 kilometers and 2.3^3 kilometers as the total annual calvings of Jacobshavn and Torsukatak fjords, respectively. It has further been assumed that the average of sizeable bergs has a volume of 50,000,000 cubic feet (0.00145^3 kilometers) and that such bergs constitute one-third of the total calved ice, the other two-thirds consisting of smaller pieces and detritus. In other words, if Jacobshavn calves 5.8^3 kilometers a year, then 1.93^3 kilometers are in the form of sizable bergs, which, in 50,000,000 cubic-foot sizes, totals 1,350.[55]

The number of bergs from the respective fjords and glaciers is also subject to a slow change within certain limits over a series of years. For example, the retreat of Jacobshavn Glacier, first observed in 1850, is being reflected in the rate of its berg productivity.[56]

According to the foregoing table nearly 70 per cent of the North Atlantic bergs come from Northeast Bay and Disko Bay; about 20 per cent from the Melville Bay district; and only 2 per cent from the

[55] Drygalski (1895, p. 402), after careful measurements, found Great Karajak Glacier calved 13.5^3 kilometers annually and he criticizes Helland's estimate of 5.8^3 kilometers as far too low for Jacobshavn. If Drygalski is correct, then the output of Jacobshavn according to the relative weights given, page 95, would be 25.5^3 kilometers. But this seems too great, and does not agree with de Quervain's estimate of 10^3 to 100^3 kilometers for the entire wastage in solid form from the western side of the ice sheet. Surely Jacobshavn can not account for approximately one-fourth the total yearly output of west Greenland.

[56] Lauge Koch told me that he had not visited Disko Bay for 15 years, although he had passed it several times, but that he never saw so few bergs as in 1928. Umiamako Glacier in Umanak Fjord, on the other hand, appeared to have increased in activity, for the fjord was simply littered with bergs in 1928, very similar to conditions prevailing in Disko Bay several years ago.



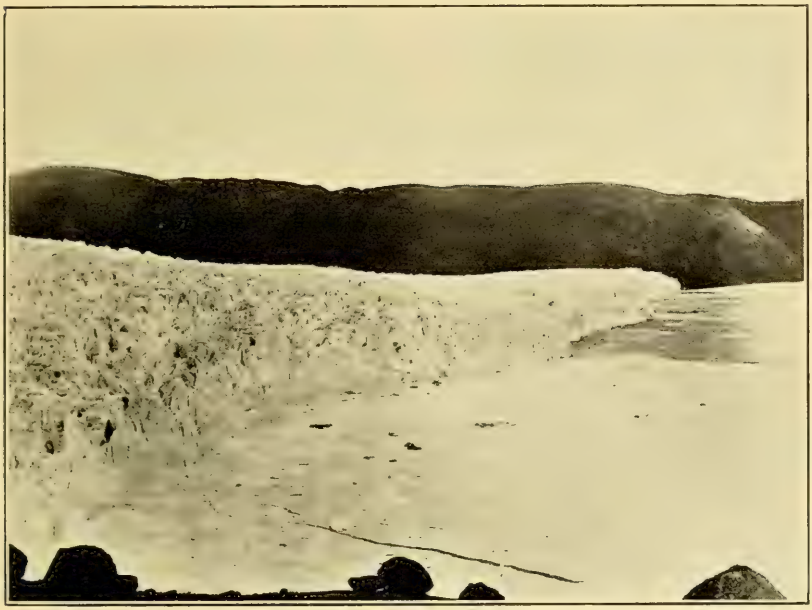
THE PRINCIPAL ICEBERG GLACIERS WHICH DISCHARGE INTO BAFFIN BAY

FIGURE 55.—The 20 glaciers named and located on the map above mark the origin of practically all of the icebergs that drift to lower latitudes in the western North Atlantic.

American shores. If 7,500 bergs be representative then normally only 1 berg in 20 finally succeeds in drifting out of Davis Strait and south of Newfoundland. The *Marion* expedition's estimate of 2,200 bergs south of Disko Bay in the summer of 1928 in conjunction with the number of bergs south of Newfoundland during the preceding spring indicates that the crop is not entirely destroyed during one summer, but that one, two, and possible three year age groups persist in Baffin Bay and Davis Strait.

Manner in Which Icebergs Calve

The so-called calving process by which end portions of a glacier break away from their main mass creating icebergs is a phenomenon

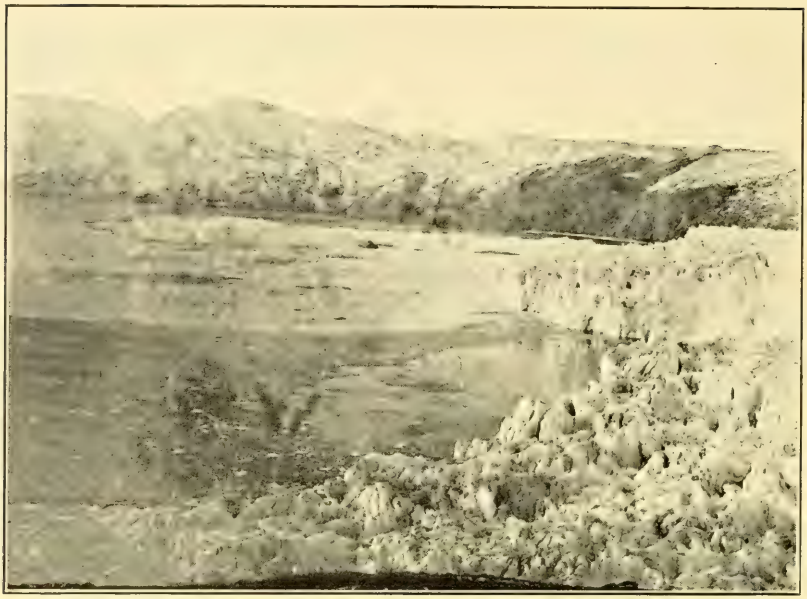


GREAT KARAJAK GLACIER PRIOR TO A MAJOR CALVING

FIGURE 56.—The front wall of Great Karajak Glacier in Umanak Fjord, west Greenland (lat. 70° 25′ N., long. 50° 30′ W.) before a major calving, August 13, 1892. Great Karajak is one of the most productive iceberg glaciers in the north. (Photograph by E. von Drygalski.)

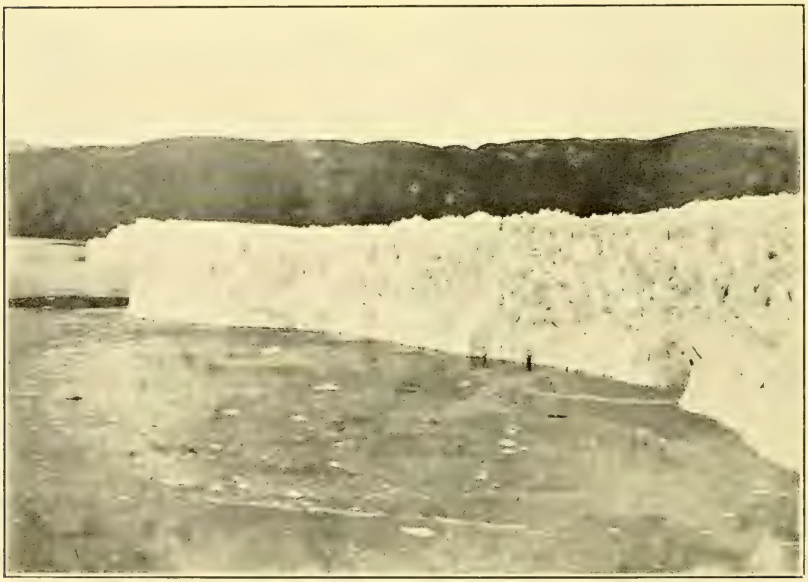
which has been under discussion for many years. The original conception held that the glacier end protruding out across the threshold of its molded bed lost support and through the continual accumulation of weight finally exceeded its structural strength and broke off. Such a process assumes, of course, that the bed on which the glacier slides has a steeper slope than the ice stream, which is not invariably the case.

A later theory introduced by Rink (1889, p. 276), maintains that icebergs are formed when the downward, slanted axis of the glacier prolonged into the greater depths of the fjord is more and more buoyed up until finally the end fractures, the iceberg rising to the surface. Rink qualified the foregoing by admitting that small pieces of ice are also calved by splitting away and falling down



GREAT KARAJAK GLACIER JUST AFTER A MAJOR CALVING

FIGURE 57.—The front wall of Great Karajak Glacier immediately after a great calving that took place on August 13, 1892. The major calvings, when several hundred yards of the glacier end breaks away, occur on the average of twice a month, during summer, near or at the dates of spring tides. The icebergs floating in the fjord were calved from the right and left of the glacier's front, as can be seen by comparing this view with that taken sometime previously. (Photograph by E. von Drygalski.)



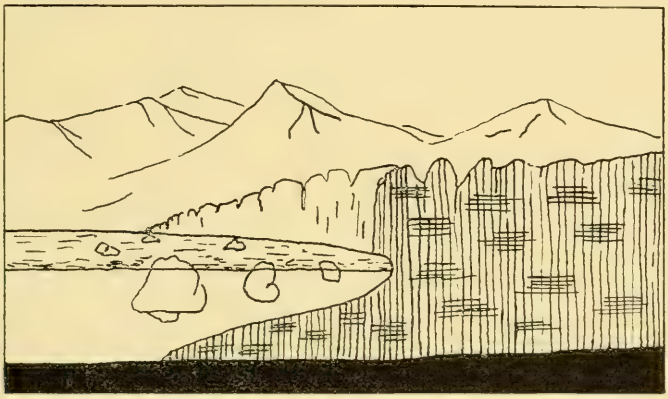
GREAT KARAJAK GLACIER EIGHT DAYS AFTER A MAJOR CALVING

FIGURE 58.—The front wall of Great Karajak Glacier eight days after the major calving of August 13, 1892. A comparison of Figure 58 with Figure 57 strikingly portrays the rapid rate which the end of Great Karajak Glacier in eight days has pushed outward into the fjord. The west Greenland glaciers, with a recorded rate of 65 feet per day, are probably the fastest moving ice streams in the world. (Photograph by E. von Drygalski.)

from the glacier front. Helland (1876, p. 99) actually witnessed the buoyancy process of calving—

> an immense dentate piece of the glacier turning over and over, and rearing on its edge high up in the air in front of the glacier, then the moment it rose, large towerlike parts of the latter fell down. On the following day one of the newly formed icebergs was measured and found to be 89 meters in height, while the height of the glacier wall hardly exceeded 49 meters.

These observations were made on Jacobshavn Glacier, which it will be recalled is one of the few glaciers whose front is lower than many of its icebergs. Drygalski (1895, p. 402), in agreement with the buoyancy theory divides the calving into three classes. The first class includes large detachments of the front, from top to bottom, unassisted by previous fragmentation, and described as follows: The front face of the glacier slowly rises above the main portion, sways back in the direction of the parent body, the foot of the berg shooting forward and then settling down in the fjord, rolling back



THE MANNER IN WHICH GLACIERS CALVE

FIGURE 59.—The glacier in profile is usually grooved at the fjord water line, causing an overhanging upper and a jutting lower underportion. The ice of the upper ledge detaches by its own weight, while the underwater shelf is continually broken away by its own buoyancy. (Drawing after Russell in Kayser, 1928.)

and forth several times. Calvings of the second class refer to the breaking away of a submarine toelike underbody which protrudes into the fjord. Drygalski states that he heard a tremendous roar and suddenly saw an iceberg breach out of the water beneath and directly in front of the glacier wall. The underwater projection of the glacier front is presumably formed by the many disintegrating processes which proceed during summer at a faster rate in the air above, than in the water below. Calvings of the third class are simply pieces of ice breaking off and falling down from the upper body of the glacier. The third class of calvings produces only small bergs; the second class medium to large bergs; while the first class calvings account for the largest run of bergs.

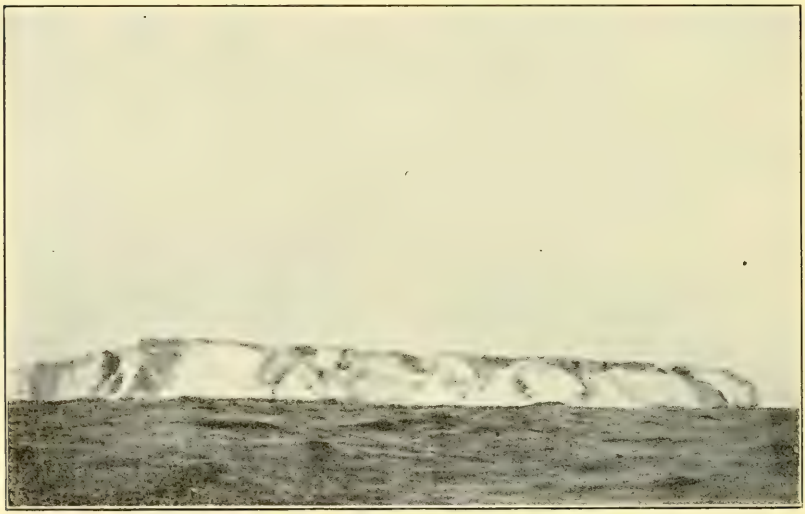
Steenstrup (1883, p. 92) warns that too much emphasis should not be placed on calving caused by the submergence of the outer end of the glacier, calling attention to the sensitive balance that he has found out on the bottom of the fjord. The outer end of the ice

tongue, gradually freed of its upper part, finally acquires the reserve buoyancy necessary to initiate calving.

It seems most likely that the two principal factors responsible for calving can be defined as disintegrative processes due to (1) the static position of the glacier end, and to (2) the dynamic form of the glacier. Following the general rule for ice and water in contact, melting proceeds fastest at the water line, the profile of the glacier fronts tending to be deeply grooved at the surface of the fjord, with an overhanging upper and a jutting underportion. The upper ledge is broken back solely by its own weight, while the underwater ledge is continually broken away by its own buoyancy. Superimposed on this type of wastage and of first importance in the case of the very active glaciers is the continual fraction and disruption due to the forward movement of the glacier.

Form and Size of Bergs

Icebergs in the northwestern North Atlantic vary greatly in form; they may be pinnacled, domed, roofed, or ledged; but spired, min-



THE BLOCKY PRECIPITOUS-SIDED CLASS OF ICEBERGS

FIGURE 60.—Second only to the rounded and pinnacled class of bergs is the blocky, steep-sided type which often reach the waters south of Newfoundland in the same plane of equilibrium that they left the glacier. This class of bergs is calved mostly from points where the topography permits the ice cap to directly face the sea. An example of such a locality is Nansen Glacier in Melville Bay. (Official photograph, international ice patrol.)

areted bergs, or bergs clustered with icicles are rarely seen outside of the school geographies. Their irregular shapes are not surprising when one considers how calving from the glacier takes place, and the many subsequent processes of disintegration that are constantly wasting the berg away. Despite the many agencies which tend to destroy uniformity, there are however, in any unassorted run of bergs several which bear a common resemblance.

In this respect the irregular, dome-shaped bergs which constitute the largest group are strikingly distinguished from the flat-topped

precipitous-sided type. The glaciers of Disko Bay and Northeast Bay discharge most of the picturesque forms, while northward in Melville Bay, where the ice sheet itself is thrust directly out into the sea, many of the square, blocky icebergs are produced. The marked difference in form, therefore, may be largely attributed to the type of the coastal land underlying the margin of the inland ice. Plateau-like bergs may maintain their equilibrium for a year or longer, for such have been sighted even off Newfoundland with the original dust and wind-blown sand of Greenland coloring their tops.

But the triangular, dome-shaped berg and the straight-sided type enter the Atlantic as bulky, massive bodies, worked upon henceforth by the various agencies of destruction. Bergs in the far north, with few exceptions, maintain their original outlines but after in-

THE DOME-SHAPED CLASS OF ICEBERGS

FIGURE 61.—A Davis Strait berg observed by the *Marion* expedition the summer of 1928. Its domed and rounded shape is characteristic of the largest class of bergs to be found in the northwestern North Atlantic. Its proportions of air-exposed body to be submerged are 1 : 5 to 1 : 4. (Official photograph, *Marion* expedition.)

vading the warm oceanic waters of the Atlantic they melt so rapidly that they change appearance often. The boxlike berg is first washed and melted around the water line, and this continual undercutting of the perpendicular sides results in a corresponding breaking away of the prominences above. After a time the unequal detachments cause the ice to lose equilibrium and the roof tilts, forming a V-shaped berg. The absence of the square, upright types after mid season at the Atlantic terminus is noteworthy.

The well-known girdling incision at the water line is a common characteristic from which few bergs are free. Overhanging ledges strained beyond the structural strength of the ice detach, causing the berg to roll, so that it floats in a new plane, leaving the old water line clearly visible for miles. We have seen bergs off the Grand Bank with as many as three or four old water lines marked

diagonally across them. The surface of the berg above water is often sculptured and chiseled in marked contrast to the part below surface which has been smoothed and rounded in contour by the effect of the sea. The continual shifting of stability in the irregu-

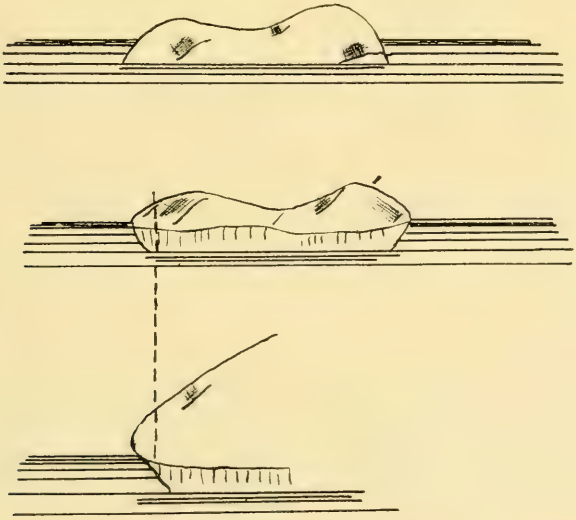
THE PINNACLED CLASS OF BERGS

FIGURE 62.—The fact that many icebergs in the north are pinnacled and irregular causes the question of mass flotation to become a subject of mostly academic interest, and as a result the attention becomes focused on the proportions of height above the sea to draft. The berg in the photograph has a draft equal to about twice its height above water. (Official photograph, international ice patrol.)

lar-shaped Greenland bergs often results in reefs jutting out a hundred feet or more underwater beyond the visible contour of the ice. These submerged ledges usually stand out beneath the waves as clearly as a coral reef in southern climes, yet they may prove ex-

THE DRY-DOCK OR VALLEY TYPE OF BERG

FIGURE 63.—An iceberg sighted by the international ice patrol about 25 miles southwest of the Grand Bank floating in water of 39° F. The "valley" or dry-dock berg is one of the common types sighted south of Newfoundland during the ice season. The continual surging of the waves and swell develop a central bore and later a deep, wide valley such as depicted here. Compare the water-washed polished surface of the underportion of the berg with the rugged air-exposed upper parts. (Official photograph, international ice patrol.)



A CHARACTERISTIC FORM OF MELTING

FIGURE 64.—All bergs drifting into the North Atlantic melt most rapidly at the water line, as shown above. This small berg was first sketched April 13, 1913, in the mixed water around the Grand Bank. The second sketch shows the effects of three weeks of melting.

ceedingly dangerous to a ship passing close by. The British steamer *Nessmore* was severely damaged by such a projection, and it is common report that a similar obstruction caused the tragic loss of the White Star Liner *Titanic*.

The bergs which penetrate far south in summer, or deeply into the Gulf Stream, often assume a completely water-washed, polished form, like a beach-worn stone. (See fig. 74, p. 114.) No sharp prominences are visible anywhere and such bergs are usually low, hence very difficult to detect. The most striking form is given to icebergs entering the North Atlantic by washing. The continual surging of the waves and swell back and forth finds the zone of greatest weakness developing a central bore and later a deep, wide valley. Such bergs are a familiar sight south of Newfoundland and are called by various names: "Valley," "dry-dock," "winged," "bicuspid," "saddle-back," and "double-horned." (See fig. 63, p. 103.)

Steamers often report two bergs close together when correctly speaking, it is simply one old eroded iceberg whose valley bottom is submerged beneath the sea. Blowholes and reentrant caves are not uncommon features through which the swell sometimes spouts hundreds of feet in the air. The valley berg, fragile in its last stages, consists of thin walls, that eventually break off, and in periods of smooth weather beautiful, curved, and slender "swan's necks" are carved by the sea. High arches are on rare occasions sighted. In most instances they represent the washing of the waves that started with a small cave and wore it larger and larger, but in some cases they may be relics of caves formed in glacier fronts by the escape of glacial thaw water.

The icebergs are largest in the vicinity of the fjords soon after their birth and diminish in size the farther they are carried into the Atlantic. The highest berg of a group of 87 measured by Drygalski (1895, p. 401), in Northeast Bay was 447 feet. Helland (1876, p. 106), measured bergs to a height of 292 feet and Steenstrup (1883, p. 96), 249 feet. Again Helland measured 8 bergs in the mouth of Jacobshavn fjord, 142, 145, 146, 181, 198, 201, 343, and 356 feet, the largest of which contained 718,000,000 cubic feet (17,000,000 tons), with 230 feet as a common height for a large iceberg. Drygalski measured a group of 70 bergs assorted as follows: Four were over 325 feet high, 14 were between 230 and 325 feet, 25 were between 160 and 230 feet, and 27 were less than 160 feet. The figures for bergs in the vicinity of the large glaciers are as follows:

Great Karajak: 250, 253, 207, 211, 250, 244 feet. Height of glacier front, 160 to 330 feet. The average height of bergs from Great Karajak is 180 to 250 feet.

Itividliarsuk: 194, 192, 214, 224, 230, 230, 95, 125 feet. Height of glacier front, 227 to 260 feet.

Jacobshavn: 69, 105, 135, 214, 330, 336, 447 feet. Height of glacier front, 277 feet.

The foregoing shows that Jacobshavn calves the highest icebergs in the north with Great Karajak following a close second. The icebergs of east Greenland are apparently not so high as those of the west coast, as Amdrup (1900, p. 243), estimates the largest of these as 160 to 215 feet in height and about 3,280 feet in length. The loftiest berg observed by the international ice patrol south of New-

THE TILTED, FLAT-TOPPED BERG

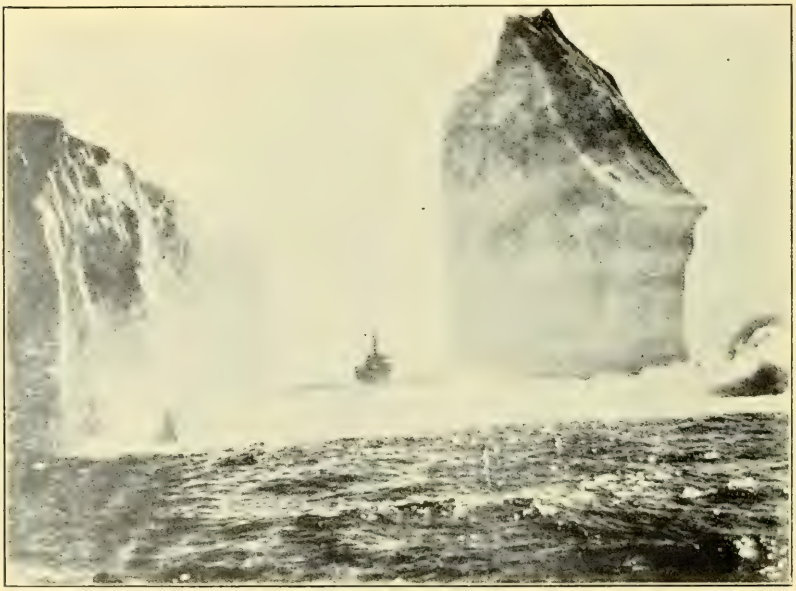
FIGURE 65.—Flat-topped bergs which roll into a new plane of flotation form V-shaped bergs, a common sight south of Newfoundland. On the right-hand side of this berg, tilted at an angle of 45°, can be seen the original top surface when the berg was a part of the Greenland ice cap. (Photograph by Lieut. Commander N. G. Ricketts.)

A COMMON ICEBERG FORM SOUTH OF NEWFOUNDLAND

FIGURE 66.—After floating some time without changing equilibrium, a large portion of ice was calved from the left-hand side of this berg. This caused the right-hand side to sink and the left to rise. The old water line is plainly discernible, inclined at an angle of 25° to the water. (Official photograph, international ice patrol.)

ICEBERG WITH AN ARCH

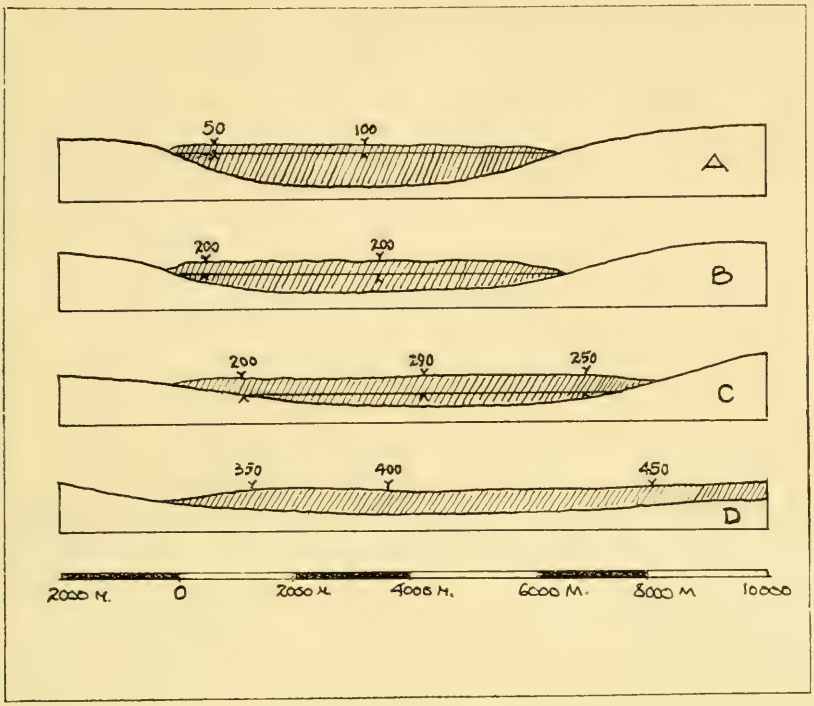
FIGURE 67.—Occasionally icebergs are observed in the western North Atlantic pierced by a large hole or bore through which the waves continually surge and spout. Such holes or bores develop from the gradual enlargement of a slight depression or embayment. Caves are developed by waves in a fashion similar to those along the sea cliffs on shore. (Photograph by Lieut. Commander N. G. Ricketts.)



THE "TWIN" BERG

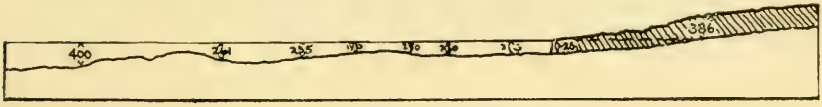
FIGURE 68.—A modification of the ordinary "dry dock" or "valley" shaped berg is the type shown here. Such ice is sometimes reported by vessels at sea as two bergs or twin bergs. It, however, is simply one berg that has been disintegrated by wave action until the mid section has retreated below the surface of the sea. The U. S. Coast Guard cutter *Tampa* can be seen through the gap, standing by the berg, warning by radio all approaching steamships. (Official photograph, international ice patrol.)

foundland was 262 feet, while the longest berg measured 1,696 feet.[57] This last was of the tabular type, with an almost level top, 65 feet



GREAT KARAJAK GLACIER

FIGURE 69.—Four cross sections of Great Karajak Glacier taken at various distances from its outer end. A. The front wall 100 meters above sea level in its middle. B, Section taken at the point of its steepest descent. C, At the Taisuak Step. (Figure from Drygalski, 1897.)



LITTLE KARAJAK GLACIER

FIGURE 70.—A longitudinal section through Little Karajak Glacier and out onto the bed of the fjord. Scale 1 : 100,000; the depths in meters. (Figure from Drygalski, 1897.)

above water, and was estimated to contain 900,000,000 cubic feet (22,800,000 tons) of ice. The average berg in the Davis Strait

[57] One of the largest icebergs ever recorded in the western North Atlantic was sighted August 28, 1928, by the steamer *Idefjord* as an "ice island." A verified report was printed in the Hydrographic Bulletin 2055 of February 23, 1929, by the U. S. Hydrographic Office as follows: "At 8.30 a. m. (ship time) the northwestern point of the ice island was abeam to port, distant 9 miles by 4-point bearing; the ship ran 4 miles before the northwestern point was abeam. When the ship was abreast the middle of the island, vertical angles were taken which showed a height of 80 feet, but there were peaks which were about 20 feet higher. After passing the island a very good view was obtained and the ice extended as far as could be seen. There was a light southwest breeze and the visibility was very good, but it was hazy over the ice." Several other steamers reported passing this huge mass of ice during the succeeding few weeks. The U. S. Coast Guard patrol boat *Marion* was ordered to investigate this report and searched the vicinity two weeks after the ice island was originally reported. Several bergs were found; the largest one, 775 feet wide and 825 feet long, however, did not even approach the dimensions of the "ice island." If the report be authentic, this huge iceberg exceeds in size anything previously encountered in the North Atlantic. Krümmel (1907, p. 520) relates a report of an ice island sighted near Cumberland Gulf, Baffin Land, in 1882. It measured 50 to 65 feet high, 7 miles long, and 3½ miles broad.

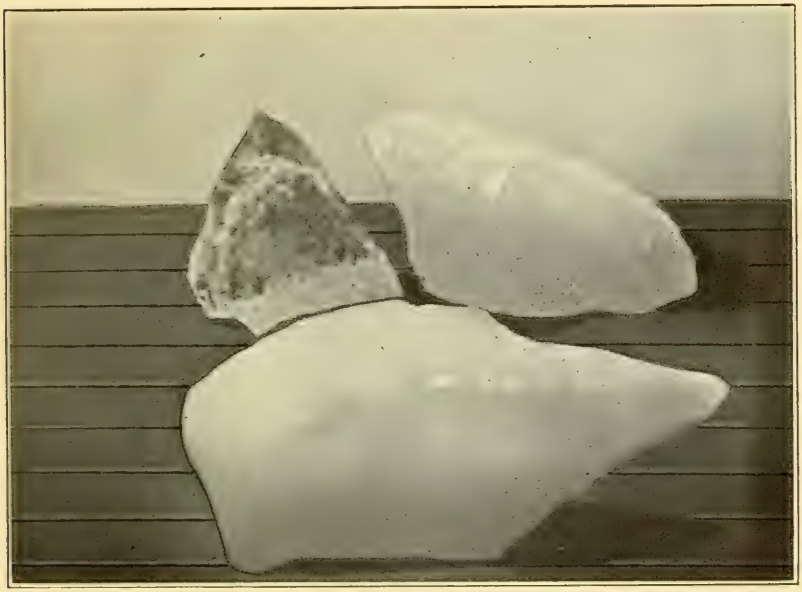
region is estimated to have a volume of 50,000,000 cubic feet, while the average for the Grand Bank is about 6,000,000 to 8,000,000 cubic feet. The average being around Newfoundland is about 100 feet high. The average berg in Disko Bay was estimated by the *Marion* expedition to be twice as high as the average Grand Bank's berg, but it is very difficult to judge accurately, because the height of the land surrounding Disko Bay dwarfs them. The general form of the Disko bergs appears bulkier, and there are few of the deep-valleyed forms that are so common around Newfoundland. The usual melting signs which also appear on the underwater surface of the bergs in lower latitudes are less apparent in Disko Bay.

Structure, Color, and Density of Icebergs

It is well known that the ice composing icebergs is formed from snow crystals which have gradually sublimated into large granular ice. This process of ice formation entangles and imprisons a vast quantity of air in the mass producing a structure which is quite different from ordinary pure ice. Barnes (1906) shows that the material is relatively hard below temperatures of 15° to 20° F., but if it be warmed, its power of resistance rapidly decreases, giving away almost entirely at the melting point, where it is soft indeed. An iceberg which invades the Atlantic and comes under summer temperatures, becomes pudgy and soft on its outside. A projectile plunges into the berg with a "chug" but in cold weather it only shakes down a shower of dustlike crystals. Steenstrup (1883) observed that by reducing the temperature of berg samples the air bubbles inclosed in the ice were subjected to a strong compression, and in that condition pricking by a needle was sufficient to burst the ice with the crack of an explosion. Barnes (1927a, p. 93), however, from his observations upon the size of the air bubbles in the ice, and after they have been released, claims that under natural conditions the air imprisoned in iceberg ice is under no greater than atmospheric pressure. It is also interesting to note that Barnes found that the air imprisoned in iceberg ice probably thousands of years ago has the same composition as the atmosphere of the present day. He has published the results of some recent experiments (Barnes 1928, p. 345) to determine the amount of air contained in berg ice, and although the work is preliminary, it was found that 7 to 15 per cent with an average of 10 per cent of the co-volume was air.[58] Small fragments of iceberg ice floating in sea water have been observed to effervesce with a sputtering sound plainly audible a foot distant quite similar to the frying of bacon. Such a reaction would probably indicate furthermore the release of innumerable minute air bubbles under greater than atmospheric pressure. Lieutenant Commander Ricketts informs me that he has observed large bubbles of air rising to the sea surface on calm days near the sides of icebergs. Undoubtedly the melting of the ice under water causes many of the small bubbles to collect and arrive at the sea surface as large air bubbles. The size of the bubbles permeating the ice may vary considerably from a diameter of one-fourth inch or more to a minuteness invisible to the naked eye. The average is less than the size of the head of a common pin.

[58] Thuras (1914, p. 68) found a content of 1 per cent, but, he adds, the ice used for the experiments was unusually clear and probably contained a minimum amount of air.

Berg ice melts slower than the artificial variety and the water obtained from it is fresh and pure. The color of the bergs is a peculiar opaque flat white, often mistaken by the inexperienced for snow, often with soft irridescent hues of green and blue. Many bergs are also tinged in places with a brown, yellowish stain, due probably to diatoms and other forms of planktonic life in which our northern seas are so richly abundant. The snowy white appearance is caused by surface weathering of a few inches to a foot or more in depth, and also to the effect of the sun's rays which release innumerable air bubbles; the ice underneath this surface film is stated to remain a clear deep, green color. If a sample be examined closely it will be seen to be covered with innumerable white threadlike lines



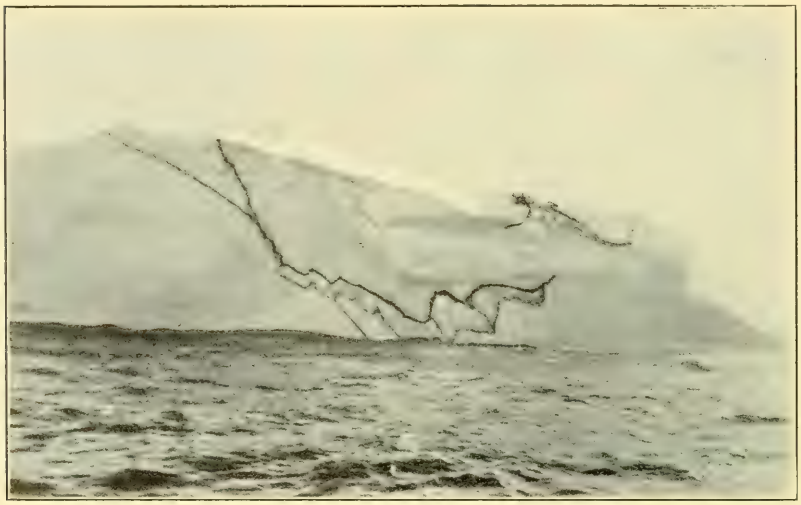
A CLOSE-UP VIEW OF ICEBERG ICE

FIGURE 71.—Three samples of iceberg ice. The two largest pieces of slaty opaque whiteness emphasize the physical character of iceberg ice, a mixture of sublimated snow crystals and air brought together under great pressure. The smallest piece was taken from a vein of clear ice often to be found in many icebergs. The veins are probably old cracks and fissures that have filled with thaw water either in the berg itself or when it was a part of the glacier. (Photograph after H. T. Barnes.)

etched all over the surface. A melting surface on close examination reveals the disarticulation of the individual glacier grains which average about three-sixteenths inch in diameter and which are separated one from another by depressions sometimes one-fourth to one-half inch across. This structure gives the ice a roughened, pebbly surface. Few icebergs are structurally homogeneous throughout, most of them being striped by one or more blue or green veins, or ribbons, of compact, transparent ice which stand out strikingly against the porous-white background. These ranging in width from an inch or more to several feet are formed by crevasses filling the water and then freezing as clear, blue ice while the berg was still a part of the margin of the ice cap. Other blue bands are also produced by the pressure and shearings that the glacier streams expe-

rience flowing outward through the foreland. The blue-green veins often give a very illusive effect as if the iceberg were cleft from top to water line, and, instead of ice, one were looking directly through the gap at the sky beyond. Another feature of interest is the rock and earth débris which has become imbedded in the mass. In this manner, doubtless, melting bergs drop considerable detritus along their paths of drift. The earth usually lies in streaks and veins striped across the sides and occasionally stones and bowlders the size of a man's body are seen imbedded in the top.

The ununiform distribution of air and the banding of the ice through melting and freezing are all processes which tend to vary the density from place to place in an ice sheet. Take one example: The bottom layers of a glacier being under greater pressure than the uppermost few meters, results in the former being much denser than the



EARTH VEINS IN A GREENLAND ICEBERG

FIGURE 72.—Many of the bergs drifting out of Davis Strait are marked with one or more veins of soil and débris which were taken up from the underlying bed when the ice was a part of the glacier in Greenland. Icebergs may thus transport such débris from the Arctic to the tropical regions of the Gulf Stream. (Photograph by Lieut. Commander N. G. Ricketts.)

latter. Ahlmann, the Scandinavian glaciologist, I am informed, has made several determinations regarding the density of glacial ice in the Norwegian mountains, from the upper surface of a glacier down to depths of several meters, and has found values which vary within wide limits. The density of the interior of a glacier, according to Ahlmann, differs little from the density of ice formed by fresh water freezing in the ordinary manner (0.9167), but surface samples of glaciers may be so porous as to give density values as low as 0.6. The density of an iceberg being no more or less than a detached portion of a glacier depends, therefore, on that part of the glacier from which it calved. A berg composed mostly of the lower layers of an ice sheet will be denser than one which broke away from the upper surfaces of the same sheet. On the other hand, a given berg in calving may embrace the entire glacier front, top to bottom, and in

such cases the specific gravity of the ice in one part of the body will differ from that in another. Such bergs will exhibit a marked stability, at least in their early stages, and the frequent number of flat bergs observed floating with the original Greenland, sand-covered tops, still uppermost, may be a corroboration of the above-described conditions.

The density of the ice in an iceberg may change after it has detached from the ice sheet. This development is due to surface weathering, a process which does not normally extend to a depth of more than a few feet. Sverdrup has suggested to me that, according to this view, an iceberg is probably composed of a core which is densest, surrounded by layers in which the density decreases towards the surface, depending upon the degree of porosity of the ice there.

Wright and Priestley (1922) found the density of a sample of ice taken from the Antarctic glacier to be 0.897. Barnes's (1928 p. 345) figure of approximately 10 per cent covolume of air taken from an iceberg stranded off Twillingate, Newfoundland, should give a density figure of about 0.825. Barnes, however, points out in a letter to me that the density value of berg ice varies by a wide margin, depending upon what part of the berg the sample was taken.

The Coast Guard cutter *Tampa* in March, 1930, brought to Boston a piece of ice taken from a berg found off Newfoundland. Determinations of the specific gravity were made by means of the displacement method in a bath of kerosene at the Jefferson Physical Laboratory, Harvard University. From two trials with samples taken from various parts of the piece the lowest density was 0.8977, the highest 0.9045, and the mean 0.8977. This figure agrees quite well with that of Wright and Priestley for Antarctic glacial ice, but it is greater than Ahlmanns's or Barnes's results. It is hoped that several more density determinations will be made from a number of samples to be collected by the ice patrol from various parts of a berg as it disintegrates. Until more such data is assembled we can not state definitely what is the figure representative of the mean density of an iceberg.

A common question met in connection with the subject of icebergs is: What proportions of a berg are above, and what proportions are below, the surface of the sea? If we assume that the mean specific gravity of iceberg ice is 0.8997, as compared with 0.9167 for pure fresh-water ice, we may arrive at the following estimate of mass buoyancy.[59] The submerged portion of an iceberg floating in sea water of 1.02690 density, a figure representative of the average density of the surface layers along the continental edge, is then about 0.8760 of the entire mass or, in other words, one-eighth of the mass of a berg is above and seven-eighths below the surface of the sea. These figures do not agree with Barnes (1927a, p. 98) who emphasizes the amount of air imprisoned in glacial ice, and states that because of this a berg will float with about one-third of its mass above water. It will require a greater number of density determinations than are now available to ascertain this figure with accuracy.

The fact that the icebergs typical of the Northern Hemisphere are as a rule irregular in shape causes the question of mass flotation to become a subject mostly of academic interest, and as a result atten-

[59] A cubic foot of iceberg ice weighs approximately 55 pounds.

tion becomes focused on the proportions of the height to which a berg towers above the sea, as compared with the depth to which it extends below the surface. Obviously, mass for mass, a pinnacled berg will rise to a greater height above the water than a more uniform or blocky shaped one. In the majority of cases of icebergs, therefore, calved from glaciers in the north, height tends to approach draft. Steenstrup, we find, gives 1:7.4 or 1:8.2 for the approximate proportions of a floating berg, height to draft; and a square blocky berg as 1:7 to 1:9 of its mass. The figures offered by Krummel (1907, p. 520) are 1:8 as one extreme and 1:4 as the other, with the common run equaling 1:5 to 1:6.

There are several records of actual measurements made on icebergs as follows: The Denmark expedition observed a berg off the east Greenland coast which measured 108 feet high and 270 feet draft, or proportions about 1 to 3. Dawson (1907) mentions measuring an iceberg August 7, 1894, stranded in the Strait of Belle Isle. It rose 105 feet above the sea in a depth of 342 feet of water, or proportions of 1 to 3. Rodman (1890, p. 7), describes a tall, spired berg 100 feet in altitude that was grounded in 16 fathoms of water in the Strait of Belle Isle; proportions almost 1 to 1. The international ice patrol (Smith, 1925, p. 81) measured a berg east of Newfoundland by sextant angle as 106 feet, and its draft by means of a wire drag was 200 feet; proportions 1 to 2. The following table of approximate proportions (height to draft) for bergs south of Newfoundland has been compiled with consideration for density values and also the results of actual measurements of bergs of various shapes.

Type:	Proportions (exposed to submerged)
Blocky, precipitous sides	1:5
Rounded	1:4
Picturesque, Greenland	1:3
Pinnacled and ridged	1:2
Last stages, horned and winged	1:1

It is seen that the Greenland bergs, characteristically pinnacled and domed, have a much smaller draft than has heretofore been the common impression. It is not surprising, therefore, that sizeable bergs can drift in relatively shallow waters, as over the Grand Bank or across the sill of Davis Strait.

Disintegration and Melting of Icebergs

As soon as an iceberg cleaves from the glacier front and is water borne, the processes of wastage commence. This phenomenon proceeds in the following three ways: (1) Melting, (2) erosion,[60] and (3) calving.

It is seldom, under actual conditions met at sea, that all three methods of disintegration are not at work, but one type is often more active than the other two until a change of the sea or a change in the shape of the ice causes a rearrangement of the three agencies.

[60] The word "erosion" is used here to designate the accelerated rate of melting that takes place when icebergs are washed by waves, by swell, or by rain. Any motion which carries away the water which is in contact with the ice and brings a fresh supply of warmer water to the ice surface materially hastens melting. Erosion, then, as used, refers not only to the frictional action of the water but it also includes the far greater melting effect.

The processes of disintegration are best observed at sea near the southern melting end of the berg's journey, and for this reason the ocean south of Newfoundland provides an excellent region to carry out such studies.

THE RUGGED SURFACE OF RECENT CALVINGS

FIGURE 73.—A tower of ice 150 feet high which the international ice patrol observed grounded on the Tail of the Grand Bank in 1922. When approached within a mile or less, a berg of this size creates a distinct feeling of impressiveness. The photograph also strikingly reveals the planes of cleavage from which tons of ice have calved. The proportions of exposed to submerged body of pinnacled berg, the group in which this one is classed, is 1 : 1. (Official photograph, international ice patrol.)

Melting processes are always at work on all bergs, but they are slowest in winter and when the berg is far north or on the continental shelf. Wastage speeds up during the warmer months and as the ice drifts farther and farther south into the open Atlantic. If the water is warmer than the air, a condition sometimes met in early

spring when icebergs drift into the Gulf stream southeast of Newfoundland, then the underwater body of a berg melts faster than the above part. If, on the other hand, the air is warmer than the water—for example, during summer in the cold waters around the Grand Bank—the air-exposed portion of an iceberg will melt the faster. But in each case account must be taken of the specific heat of the water as compared with that of air. Melting usually proceeds fastest at the water line of icebergs, the engirdling furrow being a common sight on bergs in the ice regions. Acceleration of the rate of wastage is also affected by the size of the ice body. The smaller a berg becomes the greater is the ratio between exposed surface and mass, and therefore the faster is the rate of melting at a constant temperature. Also the smaller a berg becomes the more

A WATER-WASHED BERG MELTING RAPIDLY

FIGURE 74.—This berg on May 9, 1921, in latitude 41° 45′ N., longitude 47° 30′ W., had been two days in the Gulf Stream south of Newfoundland; water temperature 63° F. Its surface is completely water washed, due to the frequent changes in equilibrium, and it has acquired a polished surface, looking much like melted glass. On close inspection the berg could be seen "steaming," like an ice cake on a hot summer's day. A flow of water also, due to the excessive rate of melting, bathed the entire surface. (Official photograph, international ice patrol.)

it comes under the control of the surface layers, which by virtue of their greater warmth and more rapid motion accelerate the rate of wasting of the ice.

The most advanced stages of melting are observed on bergs which drift into the Gulf stream off the Grand Bank in summer, when during the warm, calm days of late June they become furrowed by hundreds of rivulets from every side. Under these conditions they melt so rapidly that the plane of equilibrium is constantly altering, so that the bergs roll over and over. The constant washing polishes the surface and smoothes off all sharp prominences, and the ice that reaches these subtropical surroundings fairly steams like an ice cake on a hot summer's day.

One more factor which may further the process of melting is the descent of warm air near icebergs. The atmosphere in the vicin-

ity of the Grand Bank during summer and northward on the Labrador shelf, as over other bodies of cold water, is usually much warmer at an altitude of a few hundred feet than it is close to the surface of the sea. An iceberg which rises a hundred feet or more will form an obstruction to the wind, tending to stir up the air to a still greater altitude. Around many bergs, therefore, one might expect the air to be much warmer than in the immediate neighborhood, and fact sustains theory, for a surprisingly warm blast of air has often been experienced by the ice-patrol ship when passing close by and just to the leeward of an iceberg. On the other hand, under different atmospheric conditions, cold air is experienced in the vicinity of the ice.

Erosion, washing by the waves and ocean swell is most effective of all destructive agents for North Atlantic bergs. The erosion processes are always at work from the time the ice drifts any appreciable distance out from the protection of harbors and bays. As the seas continually wash back and forth they find a crevasse, an irregular shape, or a slight depression, all of which they enlarge. The ceaseless surging of the swell, pouring back and forth through the small channels, soon erodes valleys, which in turn grow larger and larger, until they develop into the main features of the berg. The inner slopes retreat until only two thin side walls remain, and eventually the valley bottom deepens and disappears below the waves, thus giving the impression of two separate pinnacles or towers of ice. Finally in the later stages of wasting, far south in the Atlantic, calving and fracture outstrip erosion with the falling of the fragile side walls. Calving is then in turn superseded by melting agencies which turn the last page in our iceberg's history. It is interesting to note that the attacks of erosion are always directed at the middle of the mass of ice that is above water. If the sea sculptures too much away from one side, that region through loss of weight is lifted out of reach of the waves, which then concentrate on the other side, the midsection always receiving the brunt of the attack.

During early season around the Grand Bank, February to March, icebergs are subjected to a severe washing by waves which sometimes dash completely over them, causing a great deal of wastage. The long ocean swell which makes up during a gale also sets up stresses and strains in the above-water portion, causing much rending, cracking, and fracture. But only in the small-sized bergs, even in such gales, can one detect a perceptible roll and sway. The ordinary run of bergs remain as immovable as the Rock of Gibraltar. Calving begins when the melting and erosive processes have set up strains that exceed the structural strength of the ice. Prominences and overhanging ledges calve away, sliding down from the steepest parts of the berg's sides and slopes. Unequal detachments around the edges interfere with equilibrium and occasionally initiates calving on a major scale. The berg begins to roll slowly and deeply to and fro, and when some bulging prominence swings far away from the perpendicular thousands of tons of ice rupture to fall, avalanche-like, down to the sea. In the case of many tons of ice, the effect is very curious; it seems to fall much more slowly than really is the case. Stability is, of course, seriously disturbed, and the berg may again suffer one or more successive calvings of its irregular parts

until equilibrium is reestablished. Then the cycle of destruction is reenacted.

Bergs which calve frequently, producing numerous growlers, naturally expose a greater and greater ice surface, and therefore waste faster than those bergs which resist fracture because of their particular shape, or of calmer seas, or of other causes. Bergs which are completely water washed present fewer overhanging prominences than others, and are, therefore, less liable to calve. Such shaped bergs exposed to the same conditions survive longer than any other kind.

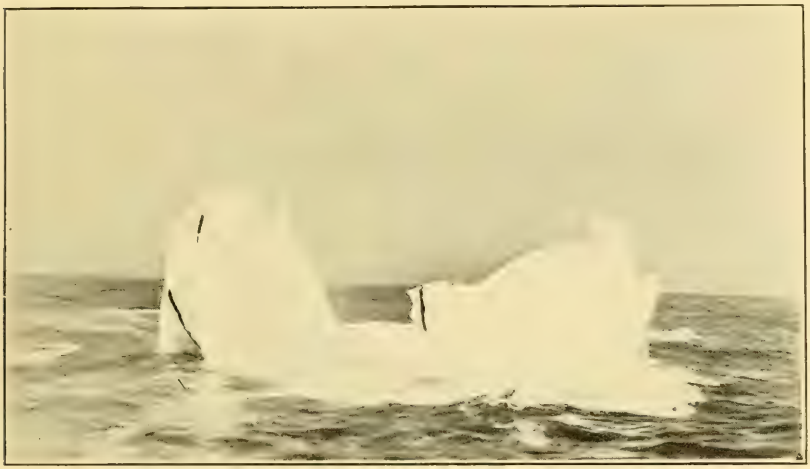
The natural processes of disintegration that attack a Greenland berg when it reaches the warm ocean south of the tail of the Grand

AN ICEBERG BUFFETED BY WINTER GALES

FIGURE 75.—Bergs drifting south of Newfoundland in early season (February to March) leave protective coast lines far behind. Then they are subjected to a severe buffeting by gales and ocean swell which sometimes shoot the spray hundreds of feet in the air. Only the smaller-sized bergs, however, display any perceptible motion; ice bodies as large as the one shown here remain as immovable as a rock. This berg measured 150 feet above the water and 400 feet in length. (Official photograph, international ice patrol.)

Bank are interesting to follow. Thus when one particular berg which on April 4, 1924, emerged from arctic surroundings of 32° F., to enter mixed water of 35° F., was then of medium size, consisting of a low ridge at one end separated by a shallow channel from a peak approximately 125 to 150 feet high at the other. The wind remained light for the next two days and the water warmed from 34° to 38°, while a slight swell washing the base of the berg materially assisted the melting processes. A growler was observed to calve occasionally, but disintegration was not rapid. By the next day the temperature of the sea had risen to 40° F., and a heavy swell making up about noon initiated rapid wastage. When the berg crossed the "cold wall," early the morning of the 8th, and floated in

warm water of 56° F., it began to dwindle fast, calving and rolling continuously. The swell subsided somewhat in the afternoon, but the tropical water combined with a moderate buffeting caused calving to continue. The most rapid disintegration took place on April 10; loud crackings were then continually heard, and the retreat of each wave exposed a surface appearing blistered, with the next wave crashing away the loosened covering, to be followed by another draining and more blistering. During this stage the wind was light southerly, a moderate southerly swell was running, and passing rain squalls were experienced, typical of Gulf Stream weather. It was apparent by nightfall that the effects of warm air, the warm waters, the rain showers, and the constant pounding of the seas would shortly complete the destruction of the berg, now rather small. It was not surprising, therefore, next morning that we found no traces of it.



RAPID WASTAGE OF ICEBERGS IN OCEANIC SURROUNDINGS

FIGURE 76.—The rapid rate of disintegration of this berg can be read in the excessive turbulence of the surrounding "white water." The seas in surging through the mid section have left the earth vein remaining only in the two side walls. A berg as small as this one—40 feet by 150 feet—reacts to the motion, riding the swells like a ship. (Photograph by Lieut. Commander N. G. Ricketts.)

The rate of wastage of about 30 feet of height per day is one of the most rapid ones on record.

From the foregoing and from other similar observations it is clear that the rapidity with which a berg that has drifted out of Davis Strait and reached North Atlantic waters around Newfoundland disintegrates depends upon the temperature of the air and water and on the state of the sea. A berg which drifts southward along the east side of the Grand Bank, except late in season, is surrounded by water of a temperature lower than 35° F. (2° C.) until after it passes the Tail of the Bank. Such cold surface layers are primarily due to winter chilling further north, but they are partly due to the presence of pack ice which assists to keep the surrounding water frigid and thereby allow the bergs to penetrate farther south. Disintegration under such conditions is rather slow, but in the mixed waters and in the Gulf Stream south of the Grand Bank it is accelerated. At 36° melting it quite noticeable; at 50° the changes can be

observed from hour to hour. A berg of average size, 70 to 90 feet in height, in mixed waters south of the Grand Bank will survive as a menace to navigation for a period of 12 to 14 days during April,

A CLOSE-UP VIEW OF AN ICEBERG AT SEA

FIGURE 77.—A close view showing the three processes of wastage—melting, erosion, and calving—of an iceberg at sea. The warm ocean swell, as it ceaselessly pours across the ledge of ice in the foreground, is the most potent factor in the disintegration of icebergs off Newfoundland. (Photograph by Lieut. Commander N. G. Ricketts.)

May, and June, but will not survive longer than 10 to 12 days thereafter. This represents a reduction of height at the rate of 5 feet per day in April and 6 feet per day in May and June. An equal sized berg farther south within the confines of the Gulf Stream, 65° to

70° F., will survive approximately 7 days, with considerable variation, equivalent to a height reduction of about 10 feet per day.

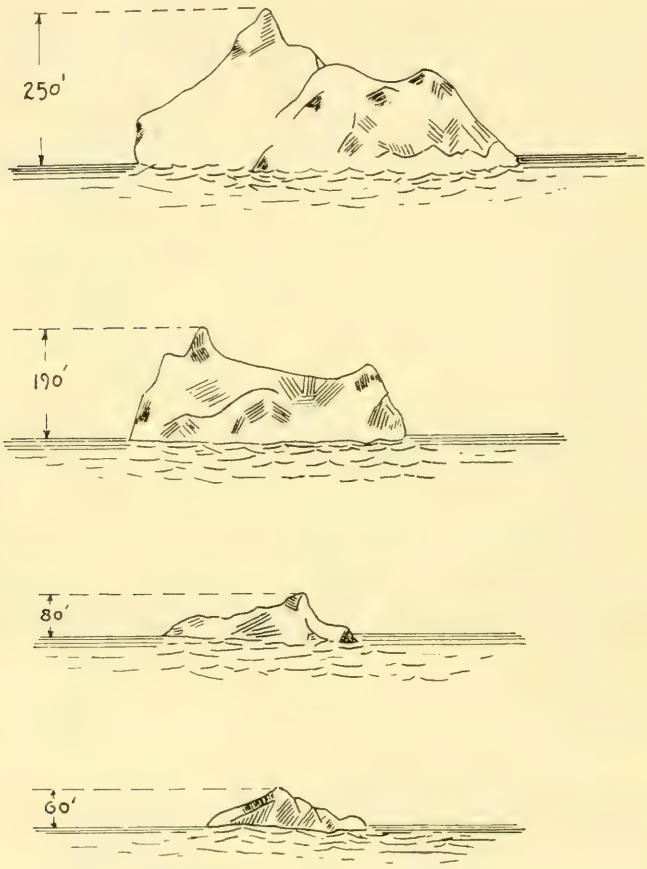
On April 11, 1921, we measured a large berg floating in water 34° F., between Flemish Cap and the Grand Bank, as 248 feet high. (Smith, 1922, Chart H.) When it was sighted again 10 days later in 33° F. water, it was 190 feet in height, and again 80 feet high on May 9 in water of 44°. When last seen on May 12 in water 63° F., it had shrunk to 60 feet. This indicates a rate of wastage in altitude of about 6 feet per day until the last few days, when the berg was surrounded by warm tropical waters, and then it lost height at the rate of 10 to 12 feet per day. Naturally, the bergs waste faster in the summer time south of the Grand Bank than in any other region of their occurrence; about six times faster than in the Arctic. In June, 1926, a large berg, 382 feet in length, floating in the northern edge of the Gulf Stream south of Newfoundland, completely melted

A MAJOR CALVING OF AN ICEBERG

FIGURE 78.—While approaching this berg, 1 mile distant, a part of the pinnacle on the left broke off. The detached piece weighing hundreds of tons dropped into the sea with a great roar. The white circle on the water around the berg is the broken ice, and growlers. (Official photograph, international ice patrol.)

away in 36 hours. On the other hand, during the same year and month there is the authentic report of a piece of ice sighted near Bermuda. (U. S. Hydrographic Office, Hydrographic Bulletin No. 1944, December, 1926.) Icebergs on the northern part of the Grand Bank sometimes survive as long as four to six weeks. And a large berg stranded south of Belle Isle in April, 1924, did not completely break up until August. These observations on the rate of wastage agree with those of earlier investigators. Thus Drygalski (1895, p. 420) measured some of the icebergs in Northeast Bay, Greenland, during the summer months and found that in a period of 7 days two of them decreased by 10 and by 13 feet, respectively, in height, while another in 6 weeks diminished from a height of 320 to 244 feet, a loss of 20 per cent in altitude. Johnston (1913, p. 23) estimates that bergs floating in water 50° F. or warmer suffer a 5 per cent reduction of mass per day.

Assuming that an average run of large bergs for the Arctic lose one-half their altitude before they reach the Grand Bank—i. e., 250 feet to 125 feet—at a daily rate of 6 feet,[61] than an uninterrupted journey of about five months is suggested. An average-size berg in the Arctic regions is estimated to contain 50,000,000 cubic feet (1,500,000 tons) of ice while an average berg south of Newfound-



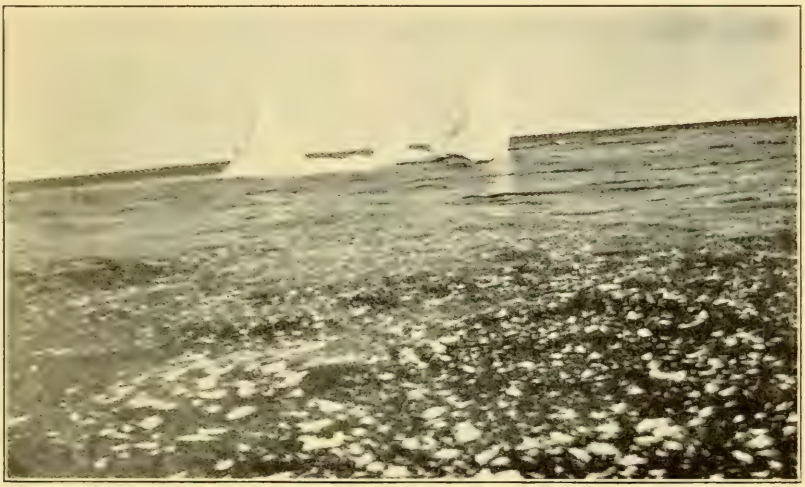
RATE OF MELTING OF AN ICEBERG

FIGURE 79.—This berg when first sighted on the east side of the Grand Bank, April 11, 1921, was 250 feet high. Ten days later it was 190 feet; 18 days later, 80 feet; and finally on May 12 it was only 60 feet above the sea. The average loss of height per day for icebergs around the Grand Bank is 6 feet; and in the Gulf Stream this accelerates to 10 feet per day.

land has a volume of approximately 6,000,000 cubic feet (150,000 tons), figures which upon comparison represent a reduction of mass from 8 to 1, Greenland to Newfoundland. These estimates of a mass reduction of seven-eighths agree very well with the height loss of one-half.

[61] A rough approximation of reduction of daily altitude is 2 feet, Baffin Bay; 4 feet, Davis Strait; 6 feet, Newfoundland and Grand Bank; and 10 feet, south of Grand Bank.

Mention should be made of published reports appearing from time to time regarding the sudden breaking up even of large bergs as a result of bursting, or of a resonant sound. Bowditch (1925, p. 264) states that bergs have sometimes been split by the single blow of an ax. The top of an iceberg has been described as often being "rotten" when it will topple down if a shot be fired at it, or even if a voice be raised in the neighborhood. Eskimos are said always to keep perfectly quiet when obliged to pass a berg at close quarters. The phenomenon is said to be due to an unequal contraction of different parts of the berg by which the imprisoned air is compressed beyond the structural strength of the ice so that it explodes. But we have not been able to produce any such effects, either by the ship's whistle or by the report of a 6-pounder gun fired close to a berg,



AN ICEBERG IN THE ADVANCED STAGES OF DISINTEGRATION

FIGURE 80.—June 15, 1922, in latitude 42° 43′ N., longitude 51° 27′ W., the water temperature was 48° F. The broken ice and small growlers can be seen floating away to leeward. (Official photograph, international ice patrol.)

though we have carried out many such experiments under various conditions.

Barnes (1927, p. 162), after carefully observing a large iceberg for a period of two weeks off Newfoundland, states that the greatest calving and cracking occurs in the early morning at and immediately after sunrise. Closer examination showed[62] that the water which had been melted on the exposed surface of the ice during the day froze again at night. The sun's rays of the following morning penetrating the ice before surface melting started, he believed to be the factor responsible by which the ice expands unequally. This does not happen later in the day because the heat from the sun is absorbed by the great quantities of melt water bathing the entire surface.

[62] Zeusler (1925, p. 38) states that the greatest amount of disintegration was noticed at noon and shortly after sunset. Ricketts (1930, p. 113) points out that the air-exposed surface of the bergs does not always freeze at night since he (Ricketts) observed water pouring from all visible surfaces of a berg on a cloudy night southeast of the Grand Bank.

This theory was first suggested by Barnes as a result of his observations on ice formations on rivers. Anchor ice, for example, in the St. Lawrence River is known to rise to the surface at the first early rays of dawn. If such relatively feeble light waves are capable of loosening great masses of anchor ice from its moorings, they might easily play an important part in breaking up icebergs.

Barnes (1927, p. 162) impressed with the belief that a small amount of penetrating energy from the sun sets up disruptive strains

THE LAST STAGES OF AN ICEBERG NEAR THE STEAMSHIP LANES

FIGURE 81.—This berg, on June 22, 1922, in latitude 41° 32′ N., longitude 49° 45′ W., south of the Grand Bank in the northern edge of the Gulf Stream, broke up completely within a few hours' time. The disintegrated part, now in the form of growlers, entirely melted within 12 hours, and the menace was removed from the ice patrol broadcast. (Official photograph, international ice patrol.)

in icebergs, carried out a number of experiments to test the effects of large quantities of heat. During the summer of 1926 (see Barnes, 1927a) several icebergs were boarded along the east coast of Newfoundland near Twillingate, and a slow-burning explosive called thermit was buried to a depth of 3 feet and more in the ice. Parts of a large plateau forming a major portion of one of the icebergs disintegrated when a charge of 100 pounds of thermit was ignited. Loud cracks and roars continued throughout that night and by the

time the berg was revisited in the morning, 400 feet of the plateau had been lost leaving a great hole surrounding the spot where the mine had been placed. Two days later, the berg had faded to only a shadow of its former self. From this he concludes that the use of thermit, or other means of thermal expansion, also offers a practical method of controlling the iceberg menace to the North Atlantic. Barnes, however, does not discuss the difficulties connected with placing a sizable charge of thermit at a vulnerable point in a berg at sea, and those who have had considerable experience around icebergs in the North Atlantic regard this problem just as difficult as determining the proper reactive type of explosive. The mining of icebergs, as developed at present, is considered justifiable in the case

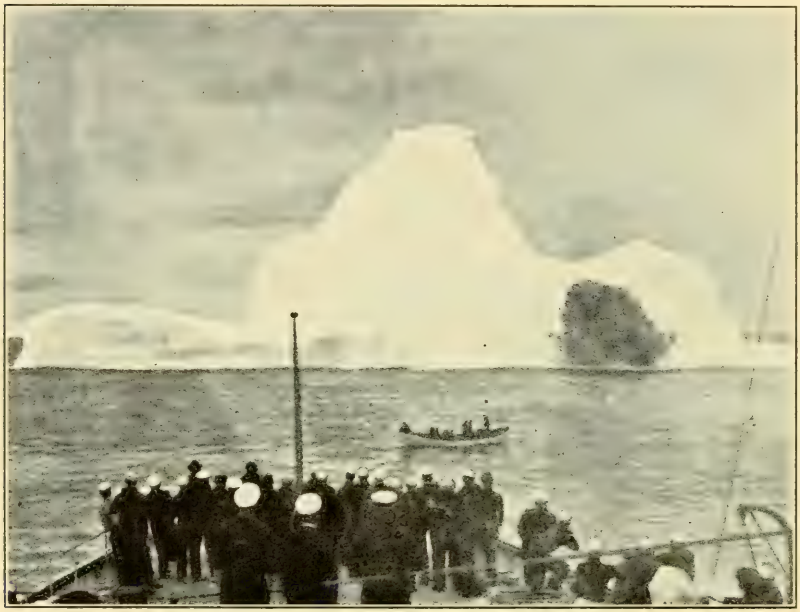
WRECKING OPERATIONS AND SMALL-BOAT DANGERS

FIGURE 82.—The career of bergs which drift across the trans-Atlantic steamship lanes and into the Gulf Stream south of the Grand Bank, such as this one did in May, 1923, can be shortened by the use of high explosives. The small boat from the U. S. Coast Guard cutter *Tampa* has just placed a 250-pound TNT mine below the water line of this berg. This is one of the most difficult and dangerous parts of the wrecking operations. (Official photograph, international ice patrol.)

of only a few bergs which lie across the main trans-Atlantic steamship tracks, and then only in periods of smooth sea.

Demolition experiments have been carried out by the international ice patrol, 1924 to 1926, on icebergs drifting far south into the Gulf Stream in order to hasten their removal as a menace to navigation. Owing to the swell and the smooth surface of the ice, such operations always entail danger, such as members of the mining crew slipping into the sea or falling onto ledges from precipitous heights, hence such work should never be attempted except at most favorable times. On May 20, 1923, a large iceberg floating in water warmer than 60° F. was mined with 250 pounds of T. N. T. on four different occasions, first on May 20; next on the 21st and finally on May 34, with the result that its life was shortened by only one or more days. The

first explosion occurred during a dense fog which shut off all view, but repeated heavy calvings were heard from time to time following the blast. Another mine suspended from one side to a depth of 10 feet brought down much loose ice, but a second charge placed at 20 feet below the surface of the water jarred the berg most of all. The last mine, being against the side wall of the berg at a depth of 30 feet, detached great quantities of ice, causing the berg to rise and then fall sideways and to break squarely in two at the point where the mine had been placed.[63] Gunfire has no material effect, but only shakes down a few tons of ice from the point of impact.



THE EFFECT OF EXPLOSIVES ON ICEBERGS

FIGURE 83.—On June 5, 1923, the U. S. Coast Guard cutter *Modoc* exploded four 50-pound gun cotton mines suspended from this berg at a depth of 60 feet. The berg was observed to quiver perceptibly the instant of the explosion, and several growlers were detached from the main body, but no other results were observed nor did the berg change in equilibrium. (Official photograph, international ice patrol.)

The melting of icebergs induces a circulation of the water in which they float. Pettersson (1904, p. 286) and Sandstrom (1915, p. 245) found by laboratory experiments that fresh ice melting in a tank of salt water tends to induce three currents: (1) A cold surface current of low salinity away from the ice; (2) an intermediate current of relatively warm salty water toward the ice; and (3) a cold current sinking diagonally away from the ice. Pettersson in several publications has offered these experiments as an explanation for many of the circulatory features of the world's oceans. Barnes (1910) as a result of ice investigations in the St. Lawrence River believes that the detection of such movements around bergs by means of precise

[63] Zeusler (1925, p. 41) gives a good description of mining operations as occasionally carried out by the ice patrol.

measurements of the temperature may be made to give warning of the proximity of bergs. An electrical resistance thermometer graduated to one-thousandth of a degree centigrade called a microthermogram was experimented with on a trip to Hudson Bay in July, 1910. Barnes (1910, p. 131) shows that a typical thermograph upon approaching a berg begins with a slight rise and then drops from 4.6° C. at 1 mile, to 2.2° C. at one-half mile. The primary peak and then the fall in temperature were explained as due to Pettersson's No. (1) current, i. e., to the thaw water spreading out on the surface, the outer fringe being heated most by the sun, was warmest. But Barnes's (1913) further investigations from the Canadian Pacific steamship *Montcalm* near the Strait of Belle Isle the summer of 1912 found, contrary to his earlier experiments, a continuous rise of temperature as bergs were approached suggesting the drop in the thermographs noted in 1910 was actually due to the generally low temperature of the cold Labrador current in which the bergs floated. Barnes concludes therefore that Pettersson's current No. (1) does not prevail under actual conditions at sea, but instead that there is an inflowing surface current toward a berg. Aitken (1913, p. 561) carried out a laboratory experiment similar to Pettersson's, finding that fresh ice melting in salt water induces a circulation whereby all of the thaw comes to the surface. Taylor (1914, p. 65), employing a microthermogram similar to Barnes's, failed to secure any definite reliable temperature graphs when the *Scotia* approached close to icebergs on the Grand Banks in the spring of 1913. Sometimes there was a rise, sometimes a drop, and usually the record was masked by fluctuations in the temperature due to causes other than icebergs. Thuras (1915, p. 67) approached bergs on the ice patrol ship using a thermometer graduated to one-tenth of a degree centigrade, finding that no definite rise and fall of the instrument was repeated with regularity. Cruising past some bergs the instrument would remain quite steady, while at others it would be very erratic. Aitken (1915, p. 561) criticizes Barnes's work in view of the contradictory results obtained by himself and by Taylor.

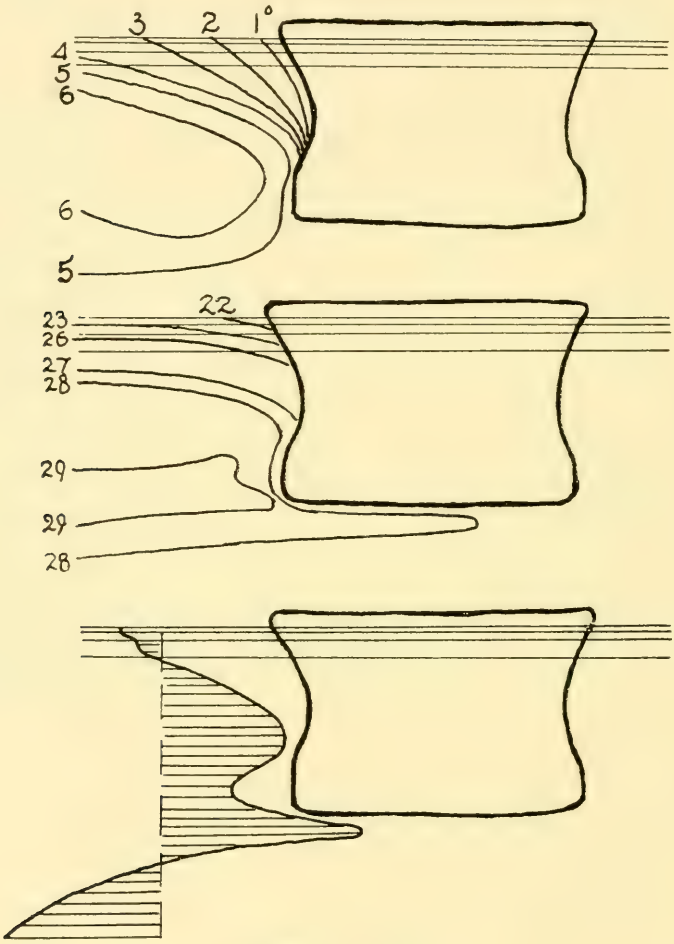
Barnes (1927, p. 92), in a recent article describes the circulation induced by melting icebergs as follows:

> Every iceberg is a hydraulic pump sinking the surrounding sea water by cooling and drawing to itself the warm surface waters and thereby contributes to its own destruction. The warmed surface layers flow more rapidly to the cold ice surface than do the cool layers, hence the iceberg becomes the central point for the collection of the warmer surface water.

By his view the warmed surrounding water is an indraft from the surface layers which compensates for the sinking thaw water. This horizontal movement keeps them uppermost and explains their greater insolation and warmth near a melting iceberg.

Ricketts (1930, p. 119) discusses the conventional theory of circulation around icebergs and he finds it difficult in view of the stratification and marked stability of the surface layers to believe that melting icebergs establish currents of any appreciable magnitude. He assumes, however, for the sake of argument, that an average-sized berg south of Newfoundland chills a 10-foot layer of the surrounding water to a temperature of 20° F. He then shows that under the given conditions the resulting horizontal indraft will be 27 feet per day

at a distance of 1,000 feet, and 270 feet per day (0.002 knot per hour) at a distance of 100 feet from the berg. Such a current is of such a comparatively small magnitude that it is masked completely by the passing surface movements due to the waves or the winds. As proof of the above, Ricketts points to the drift of growlers that even on days of light airs and breezes is away from the parent berg.



THE MELTING OF ICE

FIGURE 84.—Fresh ice melting in salt water, according to laboratory experiments, sets up a movement of water near the ice in much the same position shown in the bottom sketch. The upper two drawings show the distribution of temperature and salinity after a few moments of melting. (Sketches from Pettersson and Sandstrom, 1915.)

With conclusions so contradictory it is evident that the picture is far from clear. The experiments of Pettersson, Sandstrom, and Aitken are all open to the criticism that the laboratory conditions may not simulate those existing at sea. Barnes's conclusions are weakened by the fact that neither Taylor or Thuras obtained similar results, though Thuras did not have sufficiently sensitive instruments.

From our own observations at sea in the ice regions we can remark that the circulatory effect of icebergs is relatively so weak, and the many temperature variations in the surface waters so fluctuating that even the most sensitive thermal recorder can hardly be expected to serve as a reliable iceberg detector. It is hoped, however, that purely from the standpoint of scientific interest, information regarding the circulation induced by melting icebergs will soon be more firmly established.

Visibility and Mirage

Icebergs floating far out from shore in Davis Strait and in the North Atlantic are sighted at various distances depending upon the state of visibility, height of the berg, and of the observer.[64] Smith's (1927a) investigations in the ice regions south of Newfoundland state that bergs can be seen on a very clear day by a crow's-nest lookout at 12 to 15 miles, and they have been picked up under excellent visibility a maximum distance of 20 miles by a masthead lookout.[65] The bridge usually sights a berg on ordinary days at 10 to 12 miles, in clear weather, but with a low lying haze around the horizon, the tops of bergs have been seen 9 to 11 miles. There is a tendency to overestimate the distance, believing that one can see farther than is actually the case. In a dense fog a berg can not be seen more than 100 yards ahead of the ship, where it takes form as a luminous white mass if the sun is shining, otherwise it first appears close aboard as a dark somber shape. In a light, low fog an observer can see ice first near the surface of the sea, the initial signs being the lapping of the water on its base. A berg may be seen at a considerable distance on a clear, moonlight night, how far depending upon the altitude and size of the moon, and on the relative position of the moon, berg, and ship. Given a full moon at an altitude of 35°, a berg can be discerned at a distance of 8 miles, and is plainly visible to the naked eye as a glistening, luminous spot at 5 miles.

Mirages at sea often create fantastic berg shapes, the images appearing inverted and much larger than the actual berg proves to be when alongside. Each new berg sighted afar looks much larger than its predecessors nearer to, but upon closer approach both images gradually shrink; the upper snaps off and disappears in space, while the lower contracts until correct proportions are restored. Berg mirages are very frequent in the waters around Newfoundland, where the plane of atmospheric discontinuity is well developed in summer, a hundred feet or less above the water, the stratification of the air is sometimes so sharp that berglike reflections rise and fall even with the motion of the ship on the swell. There is a record of an iceberg the image of which was sighted over 20 miles away in the mixed waters south of Newfoundland; it first appeared as three bergs, one set upon the other, with the mirage continually changing shape as it was approached. An ice blink is a common phenomenon attending

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[64] Due to the flat opaque whiteness of icebergs and their consequent poor ability to reflect any particular color, a pair of binoculars fitted with amber sun shades are found a great help in scanning the horizon. Amber-colored spectacles for navigators in ice regions are also recommended.

[65] Zeusler (1925, p. 37) reports an incident where a berg was sighted at a maximum distance of 36 miles. Commander Roach from the cutter *Modoc* in May, 1930, reported sighting a berg under conditions of unusual refraction at a distance of 50 miles.

the presence of pack ice, but rarely, if ever, seen in the case of an iceberg.

Observations which have been carried out on the reflection of sound waves from icebergs emphasize the uncertainty of obtaining echoes. If the sides are steep or perpendicular, an echo will probably be heard from some directions but not from others, depending whether or not the face presented to the sound is normal or oblique. Echoes have been obtained from icebergs at a maximum distance of 1,000 yards under favorable conditions, but as a rule are inaudible at a quarter of a mile. A fantastically shaped berg with steep sides and a high peak toward the center was found to give echoes and reechoes, but on another occasion a similar-shaped berg gave none. An echo

AN ICEBERG IN THE FOG OFF THE GRAND BANK

FIGURE 85.—May 31, 1928, an iceberg in a light, low fog. This ice was visible from the crow's nest of the ice patrol ship when 3 miles distant, but was not visible less than 1 mile from the ship's bridge. Grave menaces to the trans-Atlantic steamships are the icebergs which are often hidden in fogs of the Grand Banks for weeks at a time. Bergs in light fogs and hazy weather often appear much larger than they actually are. (Photograph by Lieut. Commander N. G. Ricketts.)

around bergs during a fog may or may not come from the ice, as fog banks under certain conditions are known to be good sound reflectors.[66]

CURRENT AND WIND CONTROL DRIFT OF ICEBERGS

The first question to be considered in this connection is the relative importance of the two agencies, viz. the gradient current and the wind, which are responsible for the drift of icebergs. The proportion in which these factors combine may vary greatly and is dependent upon (*a*) the proportions above and below the surface of the sea at which a berg floats, (*b*) the velocity and duration of the wind, and (*c*) the velocity and depth of the gradient current.

[66] Johnston (1913, p. 22) relates a case where the Coast Guard cutter *Seneca*, lying between a berg and a growler, blew her whistle and received an echo from the growler but none from the berg.

There is little published on the effect of wind and current in controlling the drift of icebergs. Johnston (1915, p. 40) is of the opinion that the winds exert a very pronounced effect on the movement of bergs. Quinan (1915, p. 37) points out that although the wind has some effect the ocean current is the controlling factor. Krümmel (1911, p. 434) calls attention to the deep draft of icebergs, about 5 to 1, and concludes therefore that the wind has little effect and the deep-ocean currents are dominant. Drygalski (1895, p. 286) states that the föhn (east) winds drive the pack ice out of the west Greenland fjords in the spring and (p. 502) this frees the icebergs which have been blockaded throughout the winter. He does not mention, however, any tendency of the east winds to carry the bergs out into Davis Strait. Kane (1854, p. 104) speaking of the beset ship *Resolute* says, "their drift followed some system of advance entirely independent of the wind."

Mecking (1906, pp. 12–18) discusses the relative importance of currents and winds to drift the ice out of Baffin Bay. He quotes from the accounts of several observers on this subject, viz, Nansen, Greely, Blake, Drygalski, Nares, Kane, and Weyprecht. Most of the observations, however, regarding the efficiency of the winds as a dynamic agency refer to the drift of pack ice or to icebergs that are entangled in the pack. Few if any of the observations cited by Mecking refer to icebergs outside of coastal zones and consequently such bergs have not the freedom of movement as those floating in deep water. Mecking concludes that in regions where slope currents prevail the current is dominant, while in regions of weak currents the winds materially effect the drift of the bergs. He emphasizes this point on page 67 of Mecking (1906), by stating that the offshore winds along the glaciers in west Greenland during summer are the index of the number of icebergs drifting past Newfoundland the following spring.

The international ice patrol has made many observations on the relative effect of wind and current to govern the drift of icebergs. In 1922 we followed a number of bergs that drifted in the area south of the Grand Bank during a period of about two months, May 2 to July 13, with interesting results. The tracks of the different bergs entering this region all conform in general characteristics, as can be seen by examining the published charts (Smith, 1923, Charts 2 and 4), despite the fact that the winds varied much, both in direction and in force. In May the winds prevailed from the west and northwest, force 4, and during the latter half of the month were equally divided between the southeast and southwest quadrants, force 2.5. The first two weeks in July the wind changed to southwest with a mean force of 4. A striking example of this independence of berg drift and of wind direction was observed May 2 to 13, 1922, when bergs "D" and "E" (Smith, 1923, Chart F) drifted westward a distance of 90 miles at the rate of 9.5 miles per day in the face of a head wind blowing at times with a force 6. The fact that during this entire period of varying winds, the successive bergs exhibited a similar direction of drift is strong evidence against any appreciable effect of winds on icebergs. The general adherence of the ice to the stream lines of the gradient currents as shown by comparing the charts of berg drifts with the hydrographic

maps, which includes gradient currents only (figs. 103, 104, 105, and 106) is additional proof of the relative inefficiency of the wind.

This conclusion must be modified somewhat, however, when icebergs nearing the last stages of their career become extremely pinnacled and winged. An excellent example of such a case was provided in the summer of 1924 (see Smith, 1924a, p. 80) by a berg east of Newfoundland that we followed May 26 to June 12, a period of 28 days, over a total distance of 220 miles. (See figs. 86 and 87.) During the first two weeks while it was large and bulky (measuring 187 feet high and 387 feet long) it drifted in the generally accepted paths of gradient currents east of Newfoundland. The conclusion is corroborated by the salinity and the temperature of the water in which the berg was floating on June 5, as follows:

Depth (meters)	Temperature	Salinity
	°C.	
0	2.3	32.92
25	1.3	33.00
75	0.2	33.04
[1] 150	−1.2	33.28

[1] Bottom.

The upper 25 meters (14 fathoms) of the water column was of coastal origin subject to considerable fluctuation from the frictional effect of the wind. In the deeper layers was cold Arctic water, the general movement of which was toward the south. During the period May 26 to June 13 the berg drew at least 120 meters (396 feet), proof that its underwater body must then have been under the control of the deeper northern currents. On June 13, 1924, however, large quantities of ice were calved, the main body greatly changing form and reducing the draft. The drift from that date until the berg entirely melted on June 23 was irregular, not in conformity either with its former track or with the generally accepted direction of the main current. Such an erratic course can be best explained as due to control by the wind.

This view is corroborated by measurements taken of the berg's height, length, and draft on June 18 when its exposed body had a pinnacle at one end 106 feet high which sloped to a mound 30 feet above the sea at the other. Below water there were two large peaks extending downward 160 feet and 200 feet, respectively.[67]

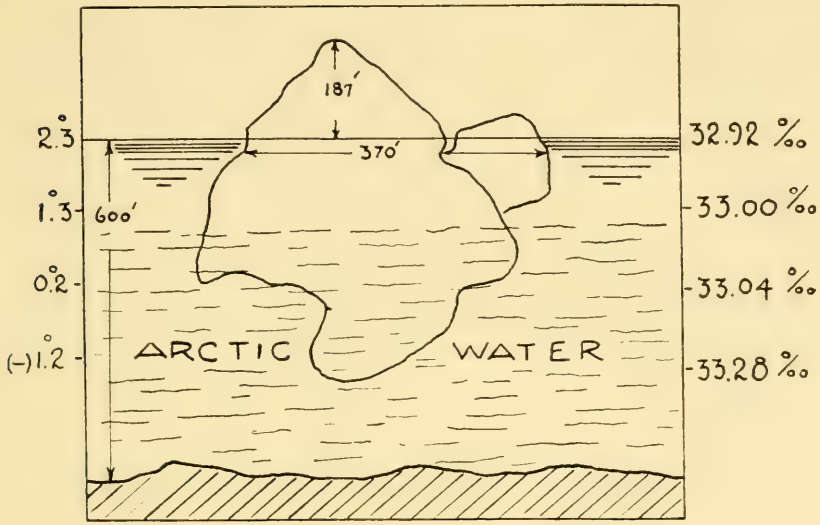
While the foregoing measurements were being taken the salinity and temperature of the water in which the berg floated were also observed.

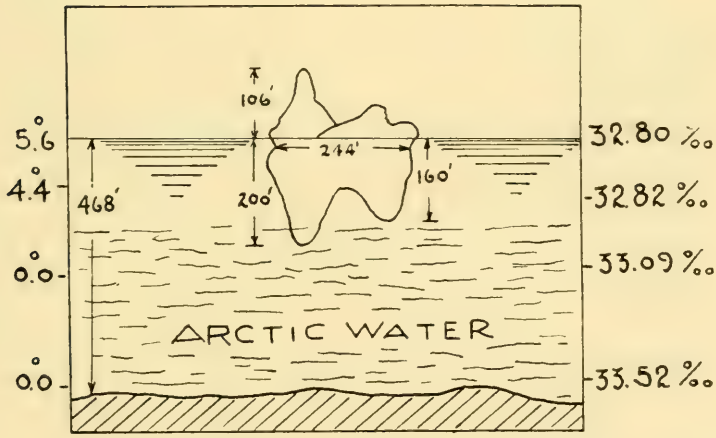
Depth (meters)	Temperature	Salinity
	°C.	
0	5.6	32.80
45	4.4	32.82
90	0	33.09
[a] 130	0	33.52

[a] Bottom.

[67] The measurements under water were carried out by two small boats lowering a heavy weight to a designated depth, the weights being connected by a fine wire about 450 feet long. Passing on either side of the berg the wire span was pulled taut and the weights lowered until the wire passed under the ice.

The transition in the character of the water column at a depth of about 75 meters agrees well with the stratification found June 5, with the exception that now the deepest pinnacle, 200 feet, fell just short of reaching down into the cold Arctic water on the bottom of the bank. These data clearly show that the berg was then floating



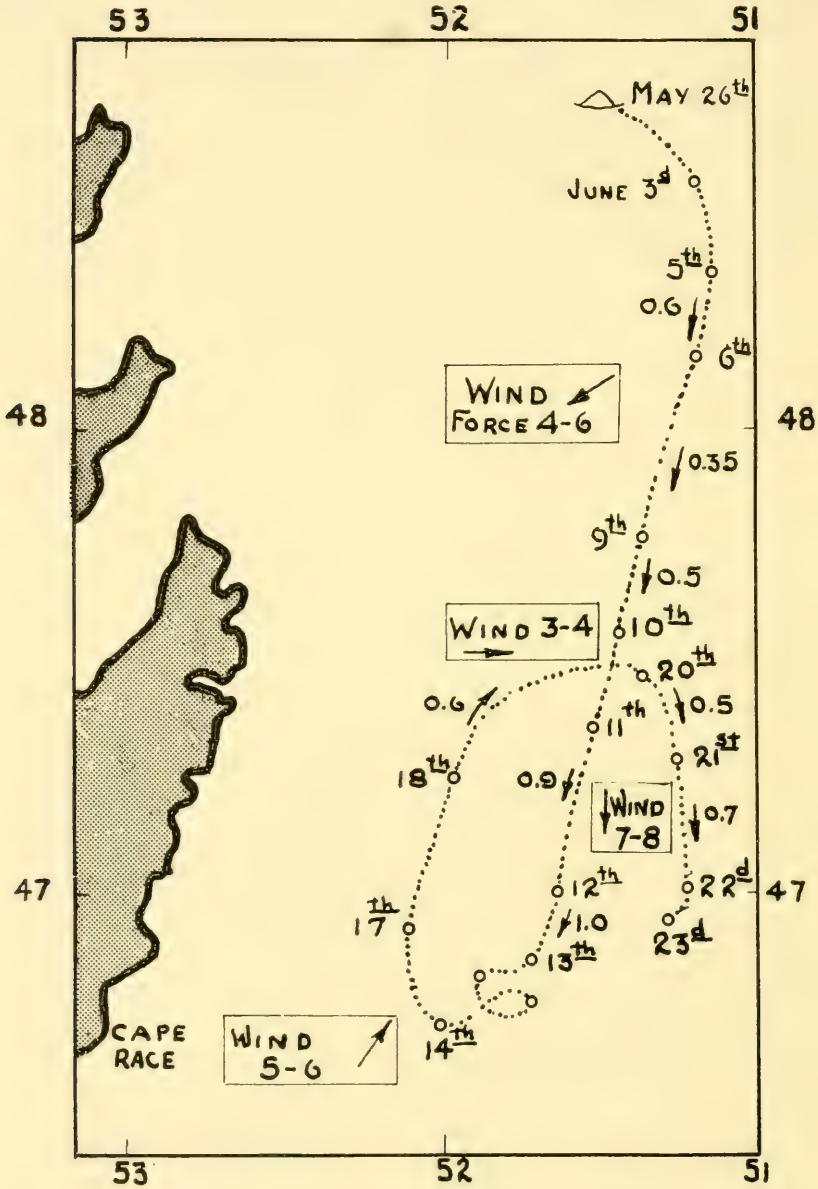


FORCES CONTROLLING THE DRIFT OF AN ICEBERG

FIGURE 86.—A berg on June 5, 1924 (see position on fig. 87), with its drift controlled by the deep-seated Arctic current. The lower sketch represents the same berg after a major calving on June 18, 1924 (see position on fig. 87).

wholly in the upper strata of the coastal waters, and being so shallow easily came under the control of the rapidly changing winds.

Many other berg tracks that we have followed to the melting end of a berg's life, similar to the case just described above, have eventually become very irregular, plainly indicating the loss of control by



FORCES CONTROLLING THE DRIFT OF AN ICEBERG

FIGURE 87.—The drift track of a large iceberg sighted May 26, 1924, under the control of the cold Arctic current. On June 13, 1924, it underwent a major calving, greatly reducing its draft. The marked change in the direction of its drift, as shown above, is attributed to the assumption of control by the wind and surface layers.

the ocean current and its replacement by the winds.[68] As a berg melts, the increasing irregularities of its contours give the wind a greater and greater hold.

The effect of the wind on icebergs is of two kinds, (*a*) the direct force of the wind as applied to the exposed surface of the berg itself and (*b*) the indirect effect of the wind as it tends to set up a frictional current in the surface layers of the sea.

(*a*) Few data, if any, are available regarding the force exerted by a wind on the sides of an iceberg. The pure wind effect on the patrol ship in the ice regions, exclusive of frictional wind current generated, is therefore employed as a guide. This is estimated to amount to 17 miles per day, with a wind force of 6 to 7 on the Beaufort scale and 11 miles per day with a wind force of 4 to 5. The ratio of exposed portion of the ship to submerged is about 2:3, while in the case of icebergs we have shown on page 112 this varies from 1:5 to 1:1, largely depending upon the form and state of disintegration. The following table gives the wind movement for the various type of bergs, based on that of the ship, assuming that the direction of its drift is the same as that of the wind and disregarding the coefficient of surface friction:

Class of iceberg	Drift of Ice [1]	
	Wind Beaufort 4–5	Wind Beaufort 6–7
Blocky	1.5	2.3
Rounded and domed	1.8	2.8
Picturesque Greenland	2.2	3.7
Pinnacled and ridged	3.7	5.7
Last stages, winged and horned	7.3	11.3

[1] Miles per day.

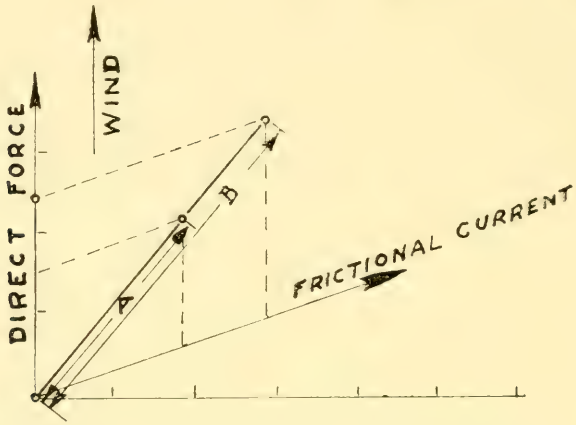
(*b*) Since Ekman (1905) has published his epoch-making theory of wind currents in the sea as deflected by earth rotation, much attention has been paid to the subject by dynamic oceanographers.[69] Admitting that the stratification of the water column and also several other conditions greatly modify the ideal theoretical development of a wind current, it may be estimated that a moderate to fresh wind blowing for a day or two in the ice regions will establish a movement of the water layers to a depth of 200 feet, and a strong breeze to moderate gale in the same interval to a frictional depth of 300 feet. The mean velocity of movement, moreover, will be that at the 80 and 120 foot levels, respectively, and the direction at that same level will be approximately 72° to the right of the wind. Under such conditions icebergs in the deep water off Newfoundland would be borne by the frictional current alone 72° to the right of the wind at a rate of 3.7 miles per day in the case of a 6 to 7 wind, and at 2.5

[68] Johnston (1915, p. 40) remarks that he has occasionally observed a berg with one or more high pinnacles catch the wind and sail along to leeward at the rate of 1 knot per hour. He claims that these "sailers" are capable of crossing the Gulf Stream. Quinan (1915, p. 37) is of the opinion also that the winds have a marked effect on icebergs, indirectly by influencing the current and directly, but in a less degree, on the bergs themselves. He points out that, although the current is the controlling factor, it is not true that the wind has no effect. He says (p. 37): "Had northerly winds continued, there is no telling how far south this berg would have traveled, but fortunately the southerly wind asserted itself and the bergs started northward with almost as much velocity as when they came south."

[69] See Bigelow (1927, pp. 962–970) and Smith (1926, pp. 46–50). Most recently Ekman (1928).

miles per day with a wind force of 4 to 5. There is, however, one modifying factor, namely, the deep draft of the ice, 500 to 600 feet, which exceeds the depth of the wind current. The bottom portion of the body dragging in the deep, dead water will, therefore, retard the ice probably as much as 20 per cent. In the case of the bulky and boxlike type of berg it will reduce the drift to 2 and 3 miles per day for 4 to 5 and 6 to 7 winds, respectively, but the smaller-winged and pinnacled bergs float entirely in the wind current and develop no drag whatsoever.

Now, combining (*a*) and (*b*) (p. 133), we arrive at a resultant drift which for the deeper-draft type of bergs is 40° to the right of



WIND EFFECT ON THE DRIFT OF LARGE ICEBERGS

FIGURE 88.—In the case of a wind with a velocity of 4–5 Beaufort scale, the berg will drift 40° to the right of the wind (in the Northern Hemisphere), at the rate of 2.8 miles per day; line A in this figure. In the case of a wind velocity of 6–7 Beaufort scale, the berg will drift 40° to the right of the wind at the rate of 4.3 miles per day; line B.

the wind. Column A represents the drift that the wind would impart to the berg if it did not affect the water and column B represents the mean frictional current for winds of the given strength. We have:

Wind strength	A Direct force [1]	B Frictional current [1]	Resultant drift [1]
Beaufort 4–5	1.6	2	2.8
Beaufort 6–7	2.5	3	4.3

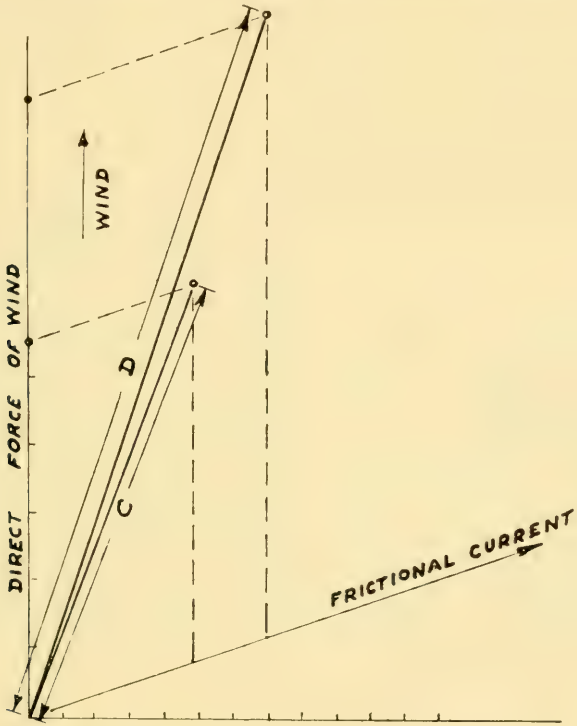
[1] Miles per day.

In the case of the lighter draft bergs, i. e., those of proportions 2:1 and 1:1, drift 54° to the right of the wind as follows:

Wind strength	Direct force [1]	Frictional current [1]	Resultant drift [1]
Beaufort 4–5	5.5	2.5	6.7
Beaufort 6–7	8.5	3.7	10.2

[1] Miles per day.

These tables tend to corroborate earlier findings that the bulky, full-contoured bergs are moved but little by the variable winds, force 4 to 5, which ordinarily prevail south of Newfoundland during the ice season, while the light-draft bergs, many examples of which come south of the Grand Bank during spring, are appreciably affected by the winds just as the tables show. In certain other regions, on the other hand, the winds may exert an important effect. Along the Labrador and Newfoundland coast during winter, for example, a series of northwesterly gales such as sometimes last for days at a



WIND EFFECT ON THE DRIFT OF SMALL ICEBERGS

FIGURE 89.—In the case of a wind with a velocity of 4–5 Beaufort scale, the berg will drift 54° to the right of the wind (in the Northern Hemisphere), at the rate of 6.7 miles per day; line C on this figure. In the case of a wind velocity of 6–7 Beaufort scale, the berg will drift 54° to the right of the wind at the rate of 10.2 miles per day; line D.

time must move the ice, especially the pack, both forward and to the right, tending to carry it in on the coast. Mecking (1906, p. 12) states that in order to move an iceberg as ordinary currents do, for example, 10 to 12 miles per day, a wind must have a force of 6 to 7 Beaufort scale. According to our observations, the only bergs that approach any such wind movement are the winged and fragile types in the last stages of disintegration. The deeper draft bergs common to Baffin Bay are moved only 4 miles a day by a wind, force 6 to 7.

In the case of the deeper bergs drifting within the bounds of an ocean current such as the Labrador current, small influences due to

frequently shifting winds are masked by the simultaneous movements imparted by the much steadier and more enduring slope current. The berg may be likened to a chip floating down a swift-running river with small eddies carrying the object in revolutions, or other tortuous paths, the while it progresses downstream.

An opportunity to compare the variations in drift due to the winds frequently becomes available on international ice patrol duty when a group of bergs is encountered at sea. During a period of even moderate to fresh winds the shoaler berg will drift about 14° to the right of the heavier ones in 24 hours and leave the latter some 4 miles astern, the bergs alternately separating and congregating with the play of the winds.

THE GENERAL DRIFT AND FATE OF ICEBERGS IN THE WESTERN NORTH ATLANTIC

The general drift of the icebergs, once they have left the Arctic fjords and reach open water, is governed chiefly by the general direction and rate of flow of the ocean currents. Pack ice in heavy floes driven for days with tremendous force by gales must tend to sweep many bergs into regions which they might not otherwise enter, leaving drift tracks not conforming to the average. Many bergs may also be detained by the contours of Baffin Bay and Davis Strait, such as shoals, capes, and off-lying skerries. Belts or bands in the currents sort out the bergs; if a berg be in one of these it is assured rapid transportation, while another berg within plain view may float in dead water. The Gateway to the Atlantic, off the Tail of the Grand Bank, marks the major turning point in the iceberg journey, because, no longer guided by the continental slope of North America, the ice is borne directly into the easterly moving masses of the Gulf Stream. The variations in the drifts of icebergs are many; two detached from the same glacier on the same day may be separated by one or two years in their Atlantic arrival. The length of their total pathway, 1,500 to 1,800 miles, and the average velocity of the current, 10 to 12 miles per day, suggests that a berg meeting no hindrances completes the journey in four or five months. As a matter of fact, this can not often happen and the usual history is to be released from the fjords in the summer, to reach the region of Hudson Strait in autumn, to "winter" there, and to appear off Newfoundland the following spring. A berg calved from its glacier during the winter might be released from the fjord in June and spend the summer drifting out to sea, winter in Melville Bay, be released there in the following summer, reach Cape Dier by that October, and arrive south of Newfoundland with the main body of bergs the next May. The sorting of this kind to which the bergs and other flotsam are subjected in the far north is well illustrated by comparing the drift of the Polaris floe party with that of the British steamer *Fox*. In seven months the former drifted directly down through Baffin Bay and Davis Strait, making a southing of approximately 1,740 miles, averaging 9 miles per day; while in eight months the *Fox*, also drifting helplessly in Baffin Bay, made less than one-third that direct distance (510 miles), averaging only 2 miles per day. The *Fox*, however, was carried on a very serpentine track that really amounted to 1,194 miles.

A major factor slowing the southward progress of icebergs is the deeply indented, broken character of the North American coast line from Ellesmere Land to Newfoundland. In the distance from Cape York to Cape Race the continuity of the shore line is interrupted in five places; and while three of these interruptions are so far north they have little effect on the iceberg stream, the 50-mile breach at Hudson Strait and the 15-mile opening at Belle Isle have an important effect. The complex movements of the water masses athwart the berg stream, as indicated by the irregular courses of the isotherms and isohalines of the *Marion's* survey, extending far out in the Labrador Sea, mark Hudson Strait as the source of much turbulence. The great indentation in Newfoundland which stretches from Cape Bauld to Cape St. Francis is a natural cachment for bergs, and

ICEBERGS MOVING OUT INTO BAFFIN BAY

FIGURE 90.—Icebergs discharged from the Great and Little Karajak Glaciers drift westward through Umanak Fjord into Baffin Bay. Great Karajak, one of the most productive glaciers in West Greenland, is estimated to discharge 1,200 sizable icebergs annually. The little native settlement of Umanak lies in the foreground. (Photograph by A. Bertlesen.)

hundreds of other small coastal configurations also catch the wandering bergs. Lady Franklin Island, a small rocky eminence lying about 30 miles northeast of Frobisher Bay on the southeastern coast of Baffin Land, forms a spur in the side of the ice stream. During certain parts of the year the northern side of the island and the long thin line of reefs to shore catch hundreds and hundreds of icebergs. Had Labrador and Newfoundland a bold, clear coast line or one similar to that of east Greenland, the annual supply of bergs to the North Atlantic would be manyfold greater.

The fields of pack ice which for four months of the year obscure all coastal contours also greatly modify the effect of the shore line on the berg drift. At a normal rate of 10 miles per day for the cold current, a band of water 1,200 miles in length will pass a given point during this 4-month period. Normally a total of about 400

icebergs scattered along this 1,200-mile train are carried past Newfoundland annually. Then the withdrawal of the pack from south to north bares the coast and the iceberg file is severed.

Such bergs as escape these traps are embraced in the 90-mile-wide shelf waters fringing the American side of Davis Strait, Baffin Bay to the Grand Bank; this zone provides an ideal straightaway sea road on which the ice travels Atlanticwards.

Distribution and Drift of Icebergs in Baffin Bay and Davis Strait

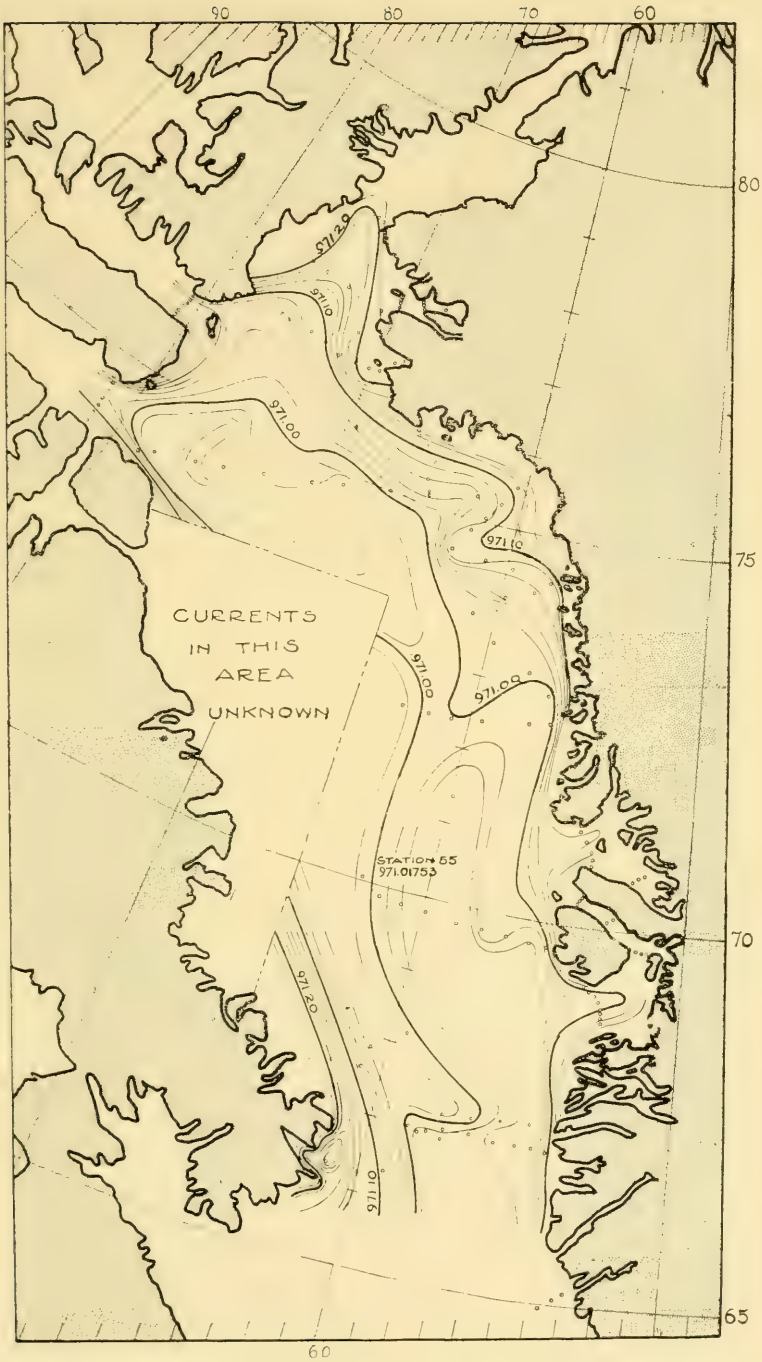
Any study of the geographical distribution of ice in the Arctic emphasizes the peculiar character of the northwestern arm of the Atlantic that separates North America from Greenland, for Davis Strait is flanked on one side, and Baffin Bay on both sides, by high walls of land ice making it unique for the Northern Hemisphere. Continual cleavages from the ice walls discharging into Baffin Bay and Davis Strait mark these regions as the most important iceberg waters of the north.

Along its eastern shore for a distance of 500 miles Baffin Bay is continually receiving a supply of icebergs, and no other body of water of this restricted size and inclosure becomes so charged with land ice. The iceberg crop which characterizes the waters of Baffin Bay and Davis Strait is believed to be completely renewed there every three or four years, and the fact that Baffin Bay does not become choked and clogged shows that the movement of its water masses are sufficient to effect the dispersal of the ice. The eastern and northern section of the bay present the densest berg concentration and the offing of the glaciers is the zone of greatest scattering.

Our knowledge regarding the circulation of Baffin Bay was greatly extended as a result of the *Godthaab* expedition in the summer of 1928. The *Godthaab* occupied about 100 oceanographic stations distributed more or less uniformly throughout the bay except along its southwestern side, off the coast of Bylot Island and Baffin Land, where pack ice made those waters inaccessible. In the absence of the text report of the scientific results of the *Godthaab* expedition we have employed the station data already published in the Bulletin Hydrographique (see Annually 1929, pp. 36–43), to construct a dynamic topographic map in accordance with the methods described by Smith (1926). The map has been used strictly to learn the general movements of the heavy deep-drafted icebergs once they have drifted away from the land and come under control of the ocean currents.

The map (fig. 91) which indicates the general direction and flow of the gradient currents is especially interesting and important in that it corroborates earlier assumptions of a cyclonic circulation for Baffin Bay, i. e., a northerly flow along the Greenland side and a southerly one along the American shore, the latter being the stronger and the more voluminous.

The surface topography of Baffin Bay in the summer of 1928 was featured by an elongated depression occupying the major central portion of the bay, while around the sides the sea surface in all places stood the highest. This picture we have no reason to doubt, is more



RELATIVE TOPOGRAPHY—1,000 DECIBARS

FIGURE 91.—The dynamic topography of the sea surface of Baffin Bay. The map has been constructed from the observational data collected by the *Godthaab* expedition (Commander E. Riis-Cartensen), and published in the Bulletin Hydrographique pour l'Annee, 1928, pp. 87–95. The reference station is No. 55, the dynamic height of which above the 1,000-decibar plane is 971.01753 dynamic meters. The isobaths are drawn for every two dynamic centimeters.

THE SURFACE CURRENTS OF BAFFIN BAY

FIGURE 92.—The arrows indicate the direction of the surface currents, and the numbers on the map indicate the velocity in miles per day

or less representative of oceonographic conditions in the bay as they normally tend to prevail. The troughlike depression extended from Disko Island northwesterly to the entrances of Jones Sound and Lancaster Sound, with the lowest points centered in two pools, one about 100 miles east of Lancaster Sound Entrance and the other about the same distance west of Disko Island. The highest elevation of the sea surface was found in Smith Sound, where off Etah, Greenland, it rose 26 dynamic centimeters above the lowest central areas of the bay. The steepest slope, according to the calculations, was in the lower end of the bay off Cape Dier where a gradient of 17 dynamic centimeters in a distance of 30 miles was recorded.

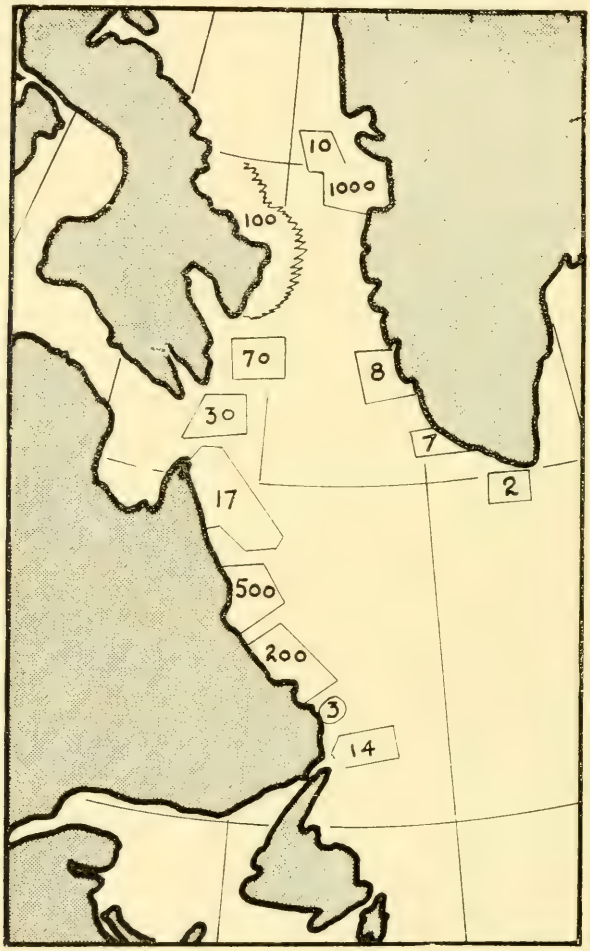
The slope currents on the Greenland side resulting from the above described topography took form as a sluggish set moving northward along the front of the principal iceberg fjords at the rate of 5 miles per day. This circulation provided for the immediate removal of the bergs that had been carried out of the fjords into the open coastal zone. The ice which remained inside the headlands, however, was subject to another sort of behavior.

About 1,000 bergs were observed by the *Marion* expedition the summer of 1928 distributed in Disko Bay. In whatever parts of these waters the *Marion* cruised, 100 to 200 bergs were constantly in sight from the ship but they appeared to be most numerous along the southeastern shore of Disko Island and around the mouth of Jacobshavn Fjord. A particularly interesting sight was the row of bergs always to be observed strewn along the southern shore of Disko Island, where they appeared to ground temporarily, awaiting the arrival of an unusually high tide or an offshore wind to start them again on their westward journey out into the open sea. During the *Marion's* excursion of the Vaigat, the 6 to 10 mile wide strait which separates Disko Island from the mainland, about 60 bergs were always in sight from the ship at once, most of them in the vicinity of Torsukatak Fjord, where they were noticeably drifting out of its northern side. A comparison of the areas of greatest iceberg abundance in Disko Bay with the current maps, Figures 91 and 92, shows good agreement in the main with the southern, major, portion of the bay as far as Claushavn dominated by an indraft from Davis Strait.

The *Marion's* section across the Vaigat did not indicate such a northwesterly discharge but the *Godthaab's* data, consisting of two sections, one not far from Jacobshavn and the other at the outer end of the Vaigat, both indicate it, and prove, therefore, that once the bergs clear the coastal waters they will be set slowly but definitely northward. This statement is contrary to the belief of Mecking (1906, p. 68), who stresses the importance of east winds to drive the icebergs across Baffin Bay into the south-flowing current along the American shore. While the bergs are still in the quiet fjords, offshore winds, no doubt, play a definite part in hastening the ice out to sea, but once they have drifted away from the inshore waters they will normally move northward. Under a 5-knot-per-day gradient current, moreover, the ordinary deep-drafted berg will be governed chiefly by the current, even if a continuous easterly gale prevailed. (See p. 134.)

The currents as drawn from the physical data collected by the *Godthaab* expedition are further supported by the distribution of the icebergs in the waters north and west of Disko Island. Vessels

bound south through Baffin Bay in September report bergs as especially numerous in a rectangle bounded by parallels 72 and 69 and meridians 62 and 57, and south of this very little ice is seen. The position of this field north and west of the most productive berg district is an indication of the direction of drift of the glacial ice. Since other parts of the bay at this particular time of the year usually have fewer bergs, this group, therefore, in all probability represents the summer crop.



ICEBERGS

FIGURE 93.—The distribution of icebergs in Davis Strait as observed by the *Marion* expedition the summer of 1928.

Immediately north of Upernivik the current and bergs spread out and the entire expanse of the surface waters from Upernivik to Cape York, and offshore for a distance of 90 miles, partakes of a slow movement toward the northwest, averaging about 5 miles per day. The direction of the current is generally parallel to the coast but its course is winding and, therefore, a berg arriving at Uper-

nivik in June will not reach Cape York in the circuit of Melville Bay until well along in August. The current embraced by this stretch of coast line appears quite uniform in that it is free from bands and the ice floating in its inner margin is carried along at approximately the same velocity as that many miles farther out from the coast.

The inner side of the current, and the distribution of bergs, upon reaching the meridian of Cape York accelerates to 7 miles per day and the map of this particular locality (fig. 91) as noted in the crowding of the isobaths, clearly indicates the tendency of the currents to divide into parallel bands. Off Cape Alexander the map shows the flow fans out to the westward and at the same time the

PROCESSION OF BERGS MOVING PAST DISKO ISLAND

FIGURE 94.—One of the main routes followed by the icebergs after discharge from Jacobshavn Fjord is westerly along the southern shore of Disko Island. Past the radio masts of the little Arctic village of Godhavn there is always a procession of icebergs drifting into Baffin Bay on the start of a long journey toward the Atlantic. (Official photograph, *Marion* expedition.)

current slackens to 5 knots per day, corroborating MacMillan (1928) (p. 429). One branch winds slowly across toward Lancaster Sound, another flows westward toward the entrance of Jones Sound, while the major portion of the water masses proceeds in the longer, circuitous route of Smith Sound at the rate of 7 miles per day. If bergs remain in this latter set, and there is little doubt but what they do, such ones will be carried northward as far as the seventy-eighth parallel of latitude before being turned sharply to the southwest and south to follow down the American side. Observations regarding the distribution of bergs, it is interesting to learn, corroborate the current map, fig. 92, constructed from the *Godthaab's* data, since

bergs are often reported as abundant in Melville Bay and in the northern part of Baffin Bay, where they congregate around the Carey Islands.

The cyclonic movement which has been traced along the eastern and northern side of Baffin Bay is augmented by inflowing currents from Jones Sound and Lancaster Sound. The dynamic topographic map (fig. 91, p. 139) plainly indicates that there is an indraft along the northern side of each of the above straits, which combines with a discharge out along the southern shores. When the *Godthaab* took its observations each of the straits, Jones and Lancaster, were dominated by a west to east current, from the Arctic into Baffin Bay, moving at the rate of 7 to 10 miles per day. A very narrow counterflow not over 2 or 3 miles in width hugged the north shore of each opening.

Bergs in the current would at first enter for a short distance on the north side of Jones Sound and then be carried out southeastward into Baffin Bay, thence through the same general movement in the eastern entrance of Lancaster Sound. In neither one of these straits, moreover, would the bergs succeed in setting very far to the westward because they would soon meet the strong discharge which normally flows out into Baffin Bay.

The closely packed position of the isobaths and a current of 7 to 10 miles a day (fig. 92, p. 140) out along the southern side of Lancaster Sound indicate this as the swiftest flow recorded by the *Godthaab* and also mark this region as the source of the icy current to the North Atlantic. If we should be asked to point to the headwaters of the famous Labrador current which flows to lower latitudes, we would unhesitatingly select this locality as the fountain head. After the bergs have slowly been carried in the circuit of Baffin Bay and they have entered the discharge from Lancaster Sound they take on henceforth a swifter, more definite transportation heading directly toward Davis Strait and the Atlantic.

The gradient currents in the waters along the Baffin Land coast between Lancaster Sound and Cape Broughton, Baffin Land, in latitude 67° 40′, have never been surveyed so thoroughly as the other parts of Baffin Bay,[70] but from various sources of information, such as the drift of ships beset by the pack, example (*Resolute*, see p. 48), and the behavior of other waters under similar conditions, a well-marked southerly set paralleling the coast is definitely established. The most northerly observations on the Baffin Land side, near Cape Broughton in latitude 67° 40′, were made by the *Godthaab* the summer of 1928. The stations 160 to 165 (see Annually, 1929, p. 42) indicate a continuation of the current previously mentioned and recorded farther north off Bylot Island.

This flow, which we shall call the Baffin Land current, embraced a band of water 15 miles in width, the axis of which lay about 12 miles out from the coast. Icebergs keeping in this belt would have been transported southward at the rate of 12 miles per day, but farther offshore such bergs would not be carried faster than 4 miles a day, yet such southward progress would be guaranteed, neverthe-

[70] This unexplored oceanographic region, so marked on fig. 91, is made inaccessible to navigation by pack ice which persists there except in only an occasional summer when for a brief week or more these waters may be entered. The *Godthaab* found this stretch off the Baffin Land shelf unpenetrable the summer of 1928.

less, to ice even 60 or 70 miles out from the Baffin Land side. Off Cape Dier, Baffin Land, in latitude 66° 30′, the swiftest, inner side of the cold stream split, one branch coursing in under the land, and the other maintaining its distance from the coast and continuing in a more or less parallel direction. Between these two branches the *Godthaab* found a counterclockwise eddy.

The position and rate of flow of the Baffin Land current as found by the *Godthaab* in September, 1928, off southern Baffin Land, were further substantiated by the oceanographic observations, and the map (fig. 95) compiled by the *Marion* expedition in mid August of the same year. A winding floe, 50 miles in width, is clearly indicated on Figure 95, curving around Cape Dier and thence continuing southward toward Hudson Strait. The bergs on the inner side of this floe are borne along at the rate of 10 miles per day, and probably faster, 12 miles per day, farther inshore where the *Marion* was unable to buck the pack ice. The *Marion* found that the ice in the outer, eastern, side of the Baffin Land current moves slower the farther out it is from land, but the current is so broad that bergs even halfway across Davis Strait are slowly but definitely transported southwestward by it.

One of the most interesting parts of Baffin Bay from the viewpoint of its circulation and the behavior of the ocean currents is the lower end of the bay where its waters escape southward through the narrow neck of Davis Strait. Considerable importance attaches itself to the movement here, not only as it vitally affects the drift of icebergs, but also as it indicates whether or not masses from the North Atlantic flow into Baffin Bay and to what degree. Consideration of the observations made by the *Michael Sars* in 1924, and the *Godthaab* and *Marion* in 1928, regarding the position and form of the isobaths drawn from the foregoing data (see fig. 91) clearly shows two main forces present. There is the light water along the Greenland coast, and over the off-lying banks Sukkertoppen to Egedesminde, which initiates a dynamic, northward movement of the water particles; while on the American side, 180 miles to the westward, similar forces are at work developing a gradient current in the opposite direction. The important feature, well portrayed by the observations, is the fact that the steeper and larger dynamic gradients belong to the western side of this narrowest part of Davis Strait.

A comparison of the data collected by the three expeditions, viz. *Michael Sars* in 1924, *Godthaab* in 1928, and the *Marion* in 1928, reveals plainly that there is a continual fluctuation in the position and velocity of the two opposing movements in this narrow neck of Davis Strait. As one current swells, the other retreats, but as a rule, the Baffin Land flow appears to be the stronger and the more voluminous. For example, the first week in August, 1928 (see fig. 95), the west Greenland current was found to have little movement, while the Baffin Land set had a width of 40 miles and a rate of 12 miles per day. Early in September, however, about a month later, the *Godthaab* (fig. 91) found the Baffin Land current had overflowed more than halfway across the strait, and the west Greenland stream had been pushed over against the Greenland coast and now had a width of only 12 miles. The eastward encroachment of the Baffin Land current is plainly discernible in the form and the position of the isobaths between parallels 67 and 68. (Fig. 91.)

The behavior of the currents in this particular region has been investigated with considerable interest because interactions of the two flows must naturally exert an important effect on the number of bergs that are carried southward out of Baffin Bay. The fact that the cold Baffin Land current at times spreads out nearly three-quarters of the distance across to Greenland indicates that some of the bergs at that time must be carried eastward away from the axis, and the swiftest part of the current, and therefore diminish their chances of reaching the Atlantic. On the other hand, during such periods of flood of the Baffin Land current, more bergs are liable to be carried southward through the neck of Davis Strait than otherwise, and a heavy ice year may be expected off Newfoundland. However, if the west Greenland current experiences an abnormal intensification, it will tend to dam the cold polar stream and thus reduce the number of bergs which are borne southward out of the Arctic.

The general bounds of the currents in lower Baffin Land and Davis Strait as determined from the oceanographic observations of the *Marion* and *Godthaab* are more or less confirmed by the distribution of the icebergs noted by the *Marion*. The striking feature of their distribution in the narrow neck of Davis Strait in August, 1928, was that no bergs were sighted three hours, i. e., 25 to 30 miles, after the *Marion* had departed from Godthavn to cross Davis Strait. No ice whatsoever was sighted out in the central part of the strait, nor until the ship had arrived within 40 miles of the Baffin Land coast, off Cape Dier. The absence of bergs in the middle of the strait at that time strongly indicates that the ice does not follow a direct path from the glaciers across to the south-flowing current under the American shore. This condition, combined with the fact that only very few bergs were sighted to the southward in Davis Strait, corroborates the oceanographic observations, namely, that the water and the ice tend to follow a cyclonic circuit of Baffin Bay. The general behavior of the pack ice as described on page 47, also fits in well with the picture of the prevailing currents as shown on Figure 92.

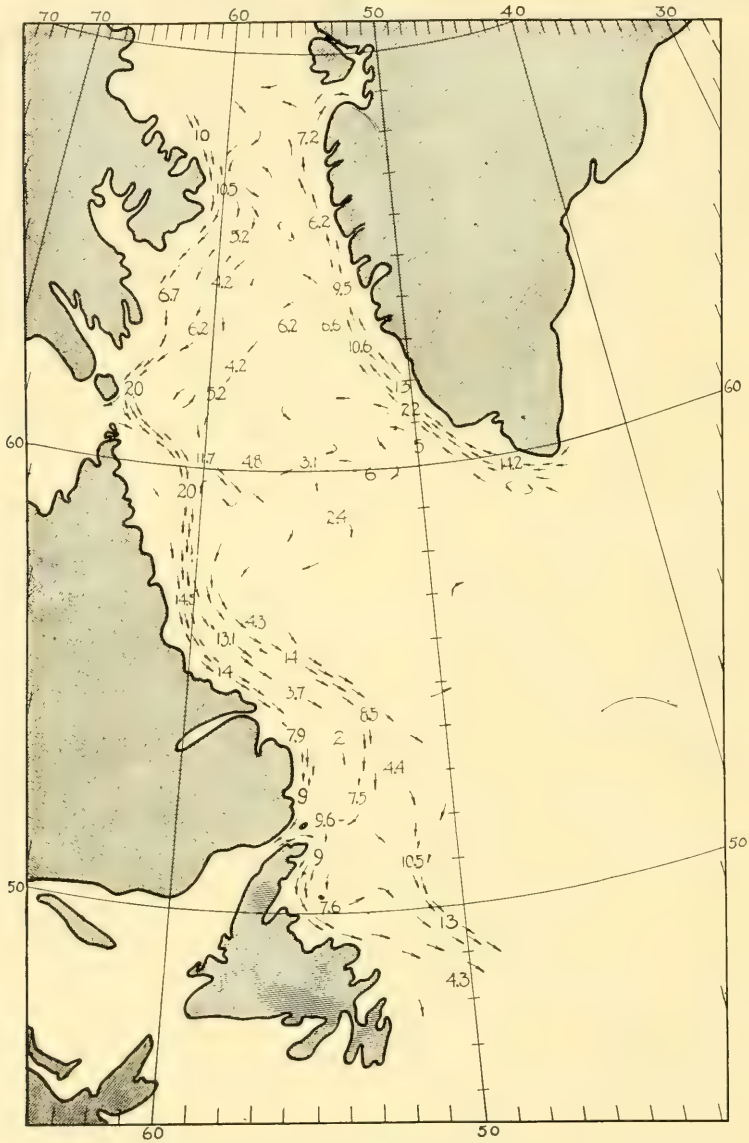
According to the *Marion's* survey, the Baffin Land current continues down the coast toward Hudson Strait at a speed of 6.7 miles per day, spreading to a width much greater than the width of the compensating stream on the Greenland side. Scattered bergs sighted 90 miles out from the Baffin Land coast were moving southward in the set. As the flow approached Hudson Strait, we found it narrowing to only 20 miles in width and simultaneously accelerating to 20 miles per day, to turn into Hudson Strait close under Resolution Island, an indication of the most probable course for the ice. McLean (1929, p. 13) says icebergs also enter Hudson Strait via Gabriel Strait, between Resolution Island and Baffin Land. This agrees, furthermore, with MacMillan's verbal statement that he has witnessed bergs drifting into Hudson Strait in narrow rows along the northern side for a distance of 150 miles, as far as Big Island, where nearly all of them recurve, to drift out again past Cape Chidley on the Labrador side. Not infrequently bergs drift up the strait as far as Salisbury Island before being caught in this current and very rarely, according to Huntsman (1930, p. 3), one is reported in Hudson Bay itself.

As previously remarked, the berg stream becomes much dispersed in the offing of Hudson Strait. The outer bergs may for a time lie in

RELATIVE TOPOGRAPHY—1,500 DECIBARS

FIGURE 95.—The dynamic topography of the sea surface of Davis Strait, July 19 to September 11, 1928, as determined by the *Marion* expedition. The sea-surface relief has been obtained by comparing it with that of a plane at a depth of 1,500 decibars. The dynamic height contours (isobaths) are drawn for every two dynamic centimeters. The survey is based on 191 oceanographic stations represented by the small circles on the map. The reference station is No. 947, the dynamic height of which above the 1,500-decibar plane is 1454.54390 dynamic meters.

dead water, but eventually they are likely to be carried again to the southward along the Labrador shelf. The inshore bergs that enter the strait follow a longer and more circuitous route but a more cer-



THE SURFACE CURRENTS OF DAVIS STRAIT

FIGURE 96.—The arrows indicate the direction of the surface currents and the numbers on the map indicate the velocity in miles per day.

tain one. Probably the most important feature in this locality is the narrow banding of the current in the offing of Hudson Strait.

The chief high road for the Arctic ice to lower latitudes is along the Labrador and Newfoundland shelves, and knowledge of the state of sea ice and currents in these waters during the winter and early spring is essential for corroborating the iceberg menace to the southward in the Atlantic during later spring and summer. The Labrador coast is seldom free of icebergs, an intermittent procession of which moves slowly southward. The ice, due to the effect of the rotation of the earth, tends to hug the coast, working in, out of the axis of the current, and much of it strands, sometimes permanently, but sometimes floats again and continues its southward journey. That hydrostatic forces, acting locally, are sufficient to maintain a southward current here, along the continental shelf, independent of additional momentum from the north, is clearly proven by the dynamic topographical maps. Iselin (1930), employing Bjerknes's formulæ, calculated that the current off Nachvak Fjord was confined to a breadth of 25 miles over the steepest part of the continental edge, moving at 20 miles per day but farther south off Sandwich Bay, the stream was broader 80 miles and slower, 8 miles per day. He observed furthermore that the bergs were characteristically strung out in lines more or less parallel with this part of the coast, moving fastest out near the continental edge, slower inshore. The *Marion* also found the Labrador current clearly banded, with the belt over the continental edge, about 20 miles wide, flowing off Nachvak at the rate of 20 miles per day. (See fig. 96 p. 148.)

Tracing the stream southward we find it approaching the coast in mid-Labrador, where it slows to 14 miles per day, while 60 miles farther out, it flows at the rate of 11 miles per day. As we proceed southward to the offing of Hamilton Inlet we note that the current widens to a breadth of 120 miles off Sandwich Bay, and changes in velocity so that the outer edge progresses at the rate of 14 miles per day, while inshore it is only 3.7 miles per day. This condition agrees well with those found by the *Chance* (see Islein, 1930), who describes the entire breadth of the water over the coastal shelf off Sandwich Bay as being in movement. (See our fig. 96.) Farther south there is an indraught to the Strait of Belle Isle and a band of the current on the continental edge off Newfoundland is observed to flow at the rate of 13 miles per day. The banding of the current and its importance on th eiceberg drift is clearly shown on Figures 95 and 96. From an examination of these figures a mean velocity of 12 to 14 miles per day is indicated for the Labrador current.

The undulating course of the stream lines on Figure 95 emphasizes that the Labrador current does not flow straightaway southward, and the well-marked inshore swirl there indicated between parallels 55 and 57, where the shelf shows its only major embayment, is certainly more than a mere coincidence but almost certainly a prevailing characteristic of the current and one which normally deflects a number of icebergs inward to the coast where they become trapped at this place. Years characterized by few icebergs south of Newfoundland will record an abundance of icebergs in this section of Labrador.

In fact, it is known that more bergs strand in this locality than at any other point along the Labrador shelf, and the summer of 1928 was an excellent time to reveal such behavior. This appears clearly from the distribution of bergs along the coast plotted from the

Marion's records and from other various sources: Cape Chidley to Hebron, 12 bergs; Hebron to Cape Harrigan, 500; Cape Harrigan to Cape Harrison, 200; Cape Harrison to Belle Isle, 0; Belle Isle offing, 14. This shows that 700 out of the total of 726 bergs were stranded on the coast in precisely the region where the *Marion's* current map shows the prominent inshore encroachment of the stream lines. Our repeated experience with a similar inshore swirl of the current in the Grand Bank region and the resultant drift of the ice emphasizes the importance of hydrographical features of this sort.

Pack ice was noticeably scarce in 1928, as proved not only by the observations of the ice patrol during the spring but also by early reports along the Labrador coast.[71] The failure of the pack to appear in normal quantity and blockade the coast line permitted the Labrador current to set the icebergs farther inshore than usual so that more stranded.

The course of the current and of the ice widens off Newfoundland. The form of the dynamic isobaths, on figure 95, page 147, in the offing of the Strait of Belle Isle, indicates the presence of an anticyclonic eddy which often controls the movement of icebergs especially in summer and early fall. There is also indicated some inflow into the entrance to the Gulf of St. Lawrence, and for that reason bergs in considerable numbers pass through the strait during the season. A few bergs are even carried far into the gulf along the Quebec shore; they have even been sighted off Cape Whittle, 200 miles in. Such ice rarely, if ever, reaches the Atlantic again, but disintegrates in the gulf, a process helping to cool these waters.[72] A few bergs also linger during the summer in the offing of the Strait of Belle Isle.

The current that skirts the east coast of Newfoundland sets a considerable number of bergs and growlers on that shore during the late season after the pack ice has disappeared. The observations of the ice patrol during a cruise in May, 1924, to this region illustrate the drift of ice in this great coastal bight. (Smith 1924a, p. 76.) Farther south the current, and probably the bergs to a certain extent, reaches the northern part of the Grand Bank region in two belts, one following along the steep part of the continental slope and the other close in to the coast. This inshore stream carries ice as far south as Cape Race and even farther westward around the latter through the Gully, between the coast and the banks.

Drift of Bergs South of Newfoundland

The drift and fate of icebergs in the region south of Newfoundland have been intensively studied by the ice patrol for the past 15 years. The published reports: Chiswell (1923, 1924); Fisher, (1920, 1926, 1927); Fries (1922, 1923); Gamble (1922); Johnston (1913, 1913a, 1915, 1915a, 1920); DeOtte (1921); Molloy 1930); Quinan (1915); Ricketts (1929, 1930); Smith (1922, 1922a, 1923, 1924, 1924a, 1925, 1926, 1926a, 1927, 1927a, 1927b, 1927c, 1929, 1929a); Thuras (1915, 1916, 1921); and Zeusler (1926, 1926a, and 1926b).

The various paths which the icebergs are most liable to follow in this region is diagrammatically shown by Figure 97, constructed from

[71] MacMillan, wintering in Labrador 1927–28, reported a remarkable absence of pack ice.
[72] Huntsman (1925, p. 2) found an indraft of 9 miles and an outflow of 8 miles, but the circulation is largely controlled by the winds.

all available data. The separation of the ice stream around the northern buttress of the Grand Bank, as represented by branches a, b, and c, is referred to on page 53. The first two paths head the icebergs into shoal waters, where they meander. The ice closest inshore is dispersed via the Gully along the current lines between Cape Race and

FIGURE 97.—The main paths that icebergs are most liable to follow in the western North Atlantic when advancing southward toward temperate latitudes. Branches *d*, *e*, *f*, *g*, *h*, and *i* are variations in the drift once a berg has embarked along path c. This illustration is purely diagrammatic, but it contains information where icebergs are apt to be found.

the Grand Bank proper. Sometimes bergs are carried through this gully at rapid rates; at other times they remain there nearly stationary while, for short periods prior to a decided change of wind, the current has been known to reverse and such bergs to move back northward. The behavior of the currents along the coast from St.

Johns, around Cape Race, and as far west as the Miquelon Islands, has been studied by Dawson (1906, p. 21), by Huntsman (1930, p. 11), by Matthews (1914, p. 32), by Smith (1924a, p. 131), and by Iselin (1930, p. 3). The outstanding features as affecting the distribution of arctic bergs are (*a*) the great variations and occasional reversals of the current and (*b*) the western and southern limits of the current as marked by the outline of Green Bank and the continental edge. Icebergs have been sighted at various distances to the southwest of Cape Race, but never, to our knowledge, farther west along the coast than Placentia Bay. (See Huntsman, 1930, p. 4.)

Hundreds of bergs ground on the northern part of the Grand Bank, where processes of accumulation and disintegration proceed throughout the entire season. The coldness of these shelf waters preserves the ice longer than if it had continued southward; and so in August bergs may still be on the northern part of the bank long after most of the surrounding localities are clear. As they disintegrate and grow smaller bergs may again be carried off the bank to resume their southward journey or occasionally they may be driven farther in the shoal waters of the bank itself. The central regions of the latter, as a rule, are free from bergs.

Branch *c*, the richest of all the berg streams south of Newfoundland, follows southward in the deep water along the eastern slope of the Grand Bank, as shown on Figure 97 and also by the actual drifts on Figure 102. The behavior of bergs once started along path *c* depends primarily on the manner in which the water masses of this northern discharge meet and conflict with the warm easterly-moving oceanic masses. The axis of the icy current tends to hug the eastern slope of the Grand Bank and curve westward around the "Tail" of the latter but the outer edge of this flow is continually sending out temporary offshoots, offshore. The mixing zone of the currents which carry the ice has the form of an undulating front extending from southwest of the Tail of the Grand Bank to somewhat northeast of Flemish Cap. This demarkation is called the cold wall or temperature wall. Whenever and wherever the boundary shows a salient, there similarly a vortex tends to develop (similar in many respects to the "polar front" of meteorology), but usually the chief point of discharge is near the Tail of the Grand Bank. When the Arctic current swells, the mixing zone between the two waters often moves farther and farther offshore, and in consequence the bergs fan out along the northern edge of the Gulf Stream. During late summer, however, according to our observation, the Labrador current in this region probably dwindles and there are times during the autumn and winter, as proved by the observations of the ice patrol (see Smith 1923, p. 85 and also fig. 99) when the icy current is not traceable at all to the southward of the Grand Bank. At this season, largely reflecting these oscillations, few icebergs drift far south or east of Newfoundland, though an occasional berg may be reported south of Newfoundland in any month of the year. They are at a minimum during November, December, and January, and at a maximum during April, May, and June.

In March there is a marked tendency for the bergs to take the offshore path (branch d, fig. 97) just south of Flemish Cap. Such a dispersal is due to several factors: (*a*) The presence of pack ice farther north along the American coast, which prevents the bergs

from following the main drift; (*b*) a shoulder of warm Atlantic water often deflects the outer part of the cold current with its ice out past Flemish Cap (see fig. 103); and (*c*) the continual northwesterly gales drive off great fields of pack ice and whatever bergs are entangled therein.

Any retreat of the warm salty inthrust between Flemish Cap and the bank, any change in atmospheric circulation, or the dissipation of the pack ice may initiate the file of icebergs down the east

GREENLAND BERGS IN TROPICAL WATERS

FIGURE 98.—While the crew of the ice patrol ship dive into tropical waters of 72° F., Greenland icebergs float in the offing. It has been found that the bergs which enter the warm confines of the Gulf Stream, south of Newfoundland, seldom survive more than a week or ten days. (Official photograph, international ice patrol.)

side of the Grand Bank. The first bergs reach the Tail of the Grand Bank early in April. These are believed to be the remnants of the previous season that have survived the summer's heat in high latitudes and have remained frozen in the pack ice along the American shore over the winter and started south again in the spring. Bergs of this group usually show much disintegration, therefore they do not as a rule survive long in North Atlantic waters. The pack ice which has held them in the far north over

winter, drifts south faster in spring than do the bergs, hence the latter arrive as the fields and floes are disappearing.

Upon approaching the Tail of the Grand Bank the bergs either wheel abruptly to the eastward, to drift northward along the northern edge of the Gulf Stream, or they follow around to the westward. If the bergs ground on the southern part of the bank, as many do, they usually disintegrate there, and thus do not menace navigation farther south.[73] Some bergs, however, which drift westward around the Tail, continue northward beyond the forty-third parallel, where they are usually caught in the large counterclockwise eddy which characterizes the water mass of this region.[74] (See fig. 101.) Still other bergs of this group may be carried directly northwestward, where they eventually ground well in on the Grand Bank, even as far as 120 miles from the Tail, and survive for several weeks before finally disintegrating. Occasionally a berg is even carried as far west as longitude 54° or 55° W. in this way. Shortness of drift, together with a slow rate of travel, probably reflecting the weakness of the northern current, are the outstanding features of the drifts of early April bergs.

Bergs drifting southward toward the Tail attain maximum abundance during May, reaching a normal total of 130 south of Newfoundland. At this, the height of the season, they tend to drift along paths *h* and *i* (fig. 97), provided the cold current is normally developed. The prevailing drift is thence southwesterly to a region bounded by the forty-second and forty-third parallels and the fifty-first and fifty-second meridians, in which vicinity most of the bergs of May swing sharply to the eastward, paralleling the northern edge of the Gulf Stream. As a modification of this type, the ice may set more to the southward and even southwest of the Tail (see figs. 105 and 106), but in most cases it does not drift deeply into the Gulf Stream before being borne off eastward, and finally northeast, where it rapidly melts. However, in a year when arctic influences are weak, or the Gulf Stream abnormally strong, bergs do not reach the Tail but are deflected eastward along paths *d* or *e*, Figure 97. (See also figs 100, 101, 103, and 104.)

June witnesses a decline from the season's maximum iceberg crop. The normal number is 68 bergs south of Newfoundland. The first week of this month may not show any apparent slacking in the stream of bergs, but by the latter part of June, as a rule, the drift of the ice past the Tail of the Grand Bank noticeably diminishes, due to any one or all of the following causes, the relative importance of each still remaining to be determined:

(*a*) The encroachment of warm oceanic water northward in the surface layers toward the Tail of the Grand Bank.

(*b*) The westward encroachment of the inner side of the Gulf Stream toward the eastern slope of the Grand Bank.

(*c*) The withdrawal of the pack ice along the coasts and shelves to the northward allowing the bergs to work inshore and strand.

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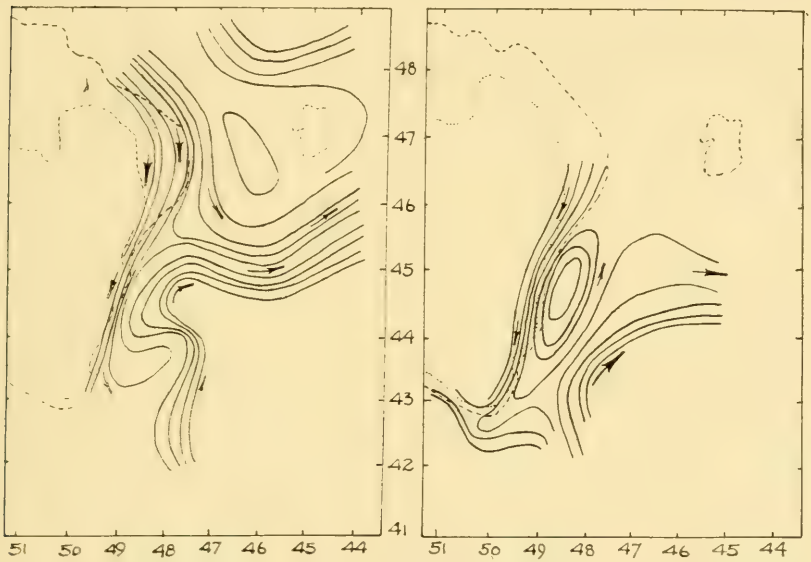
[73] During the ice season of 1922 we saw several bergs float again after several days delay in on the southwest slope of the bank, so to resume their journey southward. (Smith, 1923, p. 58.)

[74] In 1921 (see Smith, 1922, Chart II) the large elliptical track taken by an iceberg off the western slope of the Grand Bank gave the first inkling that such phenomena prevailed in the current system. But several times since then (see Smith, 1927, p. 86, and 1927b, p. 70) the dynamic topographic maps of the sea surface have proved conclusively the existence of such a cyclonic depression.

(*d*) The seasonal cessation of the prevailing northwesterly winds and the consequent removal of this frictional effect to build up a slope current along the North American coast to the northward of the Grand Bank.

The first advanced information on the impending failure in the supply of ice comes from an increase in the number of bergs which begin to set down past St. Johns close in along the Newfoundland coast. Icebergs also about this time collect in the greatest numbers on the northern slopes of the Grand Bank. The gradual swing in the axis of the berg stream from out near Flemish Cap in March to in close to Cape Race in June and July has been likened to the sliding of grain to a bin, as the chute is slowly deflected the supply



CHARACTERISTIC TYPES OF CIRCULATION EAST SIDE OF THE GRAND BANK

FIGURE 99.—The ice patrol dynamic topographic maps have disclosed two characteristic forms of circulation which often prevail on the east side of the Grand Bank. The one on the left shows how the Gulf Stream presses in toward the eastern slope of the Grand Bank, blocking off the bergs, and guaranteeing the safety of the United States-Europe lane routes. The map on the right shows a cyclonic depression in which bergs have often been observed.

is likewise cut off. The fact that few or no icebergs drift southward past the Grand Bank and Newfoundland during the balance of the year, August to December, seems in no way due to the lack of icebergs in the north where they are continually calving from the ice cap and drifting out into Davis Strait and Baffin Bay.

July witnesses a still further slackening in the number of bergs south of Newfoundland; the normal number by months is: July, 25; August, 13; September, 9; October, 4; November, 3; and December, 2. The bergs that do succeed in drifting past Newfoundland are found hugging closely to the edge of the shelf and working inshore to the shallow waters, where the ice usually breaks up and disappears. Bergs in the area south of Newfoundland during the summer dissipate very rapidly, due to the warmth of air and water.

Many direct observations on the rate of movement of icebergs south of Newfoundland have been made by the ice patrol, and, supported by numerous current surveys run simultaneously, furnish us with accurate information. Any student interested in examining this matter in detail is referred to the file of annual reports of the patrol. (See p. 150.) The records may be summarized as follows: During March the average rate is 0 to 7 miles per day, and 10 to 17 miles per day from April to June around the southern parts of the bank. The bergs on the northern edge of the Gulf Stream move fastest of all, i. e., at rates of 19 to 36 miles per day. In late summer the ice drifts more slowly, but at all seasons the rates are subject to considerable variations from week to week or even from day to day. The velocities of ocean currents, especially at this junction region, are constantly changing, changes reflected by the drifts of the icebergs. For example, on one occasion a group of bergs on the east side of the Grand Bank just north of the Tail suddenly increased their rate of drift overnight from 7 miles to 28 miles per day. (Smith, 1924, p. 76.) If the current can accelerate suddenly to such a high rate off the Grand Bank, may not such events occur equally along the Labrador shelf?

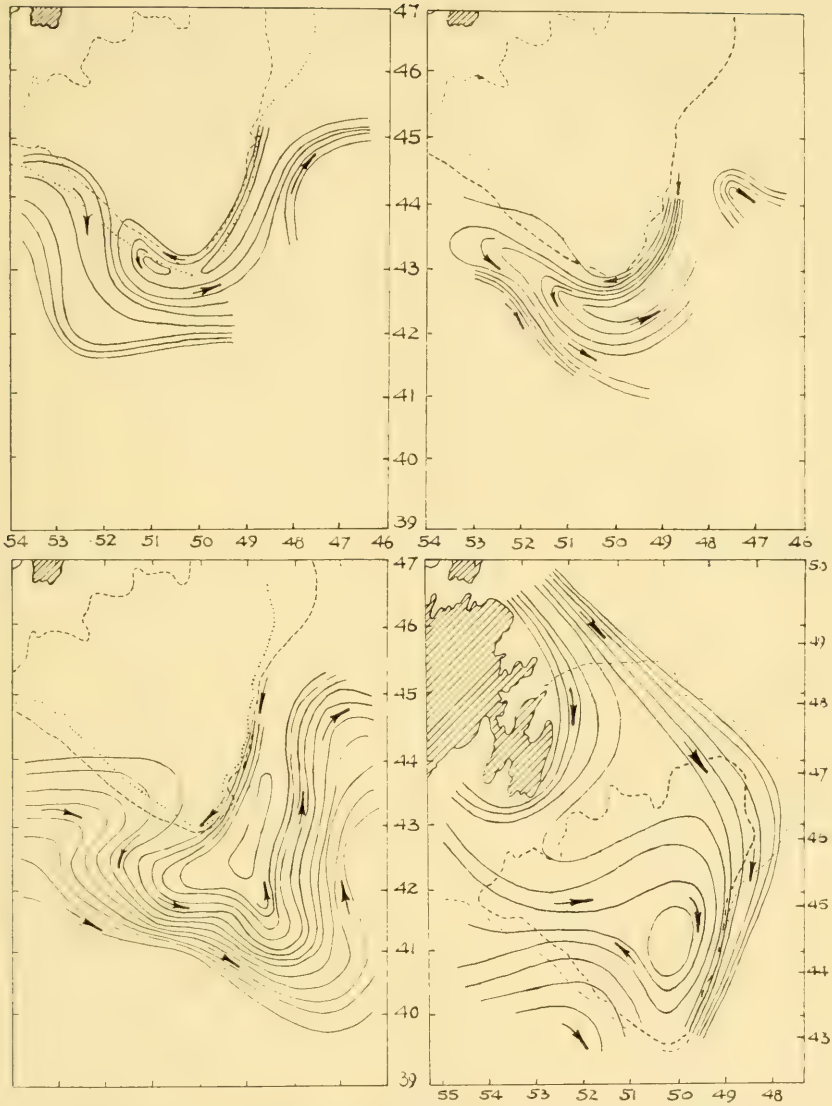
The ice patrol's investigations have revealed the following features in the circulation south of Newfoundland affecting the behavior of icebergs:

(*a*) An elliptical depression in the sea surface makes a slow cyclonic vortex in the surface layers off the southwest slope of the Grand Bank west of the Tail. This "low," covering an area of over 2,000 square miles, was first discovered by chance in 1921 by following an iceberg as it made the circuit. 1926 the "low," apparently similar in character to an atmosphere cyclonic depression, was accurately charted by several successive dynamic surveys (see Smith, 1927, pp. 109, 112, and 115) prevailing throughout the season. It was present also in April, 1927, but disappeared early that May. Bergs are often carried northwestward along the slope in the northern semicircle of this eddy and sometimes a complete circuit of 365° in its periphery. (See fig. 101.)

(*b*) Sometimes the counterclockwise eddy is wanting; displaced by the Gulf Stream encroaching toward the Tail of the bank. An earlier series of current maps (Smith 1927b) shows the development of such a phenomenon in the first week in May, 1927. As long as the Gulf Stream maintains this position ice is blocked from pursuing a course west of the Tail. (See lower right-hand sketch, fig. 101, also overlay figs. 105 and 106.)

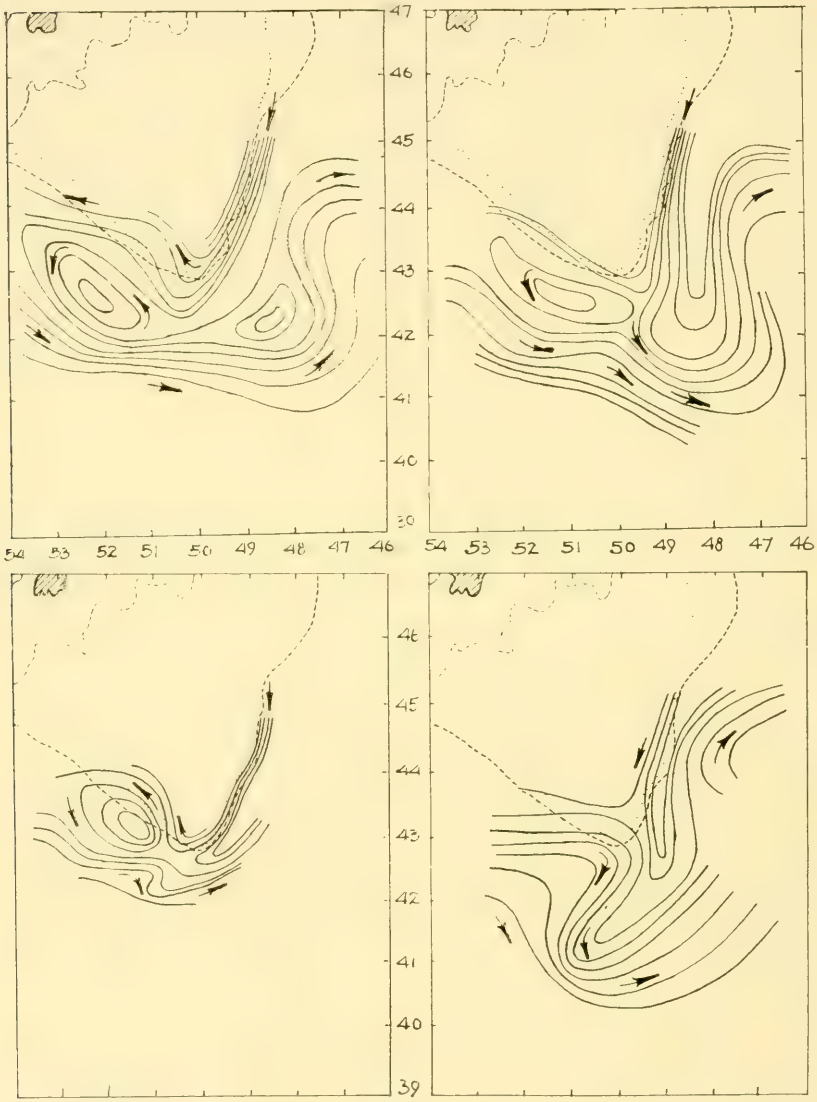
(*c*) The cold and warm currents after meeting off the Tail of the Grand Bank proceed southeastward in a parallel set to the vicinity of latitude 41°, longitude 47°, where the streams bend sharply to the left. The existence of this cul-de-sac of the current, and occasionally for bergs, 150 miles southeast of the Tail shows evidently that the drift is affected by the bottom configuration, even to as great a depth as 4,000 meters. (See Smith 1922, Chart H and lower left-hand sketch, fig. 100.)

(*d*) The ice patrol suspected for several years that the Gulf Stream after turning sharply northward on the forty-seventh meridian pressed in toward the bank up the submarine embayment between



CHARACTERISTIC TYPES OF CIRCULATION AROUND THE GRAND BANK

FIGURE 100.—The two upper maps show the Labrador current hugging the Grand Bank slopes. The type on the left often prevails in early season, while the one on the right is associated with a stronger cold current and a relatively weak Gulf Stream. The lower left-hand sketch reveals a Labrador current developed to maximum; a condition often found in mid ice season. The last map shows the primary circulation which prevails over the Grand Bank. The various systems of circulation as shown above have been obtained from many ice patrol surveys in accordance with the Bjerknes's method of hydrodynamics. Iceberg tracks have also been found to agree with the stream lines of the currents.



CHARACTERISTIC TYPES OF CIRCULATION AT JUNCTION OF LABRADOR CURRENT AND GULF STREAM

FIGURE 101.—The two upper maps and the lower left one feature a cyclonic vortex in the circulation at the Tail of the Grand Bank, where the Labrador current meets the Gulf Stream. The lower right-hand map shows the Gulf Stream crowding well in over the southwest slope of the Grand Bank, while a strong Labrador current penetrates deeply toward the south. All these types of circulation have been mapped by the ice patrol, and icebergs also have many times been observed in the control of these systems.

the forty-fourth and forty-fifth parallels. The current surveys of 1926 and 1927 (see Smith 1927 and 1927b) proved such is the case when the cold current is weak or the warm abnormally voluminous. In some years this phenomenon is more pronounced than in others, but whenever well developed it definitely blocks the path of the bergs from reaching the Tail of the bank, and deflects them out southward and to the eastward. An excellent illustration of this is to be seen in the charts for 1926. (Smith 1927b, pp. 86 and 90, and fig 99, p. 155 of this paper.) For several years past this inshore development of the Gulf Stream on the east side of the bank in June and July, insuring safety to the steamship tracks, has been thought to warrant the discontinuance of the ice patrol for the year.

(*e*) The change in alternation in direction of flow between the warm and cold water masses takes place about 25 miles shoreward from the "cold wall" or temperature wall. (See fig. 122, p. 204.) It is for this reason that icebergs are seldom sighted offshore of the cold-water area south of Newfoundland and that they often drift along parallel to the Gulf Stream, though lying many miles within the offshore boundary of the cold, mixed waters.

Icebergs around Newfoundland have been trailed by the ice patrol as they drifted many hundreds of miles in the current; in fact the compilation of individual berg drifts in the area south of the Tail of the Grand Bank constitute the most positive evidence to refute the old belief that the Labrador current flowed southward along the United States coast. (See figs. 102 and 107.) The first attempts to follow the movements of an iceberg near the Grand Bank were by Johnston (1913, p. 23) when the *Seneca* kept in touch with two bergs for over a period of five weeks, finding that they drifted in a large cyclonic vortex off the east side of the Grand Bank. The next instance was in 1915 when a berg was followed for 17 days while it drifted 195 miles in a wide semicircle south of the Tail, first to the westward at the rate of 12 miles per day and then eastward in the Gulf stream at 24 miles per day. In 1921 on the ice patrol we began a policy of tracking the bergs whenever possible, and the composite map of drifts (fig. 102) so compiled has now become very instructive. Some of these follow: April 11 to May 12, 1921, a large berg drifted from the northeastern edge of the Grand Bank to a point 90 miles south-southwest of the Tail at an average rate of 15 miles per day; thence it was carried eastward on the Gulf Stream for 10 days at the rate of 28 miles per day. March 19 to April 15, 1922, a berg drifted westward past the Tail of the Grand Bank, then south and east, a total distance of 315 miles at an average velocity of 12 miles per day. May 2 to 20 three bergs set southwestward past the Tail of the Grand Bank in a large anticlockwise eddy at the speed of 10 miles per day and a total distance of 175 miles. March 16 to April 11, 1924, a berg on the northeastern part of the Grand Bank drifted south into the Gulf Stream a total distance of 450 miles. The rate of drift was slow to the forty-third parallel, but south of the Grand Bank it increased to 24 miles per day. May 19 to June 30, 1925, a berg was followed over a distance of 492 miles as it drifted from the forty-fifth parallel southward along the east side of the bank to the Tail and thence southwesterly until turned easterly by the Gulf Stream. May 26 to June 2, 1928, a berg was traced 265 miles in a semicircular

path southeast of the Grand Bank. A map, such as Figure 102, showing the compiled drifts of the foregoing bergs, and many more besides furnishes the best information available as to their general paths in the currents. Its most instructive feature is its demonstration that the heaviest ice stream is past the Tail of the Grand Bank to the vicinity of latitude 42° 30′, longitude 51° 30′, where the bergs

THE DRIFT TRACKS OF ICEBERGS, 1900-1930

FIGURE 102.—The courses that icebergs have followed south of Newfoundland is a record compiled by the international ice patrol, as it has tracked bergs for distances of 400 miles and more in the ocean currents.

tend to turn abruptly to the left, and thereafter follow an easterly course until they perish.

We have on the ice patrol also collected over a period of several years the physical data needed for the dynamic computation of the oceanic circulation in the region of the Grand Bank in accordance with Bjerknes's theory of free motion. (See Smith (1926).) Dur-

ing the years 1926 and 1927 the circulation of the water masses in the iceberg region south of Newfoundland was kept constantly under surveillance by means of frequent current surveys covering areas of several thousand square miles. The general results have been discussed in earlier publications (Smith 1927 and 1927b), to which are referred any students interested in the details of the circulation in special regions.

In general the agreement between the berg tracks that have actually been followed (figs 103, 104,105, and 106) and the gradient currents for the same periods, respectively (overlays for figs. 103, 104, 105, and 106), is so close as to indicate that dynamic projections of this sort may be used as a basis for predicting the tracks that individual icebergs are most likely to follow (see p. 176).

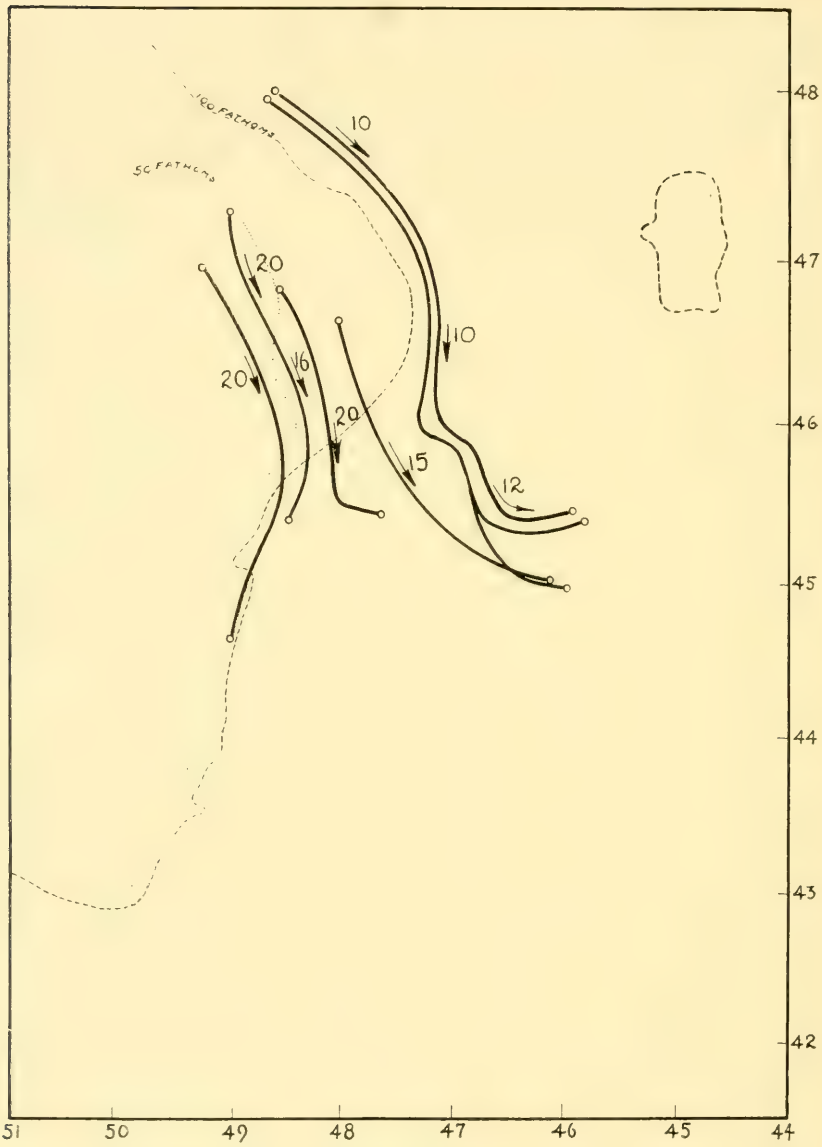
Occasionally icebergs survive relatively long periods, even when floating in Atlantic Ocean water of high temperature, and they may then make phenomenal journeys. Thus icebergs have been sighted near the Azores, near the British Isles, and even near Bermuda, such events usually taking place late in the season and during bad ice years such as 1890 and 1912. These drifts, however, do not indicate a direct extension of the Labrador current into low latitudes but simply that the bergs in question have been caught up in oceanic vortices that are continually forming over the Atlantic basin in which the ice is borne southward instead of following the normal drift.[75] The emphasis that has often been laid on bergs drifting exceptionally far southward is apt to give an exaggerated impression of the frequency of such events. As a matter of fact, it is unusual for a berg to drift south of the fortieth parallel of latitude in the western North Atlantic, the records for the past 20 years showing only one such occurrence every one to three years.

The Deutsche Seewarte (1902, tafel 3) published a map showing the positions in which ships have reported ice far south in the Atlantic. During May and June of the year 1890 some bergs attained the thirty-seventh meridian between latitude 44° 30′ and 46° N. During the period 1904–1913 ice was sighted at least once south to latitude 37° 50′ north and east to longitude 38°; with one exceptional journey to the thirtieth parallel. In May, 1905, a few bergs penetrated the Atlantic to latitude 39° 07′ directly south of the Grand Bank. In 1912, the year the *Titanic* was lost, a berg reached latitude 39°, longitude 47°.

Hennessy (1929, p. 84) has published a list of extraordinary berg drifts which shows that only 24 bergs during the past 20 years have been sighted south of the fortieth parallel.

Some of the positions in which ice has been reported seem hardly credible—due to errors in observation or in transmitting radio reports—had they not been verified. One of the most astonishing incidents is the report made by the British steamer *Baxtergate* and verified by the United States Hydrographic Office that on June 5, 1926, she passed a large piece of ice 30 feet long, 15 feet wide, and 3 feet above water in the latitude 30° 20′ N., longitude 62° 32′ W., near Bermuda. Such an occurrence is the more mysterious when it is

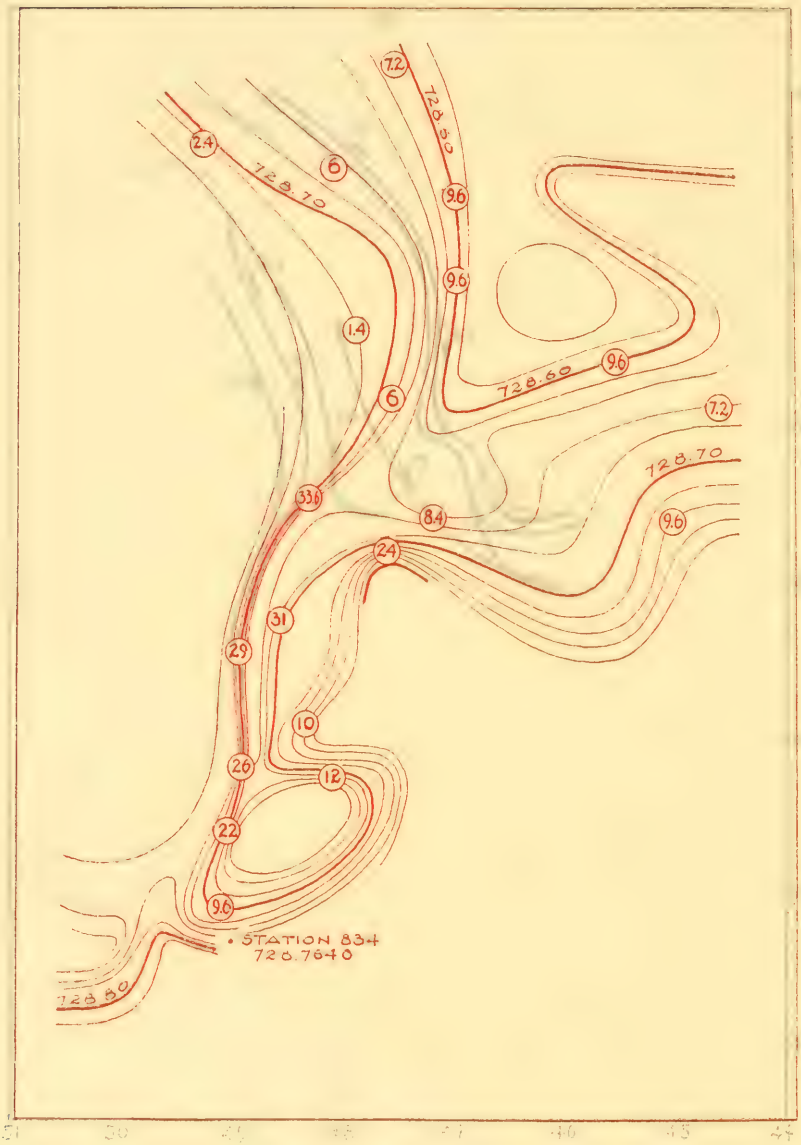
[75] It has already been pointed out that the outlet for icebergs departing on extrasoutherly drifts is noticeably confined between meridians 46 and 50, almost directly south of the Grand Bank. (Smith, 1927, p. 68.)



ICEBERG DRIFTS COMPARED WITH OCEAN CURRENTS

FIGURE 103.—The observed drifts of six icebergs, June 9–25, 1927, on the northeastern side of the Grand Bank compared with the direction and velocity of the slope currents contemporarily determined by means of Bjerknes's formulæ of hydrodynamics. The numbers on both the base map and the overlay indicate the rate of movement of the ice and the water, respectively, in miles per day.

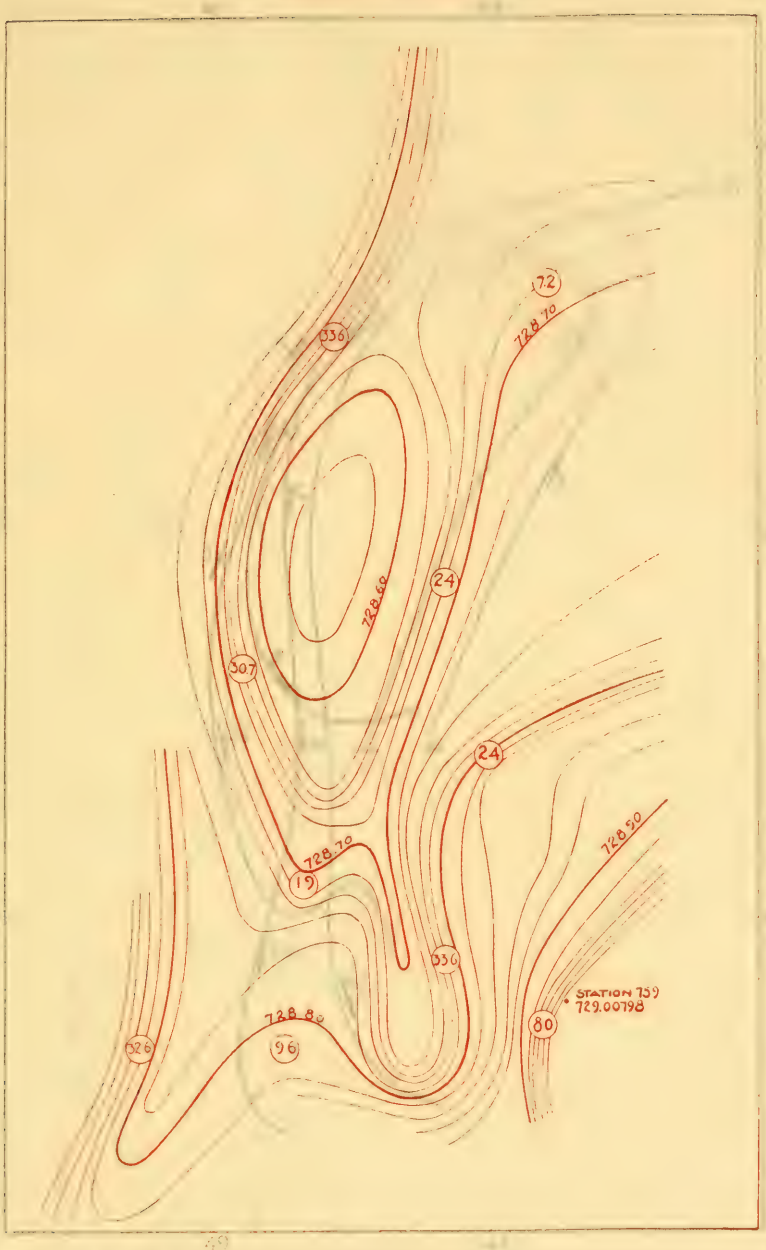
FIGURE 103 overlay



CELTAS USED AS COMPARED WITH OCEAN CURRENTS

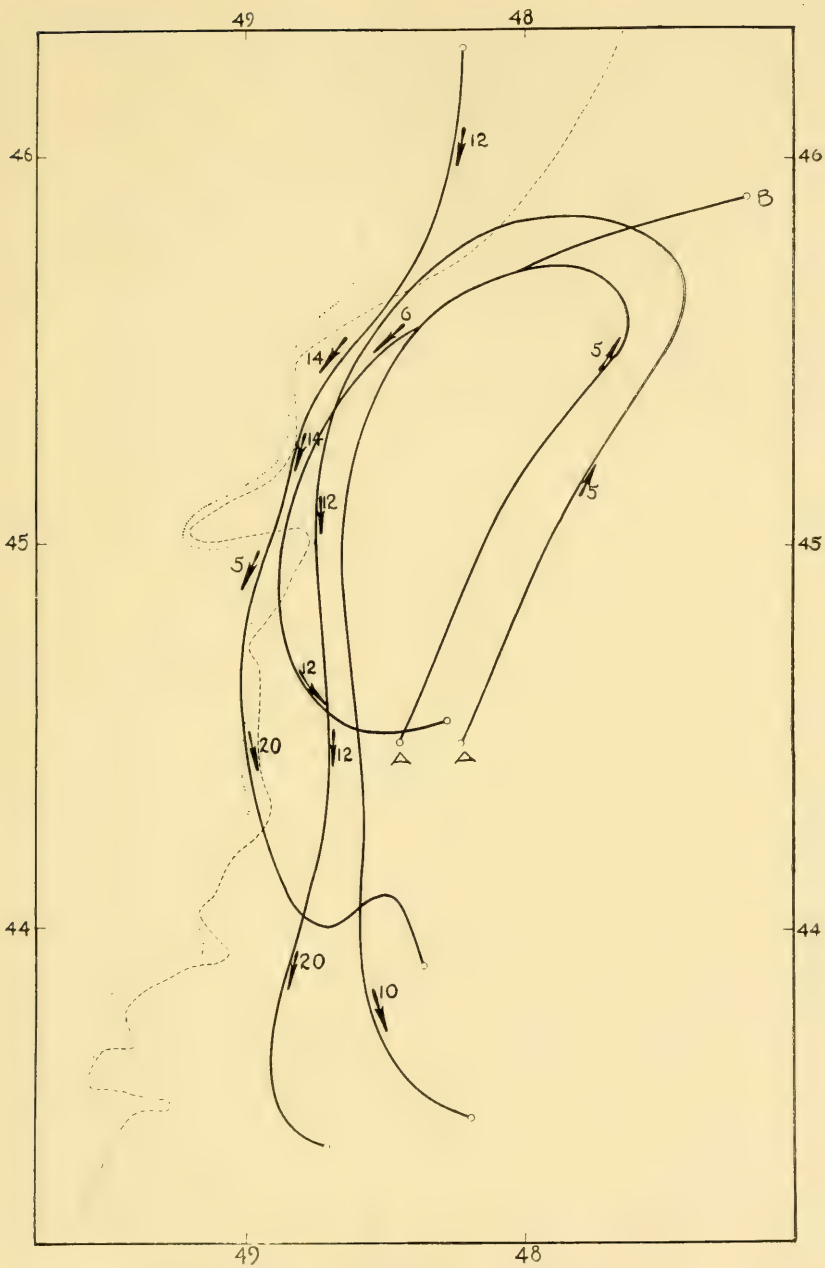
FIGURE 103.—The ab[illegible] is based on [illegible] June 8-25, 1927, on the northern side of the Grant Bank [illegible] with the direction and velocity of the surface current [illegible] Bjerknes's formula [illegible] tide [illegible]. The numbers on both the [illegible] and the overlay indicate the [illegible] of movement [illegible] the water [illegible] in miles per day.

FIGURE 104 overlay



IN FIELD DRIFTS COMPARED WITH OCEAN CURRENTS

[illegible]



ICEBERG DRIFTS COMPARED WITH OCEAN CURRENTS

FIGURE 104.—The observed drift of an iceberg May 27 to June 8, 1927, on the east side of the Grand Bank, compared with the direction and velocity of the slope currents contemporarily determined by means of Bjerknes's formulæ of hydrodynamics. AA refers to two bergs April 13–16, 1913, which drifted in a similar system of circulation. B refers to a similar drift of a berg April 30 to May 23, 1925.

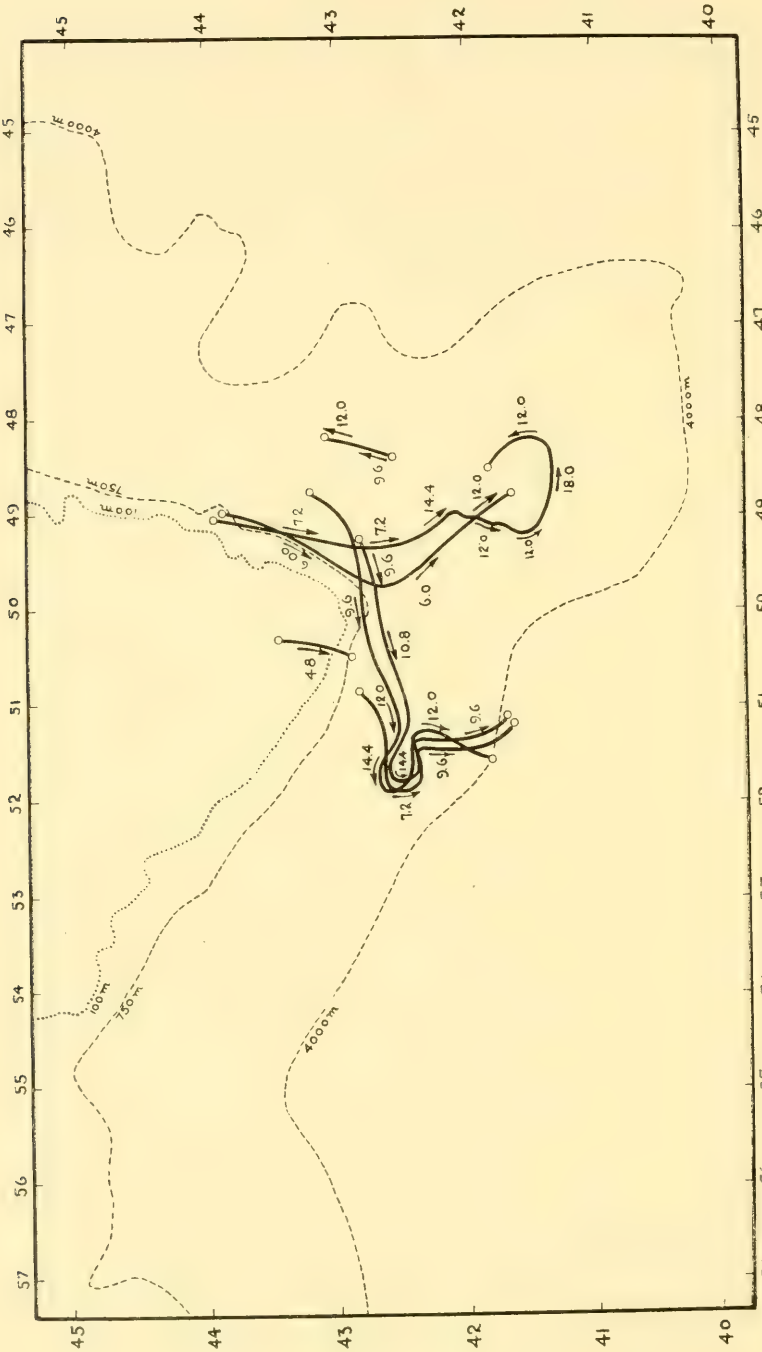


FIGURE 105.—The observed drifts of seven bergs, May 1–27, 1922, near the Tail of the Grand Bank compared with the direction and velocity of the slope currents contemporarily determined by means of Berkness's formulæ of hydrodynamics. The numbers on both the base map and the overlay indicate the rate of movement of the ice and the water, respectively, in miles per day.

Isohypse 100 meters

FIGURE 105 overlay

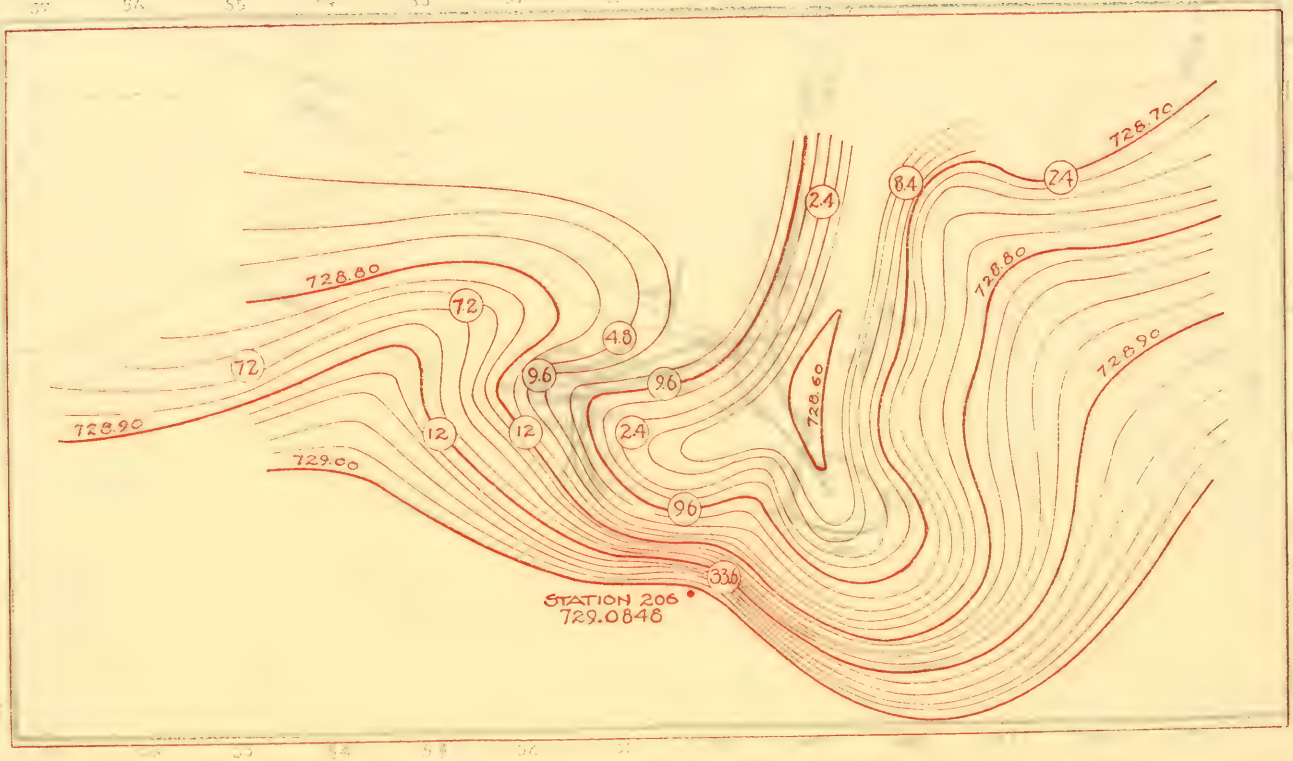


PLATE 104.—The observed drifts of seven floats, Mar. 1-24, 1922, near the S. E. tip of the Grand Bank compared with the direction and velocity of the tide currents contemporaneously determined by means of Bjerknes's crude of tyd polyrondes. The numbers on both the base map and the overlay indicate the rate of movement of the ice and the water, respectively, in miles per day.

FIGURE 106 overlay

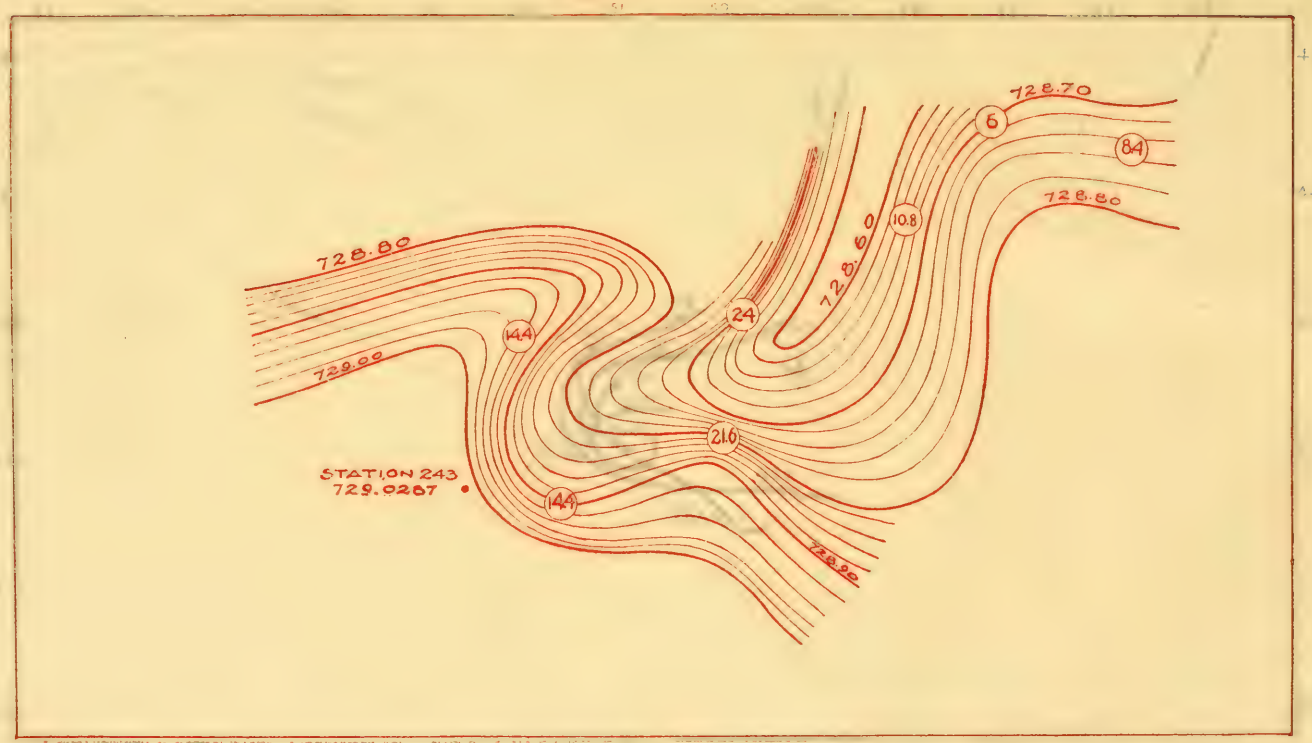


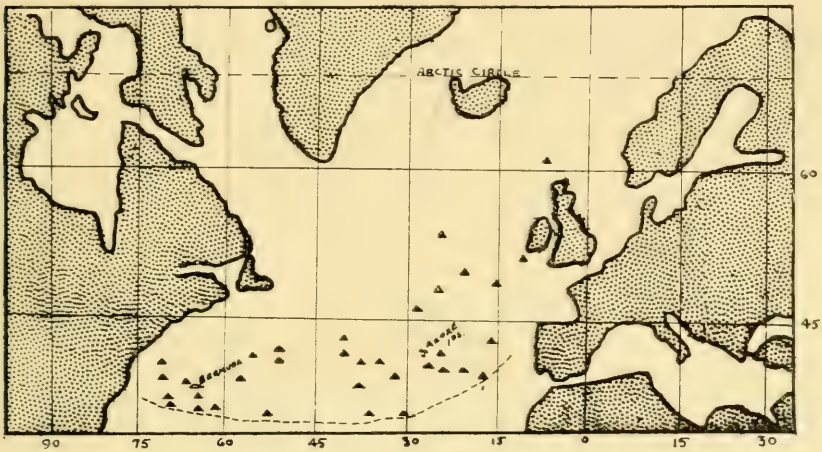
PLATE 106.—The observed drift of five icebergs, June 14 to July 3, 1922, near the Tail of the Grand Bank, compared with the direction and velocity of the slope currents, contemporarily determined by means of Bjerknes's formulae of hydrodynamics. The numbers on both the base map and the overlay indicate the rate of movement of the ice and the water, respectively, in miles per day.

THIRTY YEARS OF ICEBERGS

FIGURE 107.—The positions in which icebergs have been sighted around the Grand Bank south of Newfoundland, 1900–1930. The localities where icebergs are densest are along the eastern edge of the Grand Bank, and in an east-northeasterly direction from Cape Race, Newfoundland. This distribution agrees well with the main branches of the cold current which brings the ice southward. Note the southwesternmost position which icebergs have attained, latitude 41° N., longitude 55° W., an excellent indication of the nonexistence of a continuation of the Labrador current down the east coast of the United States. Note also the definite tendency of a few bergs in the 30-year period to drift far south between the meridians 49 and 46, 50° W.

realized that no other ice was sighted that year within a thousand miles of the *Baxtergate's* position, although the ice patrol had been maintaining a vigilant guard upstream, off Newfoundland, the entire spring.

An investigation of the records of the United States Hydrographic Office and of the international ice patrol for the period 1900 to 1926 regarding the southward distribution of icebergs in the western North Atlantic shows that a total of about 386 drift south of Newfoundland during a normal year, and that 51 of these on the average (about 15 per cent) are carried south of the Grand Bank. The distribution in the three main paths, viz, around Cape Race, down the



PHENOMENAL ICEBERG DRIFTS

FIGURE 108.—The above positions were compiled from 21 authentic and verified reports from ships at sea sighting icebergs, 1900–1916. (From Jenkins, 1921.)

east side of the Grand Bank, and southeastward between the bank and Flemish Cap, is 14 per cent, 67 per cent, and 19 per cent, respectively. The distribution by months is as follows:

Normal number of icebergs south of Newfoundland

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
3	10	36	83	130	68	25	13	9	4	3	2

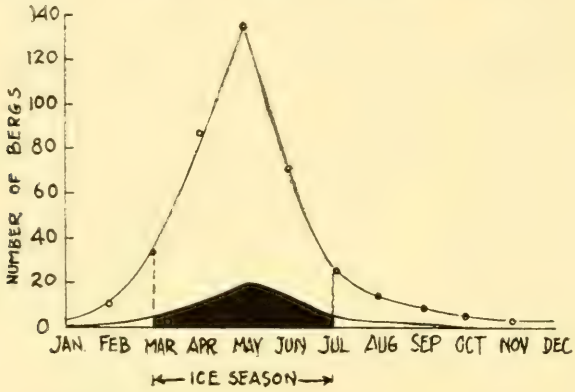
Normal number of icebergs south of Grand Bank

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
0	1	4	9	18	13	3	2	1	0	0	0

These data indicate that the iceberg season off Newfoundland may be said to cover a period of four months, from March 15 to July 15. The bergs decrease in numbers noticeably after the middle of June

and from the middle of July until the following spring the area south of the Grand Bank is practically free. An isolated berg or two may drift southward to the Tail of the Bank, but not often south of the latter, as late as October, after which month bergs are sighted rarely in the latitude of the Grand Bank until the following February.

It is interesting to note that of the fifty-odd icebergs which may be expected to drift south of the Tail during a normal year, only three will ordinarily be carried across the westbound steamship routes which run between Europe and the United States. The bergs which drift south of the lanes located along the latitude of 40° N. do so in April, May, and June: May being the month which is most dangerous to shipping. The total number sighted along the tracks south



THE NORMAL SEASONAL DISTRIBUTION OF ICEBERGS

FIGURE 109.—The upper curve represents the normal monthly number of icebergs south of Newfoundland in the western North Atlantic. The lower curve represents the normal number of bergs south of the Grand Bank. The number of bergs are at a minimum during November, December, and January, and at a maximum during April, May, and June. The ice patrol has interpreted the normal iceberg season as extending from the middle of March to the middle of July.

of the fortieth parallel during the decade 1913 to 1923 was 33, distributed as follows: 3 in April, 25 in May, 5 in June.

METHODS EMPLOYED TO PROTECT TRANS-ATLANTIC SHIPPING FROM THE ICE MENACE

Where the cold northern currents carry Arctic ice on to the North Atlantic lanes of commerce, there it becomes a distinct economic menace to life and property. The protective measures taken every spring against this danger are (*a*) a system of prescribed routes south of the normal ice barrier, and (*b*) a ship patrol to warn vessels of the position of the ice.

The interesting accounts of the early Arctic explorations by the British, the Dutch, and the Scandinavians relate to dangerous situations in which vessels were often surrounded and crushed within the vast fields of pack ice. Thus in the year 1777 the Dutch whaling fleet was unexpectedly caught in the heavy ice floes off east Greenland and 12 vessels were swept away and sunk in the dangerous waters of Denmark Strait. The tragic loss of the famous *Jeannette* and practically all of her crew was due to the great pressure of the

ice-cap where she was beset northwest of Wrangel Island, Siberia. The latter part of the nineteenth century on four different occasions, viz., 1871, 1876, 1888, and 1896, witnessed a series of unexpected "freezes," when the American whaling fleets were crushed in the pack ice east of Point Barrow, Alaska, with a total loss of more than 50 ships. Whaling, fur trading, and scientific explorations have done much to educate the many able seamen of both steam and sail in the art of ice navigation. Bartlett (1928), one of the best known of present-day sailing masters, has discussed some of his rich and varied experiences in northern seas and published a few valuable instructions on the art of ice navigation.

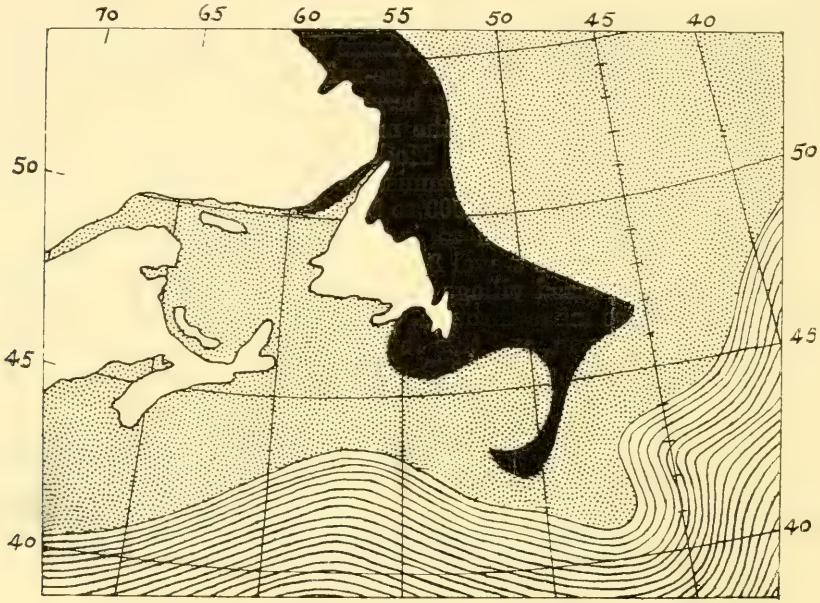
But the state of the ice in the Arctic is far different from such conditions in the more open, yet ice-infested waters of temperate latitudes. The pack ice and the beset ship of the far North gives way in the North Atlantic to the massive iceberg and the sudden impact of disastrous collision. Modern attainments in the art of shipbuilding have placed in commission enormous hulls, aggregating 50,000 tons, costing $5,000,000 to $15,000,000, and driven at railroad speeds of 20 to 25, or more, knots per hour. On board one of these new-day leviathans travel 3,000 to 4,000 people, as many souls as constitute a fair-sized village on shore; 1,500 to 2,000 total individual passages are made through, or past, the ice regions off Newfoundland every season, and this represents approximately $10,000,000,000 of property and 1,000,000 lives that come each year within the shadow of the ice menace.

The icebergs off Newfoundland have for centuries been one of the most dreaded dangers of trans-Atlantic navigators. John Cabot describes sailing past these towering monsters shrouded in Grand Bank fogs on his first voyages to America early in the sixteenth century. Pioneer navigators of the North Atlantic soon learned to determine their latitude approaching western shores by the sharp line of Arctic waters thrust southward along the forty-seventh meridian. (See fig. 110.) Benjamin Franklin through his relative Capt. Peter Folger of Nantucket, pointed out the Gulf Stream and a path to avoid most of the Newfoundland fog and ice.

A perusal of trans-North Atlantic sailing ship disasters impress one with the great number of casualties that befell those vessels which followed the shortest route (great circle course) across the ice longitudes. Strange to relate, however, the risk of collisions between ships, bound on opposite courses, especially in the fog regions south of Newfoundland and Nova Scotia, first elicited more anxiety than did the dangers of the ice. Early in the nineteenth century an unusually appalling disaster about 50 miles east of Cape Race, between the French steamer *Vesta* and the American ship *Arctic* with the loss of 300 lives brought an acute realization that remedial measures must be taken. Lieut. M. F. Maury of the United States Navy was the first to propose seperate lane routes in his Sailing Directions, published in 1855. Another active advocate of separating the east and west bound traffic was Mr. R. B. Forbes, of Boston, Mass. whose proposal to run the westbound lane diagonally across the Grand Bank just south of Cape Race, while the eastbound tracks were to cross about 15 miles north of the Tail of the Grand Bank (latitude 43° at longitude 50° W.) was the one eventually adopted.

Although collisions between ships were materially reduced after the inauguration of prescribed tracks, the dangers incident to navigating without due regard for the menace of icebergs and pack ice remained unmodified and therefore accidents continued to prevail.

It is to the credit of the Cunard Steamship Co. that in 1875 the first real steps were taken to reduce the number of casualties due to drifting Arctic ice. The Cunard Line order its shipmasters to follow lane routes across the Atlantic that were laid south of the zone into which northern ice normally drifted. The westbound lane was run



ARCTIC, MIXED, AND GULF STREAM WATERS

FIGURE 110.—The distribution of the three main classes of water in the region of the Grand Bank south of Newfoundland. The solid black area indicates the position of the coldest water, ca. 34° F., and the most liable retreat of Arctic ice. The wavy, parallel lines mark the region of warmest water, the northern edge of the Gulf Stream, about 60° F., and the least probability of harboring Arctic ice. The stippled area is the intermediate zone of mixed waters, about 45° F., and represents an area liable to contain icebergs. The marked difference in temperature between the cold, icy water on the one hand and the warm oceanic water on the other provides the greatest liability of fog over the mixed and the icy waters, while clear weather prevails over the Gulf Stream.

to the point, latitude 43° N. longitude 50° W.; and the eastbound route was placed 60 miles to the southward.

A few of the other large and more progressive passenger lines followed the policy of the Cunard Co., and as a result the number of accidents due to ice showed an encouraging decrease from the former high rate. They still, however, continued to be of too frequent occurrence, since during the decade 1880 to 1890, according to Rodman (1890), there were no less than 14 vessels lost and over 40 seriously damaged off Newfoundland. For example we note the following:

June 2, 1882, the steamship *Ashdrubal* struck a berg 20 miles south of Cape Race and sunk.

March 15, 1883, the bark *General Birch* was found fast in pack ice in latitude 45° N., longitude 48° 30′ W., with bows stove in and the vessel abandoned and full of water.

January 2, 1884, steamship *Notting Hill* collided with a berg and was so seriously damaged that she had to be abandoned in latitude 46° N., longitude 46° 20′ W.

May 7, 1885, the brig *Annie Christine* struck a berg on the Grand Bank and foundered.

February 17, 1890, the bark *Meteor* spent nine days in a large ice field south of Cape Race. The ice crushed in her bow, opened her seams, and she sank. The crew were rescued in an exhausted condition.

The majority of the trans-Atlantic traffic, however, continued to follow courses through the ice longitudes (forty-five to fifty-two meridians) despite the added safety that it had brought to the Cunard Line and a few others navigating farther to the south. The increase in distance caused by following the safer route of about 100 miles has always been the incentive for the navigator to "cut the corner."

Noting the insurance of safety inherent to circumnavigating the ice regions, the United States Hydrographic Office in 1891 urged the several principal steamship companies to meet and discuss further lane route recommendations. In 1898 the trans-Atlantic Track Conference was formed with all of the established passenger companies agreeing to the present system of prescribed tracks (see fig. 112, p. 175);

Lane route A runs between the United States and Europe. It is effective only during ice seasons when bergs are numerous south of the Grand Bank. The eastbound route crosses meridian 47 at latitude 39° 30′ N.; the westbound route crosses meridian 47 at latitude 40° 30′ N.

Lane route B runs between the United States and Europe. It is effective normally February 15 to August 1, unless severe ice conditions require lane route A. The eastbound route crosses meridian 47 at latitude 41° 30′ N.; the westbound route crosses meridian 47 at latitude 40° 30′ N.

Lane route C runs between the United States and Europe. It is effective normally August 1 to February 15. The eastbound route crosses meridian 50 at latitude 42° N.; the westbound route crosses meridian 50 at latitude 43° N.

Lane route D runs between Canada and Europe. It is effective normally February 15 to April 10. The eastbound route crosses meridian 50 at latitude 42° N.; the westbound route crosses meridian 50 at latitude 43° N.

Lane route E runs between Canada and Europe. It is effective normally April 10 to May 15, or until the Cape Race tracks are clear of ice. The eastbound route crosses meridian 50 at latitude 45° 25′ N.; the westbound route crosses meridian 50 at latitude 45° 55′ N.

Lane route F runs between Canada and Europe. It is effective normally May 15 to July 1, or until the Strait of Belle Isle is navigable. After the strait becomes closed vessels may revert to route D. The eastbound route crosses the meridian of Cape Race 25 miles south of the latter; the westbound route crosses the meridian of Cape Race 10 miles south of the latter.

Lane route G runs between Canada and Europe via the Strait of Belle Isle. It is effective normally from the first week in July until the first week in December.

The prescribed tracks between the United States and Europe are shifted southward during the spring on the advice of various steamship masters, the international ice patrol, or the United States Hydrographic Office, and in accordance with the severity of the ice season. It is worth noting that the extra-southerly tracks (lane route A), which are employed in severe ice years only, lengthen the voyage about 50 miles more than lane route B and about 200 miles more than the great circle course close under Cape Race, Newfoundland. Practically all of the passenger ships now observe the letter of the agreement of prescribed tracks and also many of the freight vessels, especially those belonging to well-known lines. Many miscellaneous freighters, "tramps," however, persist even to-day in cutting through the ice regions unnecessarily and thereby they invite disaster. Shipping on the routes between Canada and Europe, during the ice season, finds it impossible, of course, to avoid the ice zone, and consequently those vessels run a greater danger despite the exercise of much care by their officers.

It has been remarked that if icebergs and pack ice always remained within their normal limits, then nothing further would be necessary to insure sufficient safety to most of the ships, but unfortunately such is not the case. There have been, however, several accidents to trans-Atlantic ships since the prescribed tracks have been in force. Some of them are:

May 25, 1919, the passenger steamship *Cassandra* in latitude 47° 31′ N., longitude 51° 22′ W., northeast of Cape Race in a dense fog, struck a berg filling her forward compartments with water, but was able to make St. Johns, Newfoundland, unassisted.

April 8, 1921, the steamer *Castle Point* was beset in pack ice on the northern part of the Grand Bank. It was necessary for the ice-patrol ship to go to her assistance and tow her into open water.

June 23, 1925, the steamer *Saugus* collided with a valleytype of berg, running her forefoot out on the base platform, where she remained for some hours. Fortunately the damage extended only to straining the vessel considerably, and there was no loss of life.

In the spring of 1928 the passenger steamship *Montrose* struck an iceberg while navigating in the vicinity of the northern part of the Grand Bank. Two of the crew were killed by many tons of ice which fell on the deck, and the forecastle head of the ship was badly battered in.

July 20, 1929, the steamer *Vimera* collided with a berg near the Tail of the Grand Bank, badly crushing in her bow plates and totally disabling her propeller.

The most appalling of ice disasters occurred on the night of April 14–15, 1912, when the White Star liner *Titanic*, on her maiden voyage to the United States, struck a berg in latitude 41° 46′, longitude 50° 14′ (off the Tail of the Grand Bank), and sank with the loss of 1,513 lives. The toll of 300 souls early in the nineteenth century preceded the adoption of trans-Atlantic lane routes; one of the worst marine catastrophes of the twentieth century resulted in the establishment of the international ice patrol.

THE GENERAL CIRCULATION IN THE NORTHWESTERN NORTH ATLANTIC

FIGURE 111.—The general system of circulation which prevails in the northwestern North Atlantic, from the northern end of Baffin Bay to the Gulf Stream south of Newfoundland. This map is based upon the separate dynamic surveys of the *Godthaab*, the *Marion*, and the international ice patrol. The principal trans Atlantic lane routes are also shown crossing the southern end of the ice-bearing ocean currents.

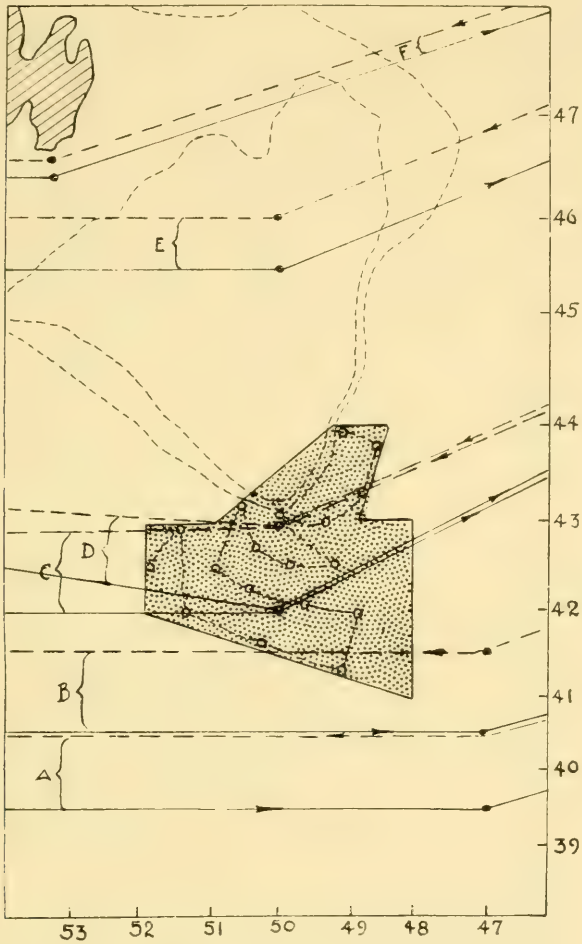
In the remaining dangerous months of the ice season succeeding the *Titanic* disaster, the United States Navy patrolled the waters south of Newfoundland, and in the following year the United States Coast Guard took up the duty. In the autumn of 1913 a Convention for the Safety of Life at Sea was called in London, where the principal maritime nations of the world provided, among other things, to establish a continuous ship patrol of the eastern, southern, and western limits of drifting Arctic ice during the most dangerous part of the year. The continuous guard is inaugurated with the first appearance of icebergs around the Grand Bank, usually the latter part of March, and the service is discontinued after the dangers largely disappear, usually about the 1st of July. Twice daily a detailed broadcast by radio informs all approaching ships the up-to-the-hour position and behavior of the ice. Extensive scientific studies are conducted which lead to a better treatment of the ice problems and a greater safety guaranteed to life and property on the North Atlantic. The *Marion* expedition to Greenland in 1928 is an example of the work being carried out by the international ice patrol. Under the London convention the United States Government assumes the operation of the patrol, the expenses being annually divided among the nations a party to the pact, each paying a quota based on its proportional ocean tonnage. The Coast Guard is the service in the United States which supplies the ships and men, and the United States Hydrographic Office cooperates to give publicity and its aid to the project.

The ice patrol during the first 15 years of its service has established an excellent record. Not a single life has been lost on the United States-Europe tracks since the patrol's establishment, and no vessels have been sunk as a result of collision with icebergs. Although the entire ice regions embraced by the lane routes are the province of the ice patrol the area is much too large for one vessel alone to properly patrol it. The United States-Europe lane routes being the most populous are therefore the ones most carefully guarded, and the more northerly routes between Canada and Europe (the most hazardous), can not unfortunately receive such close attention. A perusal of the effective dates of the various lane routes, moreover (p. 171), shows that prior to April 10, all trans-Atlantic steamship tracks cross the ice regions south of the Tail of the Grand Bank, but after that date the Canada-Europe routes separate from the United States-Europe ones, so that during the ice season when bergs menace navigation in greatest numbers, there are two paths of ocean travel separated by a distance of several hundred miles. Almost every year witnesses the damage to some craft on the Cape Race routes, and occasionally a loss of life. If the international ice patrol system is not extended to the northern steamship lane routes another serious marine disaster similar to that of the *Titanic* is liable to occur.

In 1929 the second Convention for the Safety of Life at Sea met in London and provided for a third, additional vessel for the international ice patrol.

Each year the United States Coast Guard publishes a report of the season's events. A description of the technical methods and the problems of the ice patrol are discussed and described. These bulletins clearly show that the work has been increasing rapidly in

recent years. The fact that the volume of radio traffic handled by the patrol in the last five years alone has increased 300 per cent, is eloquent testimony of the use of this ice service by North Atlantic shipping.



LANE ROUTES AND THE CRITICAL ICEBERG AREA

FIGURE 112.—The location of the principal trans-Atlantic lane routes which are explained in the text, page 171. The shaded portion of 20,000 square miles represents the iceberg area which receives the international ice patrol's most vigilant attention. Bergs in this area are potential menaces, liable at any time to drift southward across the paths of the trans-Atlantic traffic. The ice patrol during the past few years has attempted to keep on board an up-to-date current chart of this area. A satisfactory map, according to Bjerknes's method, can be constructed from a total of 18 points of observation, as shown above.

One of the most interesting scientific aids which the patrol is using with considerable success is that of current mapping according to the Bjerknes theory of dynamic oceanography. The patrol by keeping an up-to-date current map of a 20,000 square mile area south of the Grand Bank, at the junction of the Labrador current and the

Gulf Stream (immediately north of the United States-Europe tracks), is able to intelligently follow the devious movements of icebergs in this critical area and thereby more effectively protects approaching steamships. The critical area shown on Figure 112, for the duration of the ice season, is the one always under the ice patrol's most watchful guard. When the southern terminus of the icebergs has been thoroughly searched by the ice patrol, or during a light ice year, then, and only then, the patrol ship is free to scout northward to find the next southernmost threatening ice.

The current maps obtained by dynamic surveys have been found to remain reliable for a period of 7 to 15 days around the Grand Bank. Minor fluctuations of short duration have often been observed, however, especially along the boundary of mixed waters and Gulf Stream. Such swirls or vortices appear to be secondary superficial tongues 5 to 10 miles in width and sometimes several times that in length and the tracks of icebergs, especially the small shallow-draft ones, are often modified by such departures. In this connection, Ricketts (1930, p. 94) calls attention to the value of surface isotherms in predicting the behavior of icebergs.

Summarizing, the two methods now employed to safeguard North Atlantic commerce from the great dangers incident to Arctic ice are: (*a*) Prescribed tracks laid south of the normal ice zone; and (*b*) a continuous ship patrol south of Newfoundland during the ice season.

Seasonal Character of the Ice Invasions to the North Atlantic

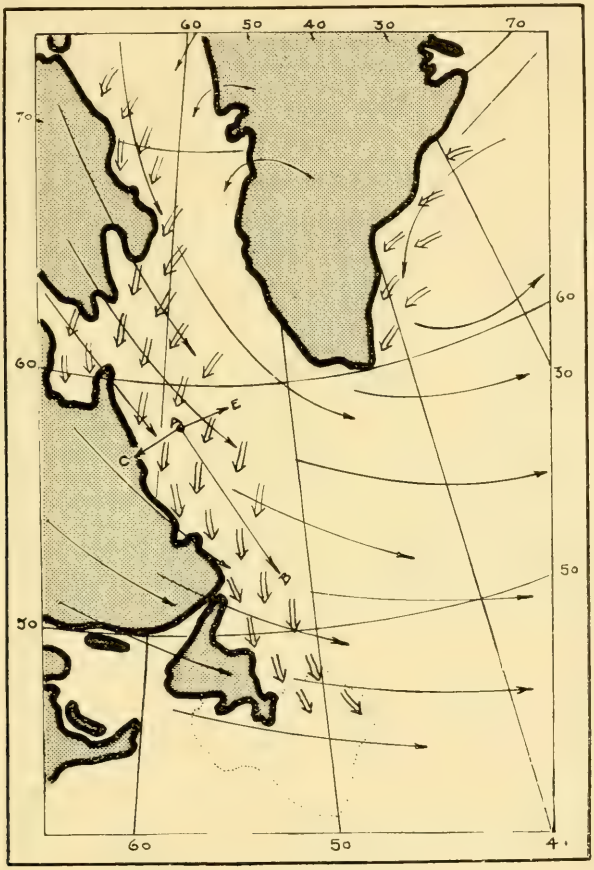
The four principal factors controlling the seasonal incursions of ice into the North Atlantic (plus several minor ones) are wind, current, pack ice, and winter temperature. They show an interrelationship so complex that each must be considered in connection with all the others.

In ice regions where steep atmospheric gradients prevail for a month or more at a time, the resulting steady winds may pile up large masses of water against shore lines, or against other hindrances, and thereby establish a deep, slope current.[76] A change in the character of the winds, due sometimes to the succession of winter and summer, may result, therefore, in seasonal variations in the size and velocity of the related ocean current.

A study of atmospheric conditions the year round shows that during the colder months when high pressure spreads over North America the northwestern North Atlantic is dominated by a strong northwesterly air stream. The exchange of temperature character between the land and the ocean, accompanying the increase of insolation in spring, calls forth a similar exchange of atmospheric pressure, and the northwesterly winds are replaced in spring by those from the southwest as well as variable ones. Thus the northwesterly winds from December to March amass great quantities of the surface water of Davis Strait against the 1,000-mile continental slope, from Baffin Land to the Grand Bank. Although the direction of the wind trends slightly off the coast, the deflecting effect of the earth's rota-

[76] An excellent exposition of currents established by winds is contained in Ekman (1928), to which the reader is referred.

tion is to the right—its force directly in proportion to the depth of the water at that particular place—causing the resulting gradient current to parallel the coast. The time required to develop a current of this sort must be calculated in weeks or months depending upon the magnitude of the wind and the width of the current. The prevailing wind system of December and March, over Labrador, correlated with the observed increase in the strength of the Labrador



THE LABRADOR CURRENT

FIGURE 113.—One of the fundamental factors causing a slope current to be built up along the Labrador coast is the prevailing northwesterly winds. The long curved arrows indicate the average wind direction December to March, and the short double arrows show the general movement of the water as a result. The diagram AB—CE represents the relative position of the Archimedian and the Ferrelian forces and the direction of resultant flow.

current off the Tail of the Grand Bank, from March to July, may represent the lag in the development of such an increase in flow. The dwindling of the cold current which is believed to occur in summer south of Newfoundland may be traced partly to the subsidence of the northwesterly winds in the preceding spring.

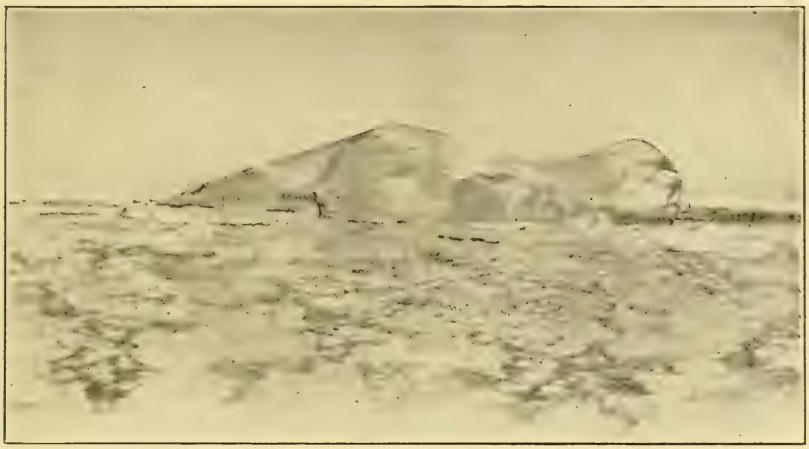
The volume and velocity of the Labrador current is also effected by purely hydrographical factors such as land drainage, and by the

melting of ice along the coast or over the continental shelves. The expansion and recession of reservoirs of light water along the margin of the current are especially significant in this connection. The greater the quantity and the lower the specific gravity of these masses, the steeper becomes the gradient and the greater the velocity and volume of the flow. The principal sources that contribute this coastal supply of light water are the melting ice,[77] and the drainage from the bordering land areas. Since contributions are at their maximum in spring and summer, their effect in imparting, therefore, a seasonal variation to the fringing ocean current is cumulative with that of the wind. Such oceanographic observations on the Labrador current as are at hand have been collected mostly in summer, and in the region south of Newfoundland. The investigations of the ice patrol, which embrace 95 per cent of the total, have been very scanty in February and March, but voluminous and in detail from May to July, with only one survey in October, 1923. Few observations have been taken of the current north of Newfoundland, except for those collected by the *Chance* in the summer of 1926, and by the *Marion* expedition and the *Godthaab* expeditions in the summer of 1928. The volume and strength of the Labrador current around the Grand Bank, as shown by comparing the ice patrol's data of temperature, salinity, and dynamic topography from year to year (see Fries 1922, p. 61; Smith 1923, p. 84; 1924, p. 109; and 1924a, p. 98), indicates the occurrence of seasonal variations. In normal years, at least according to the foregoing observations from around the Grand Bank, the cold current appears to reach a maximum during spring and early summer, after which it dwindles to a minimum in fall and winter. The fact that little Arctic ice drifts into the Atlantic during fall and early winter may also indicate that the Labrador current is then interrupted. It should be noted, however, that no year-round observations have yet been made from several sections distributed over the entire length of the Labrador current, and until they are collected we can not reach final conclusions.

The Labrador current is also marked by frequent irregular pulsations occurring within the interval of a few weeks, or of a month or two, all of which tends to mask the more important major cycle. (See figs. 99, 100, and 101.) These secondary short-term fluctuations may hark back to corresponding excesses or deficiencies developing in the light water along the continental slope from time to time. For example, a period of warm, wet weather around Newfoundland may appreciably accelerate the melting of the coastal ice and the discharge from the rivers, increasing the supply of light water, thus swelling the mass and the velocity of the southern end of the cold current. Under such conditions any ice in that particular locality will move south faster than will the ice situated farther north, although the

[77] Many statements have appeared in print that the cold Arctic currents are augmented by great quantities of ice which are melted every summer in the far north. The impression is therefore given that excess masses of light water or a high level at a riverlike source (in Baffin Bay for the Labrador current) constitutes the necessary head to discharge the water downstream into the Atlantic. It is high time that the river idea be discarded when considering unrestricted movements on a rotating sphere, under which class belong ocean currents. The pressure gradient which impels the stream flow of water lies on the right-hand side of the current in the northern hemisphere. Therefore, in order for thaw water to assist the flow of the Labrador current, the ice must be distributed and melt, not in Baffin Bay or the Polar Basin, but along the shelves of Baffin Land, Labrador, and Newfoundland.

total mass transport of the water, considering the entire breadth of the slope, may remain more or less uniform. The speeding up in one place along the slope, or retardation in another may cause the current, as noted on the dynamic topographical maps, to divide longitudinally into a series of bands or belts. In general the current is swift where narrow and sluggish where wide; for example, in July, 1928, as illustrated by the conditions met by the *Marion* expedition described on page 149. (See fig. 95.) Off Nain, under the conditions then existing, only the ice along the continental edge could have been receiving an appreciable southward set, while practically all of the ice then floating off Belle Isle was making southward progress. That an iceberg observed drifting southward at a moderate rate may suddenly accelerate to double, or sometimes to treble its former rate, is not evidence that the current throughout its entire length



ICEBERGS CAUGHT IN PACK ICE

FIGURE 114.—An iceberg surrounded by pack ice on the northern part of the Grand Bank, February, 1921. The majority of the first bergs appearing south of Newfoundland in early season, are sighted relatively wide off-shore, far to the eastward. An important factor of this seasonal variation in the course of the bergs is the entangling pack ice which, driving before the westerly winds, exerts a pronounced easterly component to the drift of the bergs. (Official photograph, international ice patrol.)

has also increased but rather suggests a contraction in breadth along the particular pathway that the ice is then following and this is an important point as yet little appreciated in oceanography.

The effect of pack ice on the seasonal distribution of icebergs is of two kinds: (*a*) When driven before the wind it entangles the bergs and tends to carry them along; or (*b*) it acts as a fender along the shoreward side of the Labrador current during a critical period, from November to July, tending to prevent the icebergs from drifting toward the coast and grounding as they are borne southward. This last assumption is corroborated by the fact that it is at the beginning of the season, from February through March, when the pack-ice belt is widest that the bergs are distributed farthest to the east of the Grand Bank.

If the velocity of the south-flowing current along the Labrador and Newfoundland continental edges, as suggested by the *Marion* expedition observations, of 12 to 14 miles per day be multiplied by the number of days of the normal iceberg season around the Grand

Bank (120), we arrive at an estimate of 1,200 to 1,500 miles as the length of the iceberg train on March 15 (an average date for the arrival of the van on the northern edge of the Grand Bank). On this date according to the assumption the rear icebergs would be about off Cape Dier, Baffin Land. It is not likely that any bergs that are still farther north at that date will drift past Newfoundland during the following summer because the coast line and adjacent shelf waters begin to be bared from south to north, as the spring advances, with the bergs tending to set on-short after June and so to become trapped. If pack ice were to blockade the Labrador and Newfoundland coasts the year round as it actually does from January to April, icebergs would drift into the North Atlantic throughout the year.

Pack ice may also tend to keep the icebergs moving faster in the currents than would otherwise be the case by holding them just outside the edge of the continent where the drift is most rapid. This banding of the current is well shown in the dynamic topographical map of the *Marion* expedition for the coast of Labrador. (Fig. 95.[78]) If, on the other hand, there is less pack ice than usual, the shelf waters are open to receive the bergs, under which conditions few of the latter ever return to the main current path. It is apparent, therefore, that the distribution of the pack ice exerts a great influence on the drift and numbers of icebergs coming out of the Arctic. An example of the foregoing effect was observed in the spring of 1924, a year when no pack ice was sighted south of Newfoundland, and when only 11 icebergs drifted south of that latitude, instead of the normal 380 to 400 bergs. This was not due to any lack of glacial ice to the northward as the *Tampa* on her northern cruise that May discovered a large number of icebergs, undoubtedly the spring crop headed for the Grand Bank, trapped and stranded along the coast from Cape St. Francis to the Strait of Belle Isle.

ANNUAL VARIATIONS IN THE NUMBER OF ICEBERGS PAST NEWFOUNDLAND

Great variations have been observed in the number of icebergs drifting out of Davis Strait into the North Atlantic—in 1929 there were over 1,300 to only 11 in 1924. As scientific observer, attached to the ice patrol, I have been concerned with the possibility of developing from these known factors, some method by which a practical forecast of the number of bergs which would drift past Newfoundland during the spring might be arrived at. The first part of this investigation centered on a thorough study of similar investigations in the past and a decision of what methods were the most promising.

Schott (1904), long ago drew attention the the great number of icebergs that came south of Newfoundland during 1903 and discussed the resulting effects. Campbell-Hepworth (1914), investigating the annual variations in the Labrador current compiled a table of the annual amount of icebergs for a 10-year period.

Mecking (1907, p. 3), with the records of the Deutsche Seewarte, the United States Signal Service, and the United States Hydrographic Office, as previously discussed, compiled a record of the

[78] Helland-Hansen and Koefoed (1909, p. 269) state that the sealing vessels along the east coast of Greenland, by forcing themselves into the border of the pack, inside the continental edge, escape the brunt of the southward current, which passed in plain view farther out.

number of icebergs sighted each year in the western North Atlantic, 1880-1897. When the yearly number of icebergs in the various tabulations were compared with meteorological conditions, it proved that during the years of record, a summer characterized by a greater than normal amount of offshore winds over the ice fjords, had been succeeded the following spring by more icebergs than usual off Newfoundland. The atmospheric pressure distribution suggested thereby as favorable for a great crop of icebergs is any one or two or more of the following conditions: A strong Baffin Bay low; a somewhat strengthened east Greenland high; a suppression of the Icelandic low; the west Iceland low, only an appendage of the Baffin Bay low; the east Iceland low is replaced by a high.

Mecking concludes that while the control of the ocean currents is fundamental for the berg drift out of the Arctic, the strength and intensity of the winds from year to year is the chief agency that causes the variation in the number of icebergs drifting into the western North Atlantic. He divides the important winds into two classes (*a*) the summer winds over the glacier fronts; and (*b*) the winter winds over Davis Strait. Mecking places most emphasis on the direction and force of the summer winds over the ice fjords of west Greenland between Jacobshavn and Upernivik. A summer characterized by an abnormal amount of east winds, according to his opinion, will blow more than an ordinary number of bergs out into Davis Strait, and this event is of proportions sufficient to cause a heavy iceberg crop off Newfoundland the following spring. With so few data from this little-known region for the period investigated Mecking's conclusions can only be regarded as tentative.

I have therefore continued with the iceberg figures where Mecking left off, bringing the records up to 1930, a total series of 50 years. (For iceberg anomaly table see Smith, 1927, p. 76.)

Number of icebergs south of Newfoundland (forty-eighth parallel) in western North Atlantic

Year	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual
1900	10	0	0	5	32	33	6	1	1	1	0	0	89
1901	1	0	0	4	13	29	22	6	5	1	2	5	88
1902	3	0	1	1	13	5	16	1	0	1	0	0	41
1903	0	2	400	166	151	52	23	7	0	0	0	1	802
1904	0	0	12	63	82	89	14	3	2	0	0	0	265
1905	3	2	168	373	109	100	50	9	8	8	0	15	845
1906	14	11	77	49	133	87	18	16	0	0	0	0	405
1907	0	1	11	162	248	138	64	11	0	0	0	3	638
1908	1	0	7	39	82	51	2	2	20	15	3	0	222
1909	0	55	147	134	321	181	121	45	19	1	0	0	1,024
1910	0	0	0	34	10	3	3	0	0	0	0	0	50
1911	0	8	41	112	72	77	21	40	3	0	8	14	396
1912	1	0	34	395	345	159	63	19	0	0	3	0	1,019
1913	2	4	37	109	292	71	14	4	7	0	6	4	550
1914	1	41	32	27	419	71	22	46	52	13	1	6	731
1915	14	72	67	96	97	71	28	17	5	0	1	0	468
1916	0	0	0	0	25	29	0	0	0	0	0	0	54
1917	0	0	13	3	3	9	10	0	0	0	0	0	38
1918	0	0	12	23	26	37	27	34	22	1	14	3	199
1919	3	4	5	25	75	56	26	36	69	2	12	4	317
1920	6	43	20	5	211	86	18	5	18	19	10	4	445
1921	17	5	43	210	198	175	53	24	4	10	1	6	746
1922	0	3	35	71	245	83	21	11	6	27	21	0	523
1923	0	3	28	65	83	42	10	3	2	0	0	0	236
1924	3	0	6	2	0	0	0	0	0	0	0	0	11
1925	0	3	5	8	58	22	13	0	0	0	0	0	109
1926	0	3	15	58	168	85	4	6	2	3	1	0	345
1927	4	10	26	93	153	95	5	3	0	0	0	0	389
1928	0	0	14	156	190	87	55	5	0	4	4	0	515
1929	0	0	45	332	460	376	107	1	0	0	18	12	1,351
1930	14	116	87	89	101	62	3	1	1	1	0	0	475

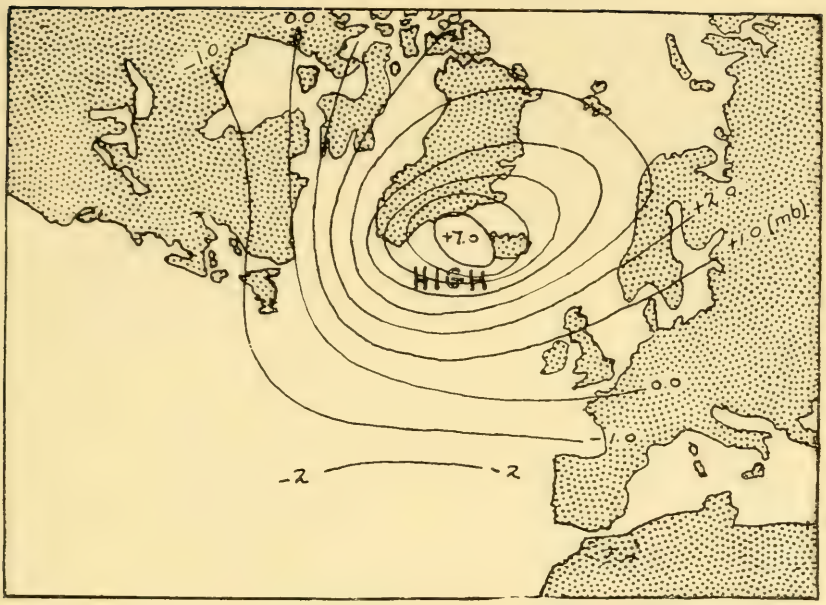
The correlations published as a result of these investigations at the British Meteorological Office in 1925 (Smith, 1927, pp. 45–48), lacked an actual count of the icebergs from 1900 to 1906, but in its place the value of the atmospheric pressure gradient Ivigtut to Belle Isle was substituted. As soon as access was had to the files of the United States Hydrographic Office a new table of iceberg values was constructed for this gap of seven years, and correlations were recalculated. The results of this work changed slightly the values of the correlations based upon the original variates.

The next task of the investigation centered upon a correct arrangement of the iceberg data, the period 1880–1926 being divided by years into six groups depending upon the relative number of icebergs present in the western North Atlantic south of Newfoundland. For example, the lightest iceberg years were found to be 1881, 1889, 1893, 1900, 1902, 1917, and 1924, and the heaviest iceberg years, 1903, 1909, 1912, 1914, and 1915.[79] The six groups were next arranged in accordance with a scale 0 to 10, but it was found by plotting a dispersion diagram that the iceberg values failed to follow the requirements of a variable; that is, the greatest number of deviations did not fall proportionately near the mean average. The fact that a value either too great or too small had been assigned to the iceberg years showed that too great weight had been given the extremes. In order to correct this error, it was necessary to arrange the years in conformity with the shape of a probability curve, and then from a transformation curve to obtain fresh relative values of the icebergs. The following table gives the values weighted according to scale 0 to 10:

Year	0	1	2	3	4	5	6	7	8	9
1880	4.7	2.4	6.1	4.7	6.4	7.4	4.0	5.0	4.3	3.5
1890	8.6	3.1	4.0	4.4	6.1	3.0	3.8	6.1	5.1	5.4
1900	3.0	3.0	2.5	7.3	4.1	7.4	4.6	6.4	3.8	8.6
1910	2.8	4.6	8.6	5.7	6.8	5.4	2.8	2.5	3.7	4.2
1920	5.1	6.8	5.9	4.1	2.0	3.3	4.3	4.8	5.6	9.0

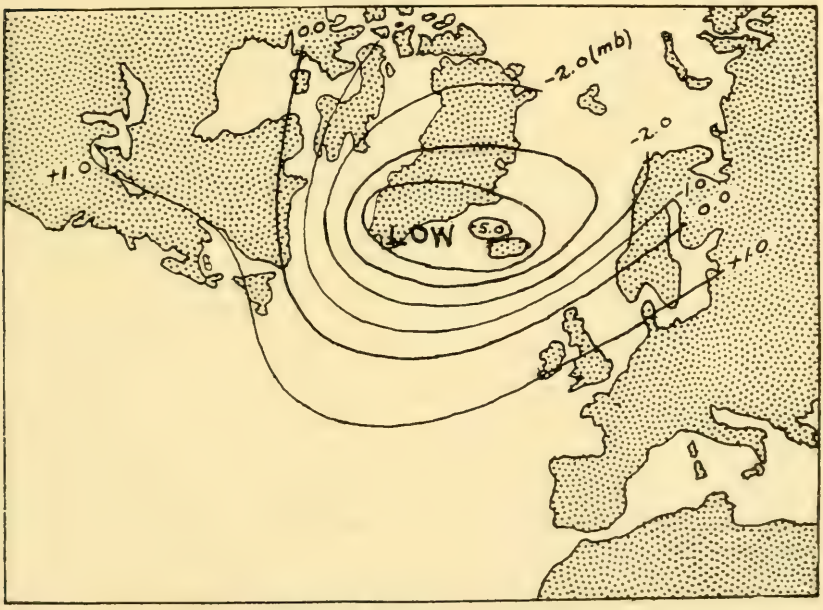
The meteorological material was next arranged in a form convenient for correlation with the iceberg values, which was made possible by the material and the assistance furnished by the British Meteorological Office. The best comparison between a series of years or months is afforded, not by pressure values direct but by the departures from normal. Station normals were obtained by arranging a long series of observations, and with the aid of these data isanomaly maps were constructed for the months October to March, for the years 1880–1926, inclusive. A classification was then attempted based upon a grouping of the monthly anomaly maps, October to March, for the northwestern North Atlantic region, from which it was apparent that two types of pressure distribution stood out clearly, one showing an excess of pressure centered in the region of Iceland and more or less dominating the ocean; the other, showing a dominance by the reverse condition. The first or "plus" type is subject to further classification into groups 1 and 2, depending upon a relative intensity of the excess mass of air. Combination

[79] The year 1929, with over 1,300 bergs south of Newfoundland, exceeds anything on record.



LESS ICEBERGS THAN NORMAL

FIGURE 115.—An atmospheric pressure map constructed by averaging the pressure anomilies for the period December to March in the years 1881, 1891, 1895, 1900, 1902, and 1917. These years are noted for the less than normal number of icebergs that drifted southward out of Davis Strait into the North Atlantic Ocean.



MORE ICEBERGS THAN NORMAL

FIGURE 116.—An atmospheric pressure map constructed by averaging the pressure anomalies for the period December to March in the years 1885, 1890, 1903, 1912, and 1921. These years are noted for the greater than normal number of icebergs that drifted southward out of Davis Strait into the North Atlantic Ocean.

of these weather-map values with the values of icebergs, shows that an excess of air over the northern North Atlantic has been reflected either in a very light, or at least a moderately light, iceberg season off Newfoundland during the following spring. The opposite type of pressure distribution is unmistakeable in showing a greater number of bergs than normal, but it does not permit of subgrouping. The plus type of pressure map, in other words, exhibits a higher correlation with poor iceberg years than does the minus type with correspondingly rich years. This indicates that there are other influences at work, such as variations in the air temperatures and in the water temperatures in the far north; variations in the precipitation, or perhaps sporadic phenomenon, such as an ice jam in the Arctic Archipelago.

It was found that pressure differences between various points furnished the most useful values for purposes of correlation, because in this method there is no room for the personal bias which may enter when charts are classified according to the types. We correlated first the annual variations of icebergs past Newfoundland with the pressure gradient for the previous summer over the ice fjords of west Greenland, the actual values employed being the pressure differences, Jacobshavn minus Upernivik, as obtained from the previously constructed isobaric maps of the Davis Strait region, 1880–1920.

Lag, in years	Correlation coefficient between icebergs and summer winds
1	−0.10
2	+.05
3	−.15
4	−.26

The months were then examined separately as follows:

Lag, in years	Correlation coefficient between bergs and winds			
	June	July	August	September
1	−0.11	−0.02	−0.11	−0.16
2	+.02	0	−.33	+.06

The negative value of the correlation is in the right direction, namely, that with pressure lower at Jacobshavn than Upernivik, the winds are directed offshore, and therefore tend to drive an abnormal number of icebergs into the current that eventually bear them past Newfoundland. The magnitude of the correlations are, however, insignificant and therefore we must conclude contrary to Mecking (1906), that although the offshore summer gradient is real, it is nevertheless of little importance, so far as effecting the dispersal of bergs south of Newfoundland.

In view of the foregoing, it was considered desirable to investigate quantitatively the effects of the following variates, beginning with those which appeared to bear the closest relationship to the number of icebergs so far as available data might allow:

(*a*) Represents the number of icebergs south of Newfoundland (the forty-eighth parallel of latitude).

We also have:

(*b*) Amount of pack ice off Newfoundland, February to May.

(*c*) Atmospheric pressure differences (mb) Ivigtut minus Belle Isle in November to April, preceding the iceberg season off Newfoundland.

(*d*) Atmospheric pressure difference (mb) Stykkisholm minus Bergen, August to January, preceding the iceberg season off Newfoundland.

(*e*) Cross gradient pressure Ivigtut minus Belle Isle, December to March.

(*f*) The pressure anomaly at Stykkisholm, December to March.

(*g*) The monthly precipitation at Upernivik, Greenland.

(*h*) The temperature during June, July, and August at Upernivik.

The correlation coefficient between (*a*) and (*b*) was found to be very high, +0.86, proving the correctness of the theory that pack ice is a vital factor in the dispersal of icebergs to the North Atlantic.[30]

If the assumptions previously set forth, viz, that an abnormal amount of northwesterly winds during the winter on the American coast south of Baffin Land results in a heavy iceberg season past Newfoundland, then we should expect to find good correlation coefficients between (*a*) and (*c*); actually they are as follows:

Between (*a*) and (*c*)—

Nov.	Dec.	Jan.	Feb.	Mar.	Apr.
−0.12	−0.45	−0.26	−0.22	−0.31	−0.09

Several other natural factors were tested in the hopes that coefficients found in the case (*a*) and (*c*) might be further raised, with the result that a correlation of −0.49 existed between (*a*) and (*f*). In order to test this effect when divorced from the Ivigtut-Belle Isle influence, partial correlations were calculated as follows:

$$r_{ac.f} = -0.30 \text{ and } r_{af.c} = -0.11$$

and the regression equation for bergs on a scale 0–10:

$$(a) = 0.33(c) - 0.05(f) + 4.8$$

The foregoing equation indicates that the Ivigtut-Belle Isle gradient has about six times as strong an influence upon the southward distribution of icebergs as have atmospheric pressure conditions at Stykkisholm. Both factors, however, are real and were employed in the following proportions:

$$\frac{6(c) + (f)}{5} \quad \ldots\ldots \quad (c')$$

[30] A correlation between numbers of icebergs and sun spots gave negligible results.

The monthly values for (c') were then correlated with (a) giving the following coefficients:

Between (a) and (c')—

Nov.	Dec.	Jan.	Feb.	Mar.	Apr.
−0.08	−0.48	−0.43	−0.26	−0.34	−0.15

It is apparent from this that the November and April gradients have little effect on the iceberg dispersal, probably because in November the berg train is just putting in its appearance while in April the character of the drift is well established. The values of 0.48 and 0.43 in December and January respectively indicate, furthermore, that those two months exert twice as strong an influence on the ice as do any of the other months of the year. When the value of (c') was expressed in monthly units and combined with the following weights:

$$\frac{2 \times \text{Dec.} + 2 \times \text{Jan.} + 1 \times \text{Feb.} + 1 \times \text{Mar.}}{5}$$

the 4-monthly group values of (c') correlated with (a) yielded the high coefficient of −0.58 for the period 1880–1926.

The Stykkisholm-Bergen pressure gradient gave the following coefficients which show that its iceberg effect does not commence until October and is mostly ended by February.

Between (a) and (d)—

Aug.	Sept.	Oct.	Nov.	Dec.	Jan.	Feb.
−0.07	−0.09	−0.26	−0.22	−0.53	−0.28	−0.10

The months were accordingly arranged in the following manner:

$$\frac{1 \times \text{Oct.} + 1 \times \text{Nov.} + 2 \times \text{Dec.} + 1 \times \text{Jan.}}{5}$$

Employing seasonal values combined as above, we obtained a coefficient of −0.62 for the series 1880–1926.

A study was also made of the degree to which the angle that the isobars make with the Labrador coast (see (e)) affects the number of icebergs passing Newfoundland. It had been assumed that the pressure gradient between Ivigtut and Belle Isle was nearly at right angles to the most effective direction of the wind; an angle of about 38° with the Labrador coast. Correlation values indicate that 30° off the general coastal trend is the most effective direction, and since the run of the wind in high-pressure centers is about 30° across the isobars, the most effective wind direction is, therefore, approximately 60°.

The remaining variates tested in the investigation were (g) and (h), with a lag of one year, which gave the following coefficients:

Between (a) and (g)—

Jan.	Feb.	Mar.	Apr.	May	June
−0.14	−0.29	−0.27	+0.19	−0.33	−0.22
July	Aug.	Sept.	Oct.	Nov.	Dec.
+0.19	−0.17	−0.15	−0.23	−0.07	−0.03

The coefficients of −0.29, −0.27, and −0.33 in February, March, and May appear quite high, but the precipitation at Upernivik involves other factors such as the distribution of the pressure, and of course, attending winds. The last variate considered was the temperature during the summer months at Upernivik, (*b*) giving the following correlation coefficients:

Between (*a*) *and* (*h*), lag one year—

June	July	Aug.	Sept.
−0.11	−0.02	−0.11	−0.16

Between (*a*) and (*h*), lag two years—

June	July	Aug.	Sept.
+0.02	0.0	−0.33	+0.06

These values are so small that we may discard any effect of summer temperatures along the ice-fjord coast of west Greenland, so far as any effect of the iceberg distribution is concerned.

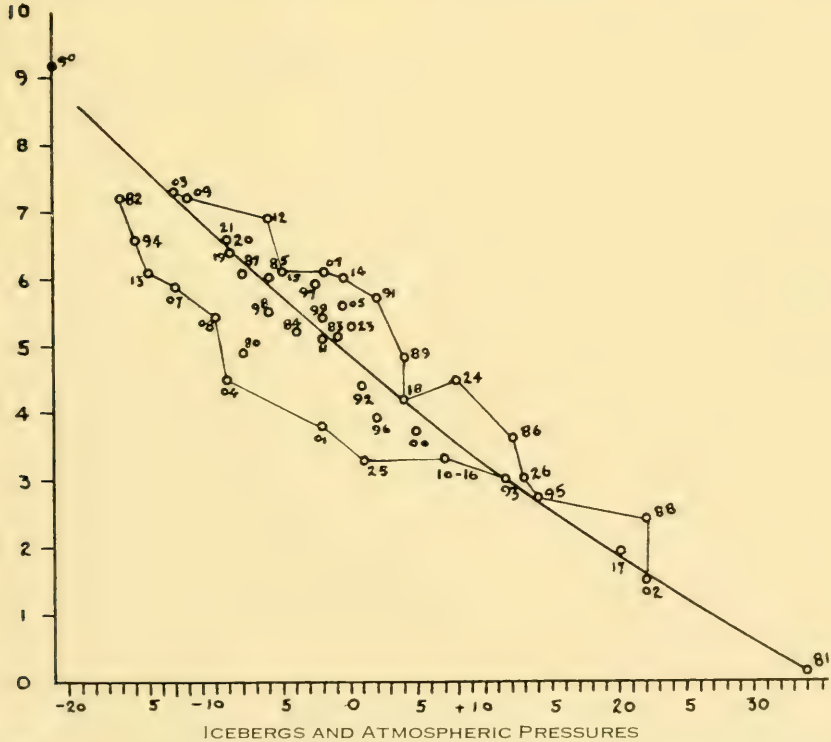


FIGURE 117.—The graph clearly shows a definite relation between the annual number of icebergs south of Newfoundland (the ordinate scale 0–10), and the departures from the normal atmospheric pressure conditions over the northwestern North Atlantic. The abscissa values were obtained by substitution of the pressure data in the regression equation on p. 188. The numbers on the graph indicate the particular year investigated.

The correlation coefficients determined by the foregoing methods indicate that of all the variates considered, the atmospheric pressure difference between Ivigtut, Greenland, and Belle Isle, Newfound-

land, is the most important in causing yearly variation in the number of icebergs. The agency of next importance, about one-third of the Greenland-Newfoundland factor, is the pressure difference, Stykkisholm-Bergen. The other variates exert so little effect that they warrant no further attention.

Both factors (c') and (d) were found to be real as shown by the following partial correlations:

$$r_{ac'.d} = -0.32 \qquad r_{ad.c'} = -0.39$$

This leads to the development of a regression equation for forecasting purposes, on the basis:

$$(a) = -0.08\ (c') - 0.12\ (d) + 4.8$$

When the calculated value of (a) was plotted for the 47 years, 1880–1926, against the actual value of recorded bergs it was found that only in 8 instances did the calculated value differ by more than ±2, on a scale 0–10. The years exhibiting the greatest departures were 1885, 1901, 1905, 1908, 1909, 1912, 1914, and 1924. Such occa-

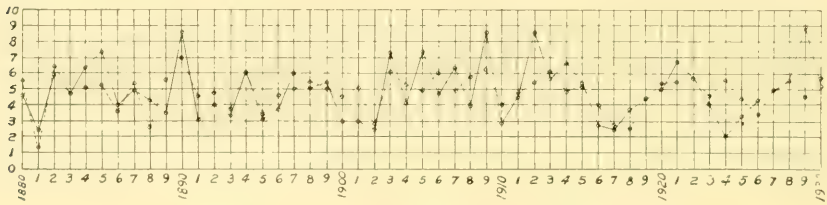


FIGURE 118.—The full line represents the number of icebergs south of Newfoundland each year on a scale of 0–10, mean value 4.8. The dashed line is the forecasted number of bergs obtained by substitution of the necessary meteorological data in the regression equation (p. 188).

sional abnormalities may be due to any one or all of several causes, for instance, ice jams in the Arctic Archipelago; variations in precipitation; winter storms; summer calms; the breaking away of the fast ice of Melville Bay and, possibly, of that fronting the great Humboldt Glacier.

The foregoing correlations, and the regression equation constructed therefrom, have been employed by the international ice patrol to forecast the number of icebergs that may be expected to drift into the North Atlantic during the spring. The recent establishment of meteorological stations in Greenland and northern Canada, and the prompt receipt of their reports by radio permits the use of monthly mean values and the immediate construction of isobaric maps. In using such in the equation $(a) = -0.08(c') - 0.12(d) + 4.8$, Isafjord (Iceland) has been substituted for Stykkisholm, and Julianahaab (Greenland) for Ivigtut; c', therefore represents:

$$c' = \frac{6\,[(2\times \mathrm{Dec.}^{\mathrm{J-BI}} + \mathrm{Dec.}^{\mathrm{I}}) + (2\times \mathrm{Jan.}^{\mathrm{J-BI}} + \mathrm{Jan.}^{\mathrm{I}}) + (\mathrm{Feb.}^{\mathrm{J-BI}} + \mathrm{Feb.}^{\mathrm{I}}) + (\mathrm{Mar.}^{\mathrm{J-BI}} + \mathrm{Mar.}^{\mathrm{I}})]}{5}$$

$$d = \frac{(\mathrm{Oct.}^{\mathrm{I-B}} + \mathrm{Nov.}^{\mathrm{I-B}} - 2\times \mathrm{Dec.}^{\mathrm{I-B}} - 2\times \mathrm{Jan.}^{\mathrm{I-B}})}{5}$$

Where ^{J} = Isanomaly pressure value for Julianahaab; BI = Isanomaly pressure value for Belle Isle; I = Isanomaly pressure value for Isafjord; and B = Isanomaly pressure value for Bergen.

A service of iceberg forecasting, covering the spring season for the western North Atlantic based upon the above correlations, was instituted in 1926 and has been continued through 1929. Forecasts compared with actual records have been as follows:

ACTUAL	Bergs	FORECAST	Bergs
1926, below normal	345	1926, below normal	150
1927, normal	390	1927, normal	386
1928, above normal	515	1928, above normal	500
1929 (estimated [51]), far above normal	1,300	1929, below normal	350
1930, above normal	475	1930, above normal	520

A comparison of the actual number of icebergs with those forecasted a month prior to the inauguration of the berg season, shows four successes out of five predictions. We are able, apparently, to predict quite accurately the smaller deviations from normal from year to year, but the occasional great iceberg seasons such as 1909, 1912, and 1929 come without meteorological warning. As our knowledge of ice, meteorology, and oceanography of polar conditions increases we have good cause to expect improvements in iceberg forecasting.

ANNUAL AMOUNTS OF GLACIAL ICE AND SEA ICE

Considerable quantitative data have been compiled by comparing the relatives volumes of glacial ice and of sea ice discharged into the northwestern North Atlantic annually. The material discharged from the iceberg fjords, set forth in the foregoing discussion (p. 95), at 4.6 cubic (nautical) miles (29^3 kilometers), more or less is estimated to amount to 70 per cent of the annual output. The total wastage for the entire coast of west Greenland is something like 7 to 10 cubic miles (42^3 to 63^3 kilometers). The American shores are believed to contribute about 0.3 cubic mile (1.9^3 kilometers).

The total area of sea ice which is fed into the northwestern North Atlantic during any one season, and which melts annually, may be calculated quite closely from a map of these regions (see fig. 121, p. 200), upon which is plotted the ice and water temperature records of the northwestern North Atlantic. The shelf area included between the coast line on one side and the 200-fathom depth contour on the other approximates the bounds of melting sea ice. The area in which we are particularly interested is equal to 467,300 square miles, and on Figure 121 it is the shaded portion bounded by the full black line. Occupying the same general inclosure but spread out somewhat farther on its outer side, to and including the dotted line, lies the melting area. The fact that sea ice remains on the whole within the

[51] The single failure occurred in the year 1929. The critical months for Davis Strait—December to March—showed December pressure normal, but January was made notable by one of the greatest excesses of atmospheric pressure over Iceland ever recorded. Europe, in consequence, suffered a severely cold winter and Davis Strait one of the warmest Decembers on record. This system of winds distinctly forbade the southward transportation of ice. February witnessed a "high" over the Canadian Maritime Provinces and a "low" southeast of Cape Farewell, thus initiating a northwesterly air stream which proved as favorable for the invasion of ice as conditions the previous month were unfavorable. The fact, however, that December is given twice the weight of January in the equation employed makes the former much outweigh the latter in the forecast. March was only moderately favorable to ice importation and therefore it failed materially to effect the final prediction of a berg year below normal. The presence of over 1,300 icebergs south of Newfoundland in 1929 can not, therefore, be explained by meteorological factors, but must have resulted from some other cause, perhaps one of those listed on p. 188.

limits of the shelf waters is the reason for the areas of accumulation and dissipation being practically one and the same. The melting area, it will be seen, includes not only the waters inshore of the continental edge, but a marginal mixing zone. This strip naturally varies in width depending to a great extent on the particular contour of the slope, the direction of the current, and like factors at any given locality. For example, off the Tail of the Grand Bank the belt of mixed waters spreads out sometimes 60 to 100 miles or more in width; a zone much broader than is found farther north in Labrador. A conservative estimate of the average width of the mixed water zone along the North American shelf is 30 miles. The greater size of the melting area over that of the ice area is balanced, however, by an area along the east coast of Baffin Land which extends from Barrow Strait to Cumberland Gulf. Pack ice is never absent from this region except for a week or two the first part of September, and then seldom more often than once in four or five years. It is safe to designate this region as one in which no melting prevails. The fact that it is in size just about equal to that of the mixing zone causes ice area and melting area to be approximately equal, viz, 467,300 square miles.

The estimate of 467,300 square miles of pack ice as the annual crop is believed to be somewhat high, but it adds conservatism to our figures on the magnitude of the ice-chilling effect to be introduced later. It should be also remarked that the pack ice area does not present a solid, unbroken surface, since leads and polynyas prevail in the pack just the same, or to a greater degree, as they do in the polar ocean. The figure of 10 per cent of open water for the polar cap ice is believed to be also fairly representative for the pack ice of Davis Strait and Baffin Bay. The 10 per cent value of open water in the pack does not, however, materially effect, for purposes of comparison, the roughly equivalent size of melting area and ice area.

In order to compare the annual amount of sea ice and glacial ice in this interesting arm of the northwestern North Atlantic we have assumed the average thickness of the pack ice to be 6 feet. Naturally the floes and fields are thicker at the northern extremity of Baffin Bay than they are much farther south near the melting end off Newfoundland. Also where rafting and hummocking has occurred the pack will be thicker than where young ice lies. But an average of 6 feet is believed to be a fairly close approximation. The total volume of sea ice melted annually is therefore 467 cubic miles. Since over a long period of years there has been no tendency toward a persistent increase or decrease in the geographical position of the ice boundaries, it follows that during the course of a normal year 467 cubic miles of water are frozen into ice and a like amount of ice is melting into water.[82] Compare this figure of 467 cubic miles of sea ice with that of 7 to 10 cubic miles as the annual output of glacial ice for Davis Strait and Baffin Bay. Though our figures may only be approximate, they unmistakably reveal, nevertheless, that *sea ice annually greatly exceeds glacial ice in volume, the latter equaling only 0.015 to 0.020 of the former.* In fact evaporation of the melting

[82] Speerschneider (1926, p 74) estimates an annual discharge from the polar sea of 26,000,000,000 cubic yards. This is far too small since, if 6 feet thick, it would amount to an area covering only 1° of longitude and 3° of latitude off northeast Greenland.

area also exceeds in effect that of the glacial ice crop and the annual volume of snowfall even, that falls on the waters in question, if expressed as ice, closely approaches the annual glacial discharge of west Greenland.

In the case cited for comparison we selected one of the most prolific glacial discharging regions in the north. If the annual production of all sea ice from all sections in the north had been compared with the total glacial discharge from all Arctic lands, the relative insignificance of the latter's mass would become still more emphasized.

The annual iceberg discharge appears much greater in volume than it relatively is, for one reason, on account of the conventional method of recording the positions of icebergs on a map as small circles or triangles. These symbols necessarily can not be drawn to scale; if so, they would be invisible, so actually they appear on the chart greatly exaggerated. If all the glacial discharge from west Greenland for one year could be amassed in a single berg and then drawn to the scale of Figure 121, page 200, it would be represented by the minute black rectangle off Cape Farewell, Greenland, marked "M." If the annual amount of glacial ice be spread out to a thickness of 6 feet (the same dimension as that assumed for the sea ice), and then drawn to scale on Figure 121 it will equal the area of the shaded rectangle south of Greenland marked "N."[83]

We have estimated that about one-twentieth of the sizable icebergs calved from west Greenland each year (400) find their way south of Newfoundland. If it were possible for the annual production of sea ice to form itself into separate sizable bergs, and the same relative distribution prevailed, the North Atlantic would be invaded every spring by approximately 230,000 icebergs.

EFFECT OF NORTHERN ICE ON TEMPERATURE AND CIRCULATION OF THE WATERS OF THE NORTH ATLANTIC

Among the most important, far-reaching events in northern seas, according to many oceanographers and meteorologists, are the thermal effects which accompany ice formation and its dissipation. That energy necessary to lower by 1° F. the temperature of a surrounding volume of water about eighty times the size of the ice is a commonplace.[84] The relatively great number of units of heat energy needed for the foregoing process represents the energy required to overcome the cohesion of the solid ice molecules as the compact crystals are changed into liquid form.[85] More than 1,000 cubic miles of ice, it is

[83] Ricketts (1930, p. 109) has similarly emphasized the exaggerated notion regarding icebergs by stating that if the whole season's bergs south of Newfoundland for the year 1929 (1350) be spread out over the Newfoundland shelf region they would make a uniform layer of ice only about one-eighth inch thick. If glacial ice did not take form as massive iceberg bodies, such ice would be a rarity south of Labrador.

[84] Heat of fusion in case of sea-water ice is approximately 67 calories.

[85] The heat of fusion is defined as the energy consumed or liberated upon transition of matter from a solid to a liquid state, or vice versa, at a constant temperature. The composition of sea ice, it will be recalled, is a conglomeration (not uniform) of pure ice crystals, salt crystals and brine, and theoretically, therefore, under natural conditions, sea ice never can exhibit a definitely fixed melting or freezing point. The heat of fusion of sea ice in its commonly accepted solid state is found to vary inversely as the salinity. Pettersson (1883, p. 268) gives the following data:

Salinity, 0/00	0	10	20	30	35	40
Heat of fusion (calories)	77	67.5	60.5	54.5	52	49.5

estimated, are annually transformed from solid to liquid and then back to solid again in the northern North Atlantic along a front from Spitsbergen to Labrador, and a phenomenon of such magnitude must bring into play a seesaw of enormous quantities of heat energy.

Several theories have been advanced, based on this effect of the latent heat. Pettersson (1904), and (1927), as a result of laboratory experiments with ice melting in a tank of salt water see fig. 84, p. 126), having reference to the distribution of temperature and salinity in the world's oceans, has presented a very fascinating theory explaining (*a*) the major oceanic system of circulation, and (*b*) the source of the great reservoir of cold bottom water. In order to test the extent of the latent heat effect he placed a block of ice in a tank of salt water and colored the water in such a manner that it would clearly show any signs of circulation. Soon after the ice began to melt three distinct movements were perceived, viz, an indraft of relatively warm water in the mid-depth of the tank and two cold outflows from the ice, one on the surface of the water and the other toward the bottom. Applying this to the general circulation of the North Atlantic (the tank representing the Atlantic Basin, its contents the ocean, and the block of ice the real front of the melting ice stretching from Spitsbergen to Newfoundland), he concludes that the current toward the ice, at mid-depths, corresponds to the warm layer which, depending on the total depth, is actually found in most northern seas. The outward expansions of melting ice, surface and bottom, correspond, respectively, to the cold icy surface currents, Labrador and east Greenland, that flow into the Atlantic and to the great reservoir of cold bottom water that fills most of the North Atlantic. The fact that the circulation established in the laboratory has so striking a counterpart in the longitudinal distribution of warm and cold water in the Atlantic is proof for Pettersson that the latent heat of melting ice is, (*a*) the main energizing agent responsible for North Atlantic circulation, and (*b*) the chilling effect which initiates great downpourings of cold bottom water.

Nansen (1909) (1912), on the contrary, is of the opinion, after many years of careful observation at sea and in the polar regions, that ice melting does not greatly affect the major processes of circulation nor chill the supply to the great bottom masses of the ocean. He believes:

(*a*) The phenomena should be considered as action and reaction.

(*b*) The processes are distinctly superficial when one considers the relatively great average depth of the ocean.

(*c*) The stratified condition characteristic of the ocean prevents the influence of melting ice from extending any significant field of forces downward into the deeper strata.

(*d*) The fact that ice in northern seas is usually separated from the cold bottom water by a warm insulating layer precludes the possibility of any relationship between the ice and the bottom water.

(*e*) The winter freezing is from top, downward; the summer melting is in similar order.

Finally Nansen (1909, p. 321) claims to have discovered the source of the great bottom water, not in the cooling effect of melting ice but

in the winter chilling of the surface layers in certain oceanic localities, notably from northeast of Jan Mayen to east of Cape Farewell as shown by Wüst (1928, Tafel 35).

The heat of fusion exhibits itself in two forms, (*a*) when ice is melted into water, as previously described, heat energy is consumed, and (*b*) when water is frozen into ice heat energy is liberated. The ice molecules represent a phase of less energy than the water molecules, the difference between the two phases being measured as heat energy in terms of heat of fusion. If a gas be transformed to a liquid or a liquid to a solid, heat energy is released and such heat exchanges, for example, have been known to constitute one of the main energizing components of cyclones. A rise in temperature has also actually been observed at cloud altitudes when water vapor has been rapidly precipitated into the rain of thunderstorms. In the case of freezing, when a quantity of homogeneous water molecules are aligned into the compact uniform arrangement of ice crystals, the heat energy released produces a material retardation in the freezing processes [56] and this phenomenon, moreover, is of such magnitude in the polar regions that it tends to stabilize the seasonal fluctuations. In other words, without the heat of fusion acting and reacting, more ice would be formed in the colder months of the year, and more and greater areas of open water would be found in summer than is actually the case.

Pettersson, in his laboratory experiment, has purposely stressed the importance of the effect of ice melting rather than that of water freezing, because in the first case the currents set up by the relatively great immersion of the ice itself causes the cooling effect to spread more completely into the water itself than into the air. On the other hand, when water in the sea is frozen as Pettersson points out, the release of heat energy is not absorbed by the water but by the atmosphere. It would be interesting to determine by tank experiment what currents are really established in the water mass as a result of freezing of the surface layers of salt water. It is difficult to see why there should be any material difference over an extended period in the disposition of heat in the case of these two processes—radiation and absorption—regardless of the high specific heat of the water, provided that the freezing and melting takes place in the same general region. The recipients or contributors of the energy are the surrounding media, i. e., the atmosphere above and the sea on all other sides.

In order to obtain a clear insight into the extent to which the heat of fusion may spread, it is necessary to take into consideration the position and size of the areas of ice production and ice dissipation in the North Atlantic and the physical state of the water which fills the northern basins. The principal regions of ice formation and dissipation agree well with the distribution of cold water. The ice nucleus is centered in the heavy polar cap extending from which are two arms one along the east coast of Greenland, the other along the east coast of Arctic North America. This ice covers about one-

[56] Barnes (1928, p. 2) describes the process of fusion as molecules of water are passing from the liquid to molecules of ice in the solid state, each ice crystal being covered by a heat mantle.

fourth of the total surface of the North Atlantic, Arctic Ocean included, of which about one-sixth is normally frozen and melted each year, so that one-thirtieth of the entire surface of the Atlantic is annually exposed to the latent heat phenomena. A comparison of vertical dimensions shows that only about 0.001 of the mean depth of the Atlantic is ice filled and if volumes are matched, only 0.0005 of the entire ocean is given over to ice. It is obvious from this that the proportions, horizontal to vertical, definitely prove the ice processes as wholly superficial.

The principal regions of ice production are the coastal seas of Eurasia and the offing of the latter, of east Greenland, and of Arctic North America. The principal regions of ice production are also the general regions in which the ice melts in greatest amount during summer. It follows that the life cycle of nearly all the ice is spent in waters whose outer bounds are the continental edges, and, although a certain quantity of ice is carried out into the deep ocean basin, this forms only an exceedingly small proportion of the whole.

Therefore, the raw materials, so to speak, out of which the great mass of ice is created, and back to which it returns, are the great reservoirs of shallow coastal waters of the north. During the colder months of the year, depending upon the severity of the air temperatures, these become readily chilled (from the top down to depths of 100 to 200 meters) close to the freezing point and ice forms in large amounts on the surface. Thus the actual appearance of the ice marks the release of heat energy, partly to the atmosphere and partly to the water, but the creation of the surface covering of ice tends to insulate the deeper waters and thereby protects them from further rapid freezing. The low temperature of the atmosphere, therefore, produces over shelves and northern seas, at the end of winter, a relatively deep frigid body of water, in the upper few feet of which then floats a covering of ice. It is obvious that practically no melting, hence no withdrawal of heat energy from the water, can take place even on the underside of the ice, as some have claimed, while it lies insulated in such boreal surroundings.

With the approach of summer increased radiation from the more perpendicular rays of the sun becomes applied to (*a*) the top of the ice cover, and (*b*) to the surface water where this is exposed. The ice and water are thus warmed simultaneously, but due to the far greater capacity of the water to absorb heat and also to spread the heat rapidly downward to 10 to 25 meters, the ice is soon completely enveloped, even on its underside, by relatively warm water. The sun-warmed surface layers, moreover, due to their great specific heat, are much more effective as a melting agency than are the direct rays of the sun on the ice itself. The latent heat of melting, it is true, tends to retard the ablation of the ice but continued solar radiation striking its surface more than counterbalances this. As the ice melts it loses draft and, floating higher and higher, it becomes more and more confined to the shallowest and warmest stratum which materially accelerates the rate of its dissipation. All this is made evident by the shrinkage of the floes, by the expansion of the areas of open water between the fields, and by the growth of char-

acteristic leads along the shore line with which visitors to the Arctic are familiar. The heat causing the melting comes from the sun above and not from warm ocean currents beneath.[87]

Observations also show that the upper 20 to 40 meters of the water column in sub-Arctic seas at the end of summer is relatively warm and fresh; actually it is heated thaw water. (Helland-Hansen-Koefoed, 1909, Pl. LXVI; and Smith 1926, p. 3.) This top film rests on a sharply contrasting icy substratum which in reality is a relic of the previous winter's chilling. The boundary surface spreads out more or less horizontally at 20 to 40 meters; the German so-called sprungschicht, or discontinuity layer, represents the limit to which the summer heat has been directly carried down by mixing by the waves, etc. Proof that this icy underwater is reminiscent of winter cooling and is not produced by the ice is shown during summer by position of the upper surface of the cold bottom layers, it lying below the deepest draft of the ice, while the latter is actually melting.

Any ocean in vertical cross section presents no homogenous basin content but is composed of a great number of horizontal layers, the thinnest and lightest floating on the top and the thickest and heaviest resting on the bottom.[88] The most pronounced stratification of the oceans in high latitudes is generally over the slopes of the basin and near the shore where the density gradient above the "sprungschicht" in the upper 100 meters becomes abrupt in summer. The stable arrangement of the water particles under such conditions resist any force that may tend to cause a vertical exchange between two different layers, for which reason both the latent heat of melting and the direct chilling effect of melting ice are prevented from proceeding to any appreciable depth. The thaw water both of land and of sea ice is much lighter than the upper 20 to 25 meters in which the ice floated before it thawed, therefore, the thaw water remains to mix with surface layers.[89]

As proof that the effect of ice melting does not extend to any appreciable depth in the ocean, the reader is referred to a series of records of temperature and salinity obtained February 28, 1921, within a few hours of each other at two stations on the Grand Bank, respectively about 60 and about 120 miles southeast of Cape Race, Newfoundland. (Smith 1922, p. 64). Conditions at these two stations normally differ little from each other but on the morning in question there was open water at station 141 while station 142 lay just within the edge of an ice field that extended northward as far

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[87] Helland-Hansen-Koefoed (1909, Pl. L–VI), by means of deep-sea thermometers serially spaced at various points in the east Greenland current, recorded the temperatures of the pools of warm surface formed in summer amongst the ice cakes and also the rate at which the solar radiation penetrated downward from the surface. Malmgren (1928, p. 14), from the *Maud*, within the Arctic Ocean pack, observed the sun melting the surface ice, while on the underside of the cover new ice was making. The thaw water with a high freezing point, it is claimed, runs down from the top of the ice cover, but freezes upon encountering the sea water with its normal summer temperature of about −1.6° C. In this manner a considerable part of the top cover of the ice is transported as an increment to the underside.

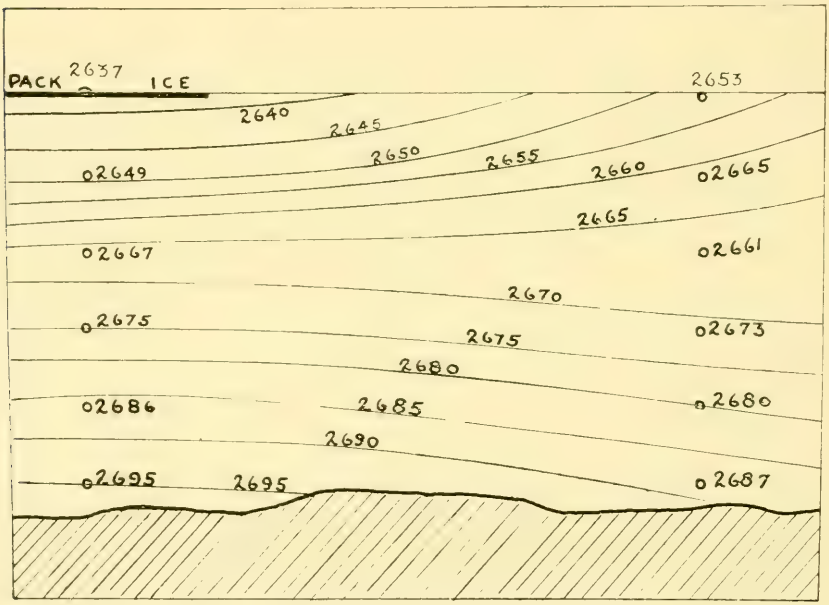
[88] Defant (1929, p. 13) has emphasized the stability of the layers in the water masses of the oceans.

[89] Pettersson (1883, p. 252) was apparently of the same opinion. He wrote, "that part of the ice which melts in the Arctic Sea leaves the water in a condition little favorable for immediate diffusion. The density of the ice water at the melting point being inferior to that of salt water beneath the ice, the melting process will tend to produce a thin superficial stratum of fresh water."

as the eye could reach. The physical state of the water column was as follows:

Depth, meters	Station 141 (open water)			Station 142 (in ice field)		
	Temperature	Salinity	Density	Temperature	Salinity	Density
0	−1.1	32.97	26.53	−0.5	32.80	26.37
15	−.8	33.04	26.65	−.5	32.86	26.49
30	−.5	32.93	26.61	−.5	32.99	26.67
45	−.8	32.96	26.73	−.6	32.99	26.75
60	−.8	32.95	26.80	−.6	33.04	26.86

While the temperatures, surface to bottom, was practically alike, the salinity of the upper 15 meters was very much lower in the melting ice field than it was to the southward, causing the density of the



THE STABILITY OF A WATER MASS IN THE ICE REGIONS

FIGURE 119.—The distribution of density which was found on February 28, 1921, near Cape Race, Newfoundland, along the southern edge of a field of melting pack ice. The marked stability of the water layers, at the time, precludes any interchange of light-thaw water from the surface with the heavier water near the bottom.

surface layers—a mixture of thaw water with sea water—to be correspondingly lower.[90] The fact that these results show no such distribution of temperature and salinity as would follow from Pettersson's tank experiment is evidence that natural conditions were not fully simulated in his laboratory experiment. The proportions of thickness of ice to depth of sea, the distribution of

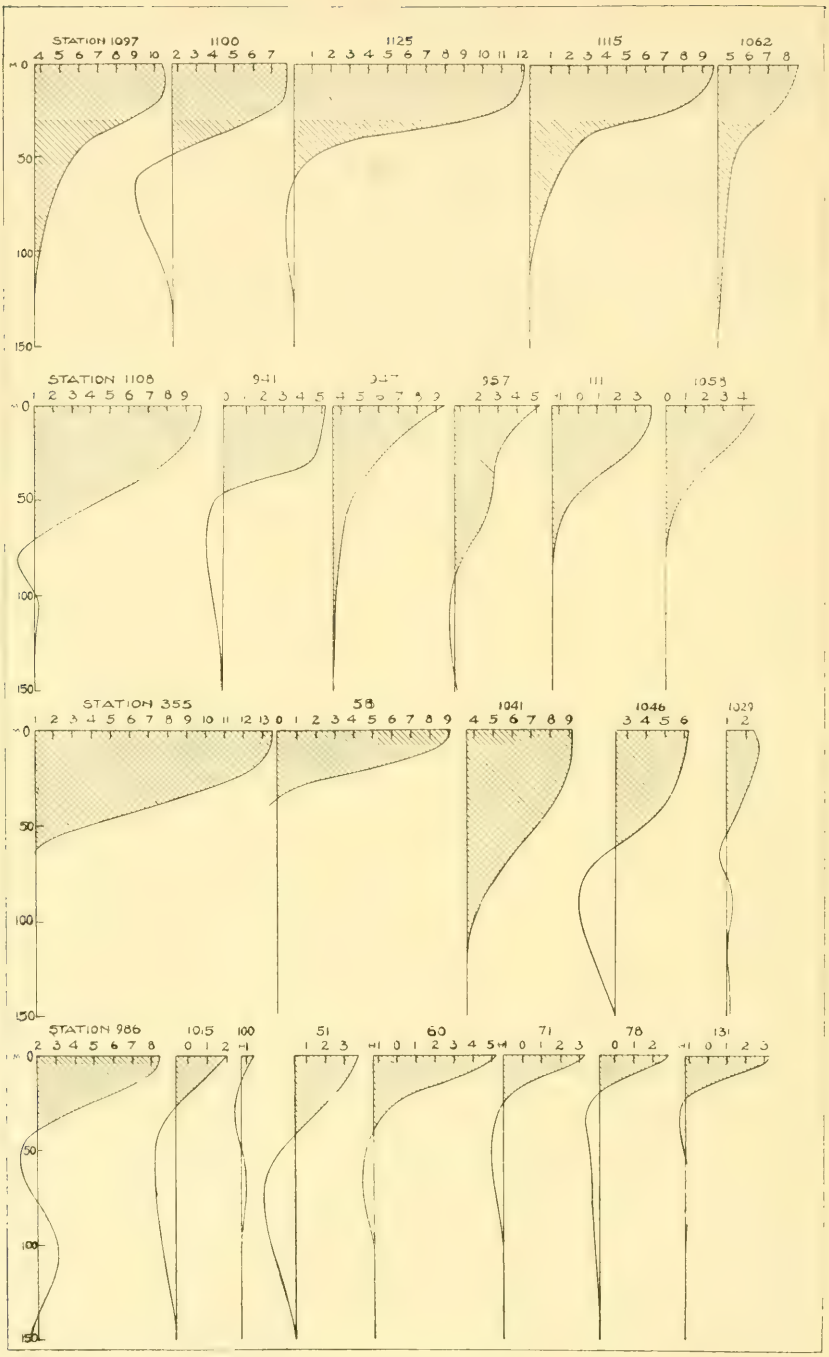
[90] Ricketts (1930, p. 120), in discussing the conditions around a melting iceberg observed by the ice patrol south of Newfoundland, concludes that the prevailing distribution of temperature and salinity in the sea with depth hydrostatically prevents the establishment of a vertical circulation.

temperature and salinity, and the stratified condition of the water under natural conditions are all difficult to reproduce exactly and failure to do so may be vital. For instance, Pettersson used a block of ice in his tank experiment which extended one-third the depth to the bottom, while under natural conditions the proportions of mean draft of ice to the average depth in the North Atlantic is 1 to 1,000.

In the formation of land ice the liberation of the latent heat of freezing is due to the atmosphere; while the melting of land ice (as icebergs) in the sea, withdraws the latent heat energy mostly from the water. The reaction, therefore, is somewhat different from that in case of sea ice where the processes of latent heat, both freezing and melting, begin and end in the sea itself. While the mean draft of icebergs is about sixty times greater than the mean draft of the sea-ice cover it is only one-fifteenth that of the mean depth of the Atlantic, consequently the thermal effect of the melting of the icebergs is similarly confined to the upper stratum.

If the effect of the two phases of latent heat, as we just qualified, were always exerted within the same general region, heat energy would be liberated and consumed in the same place. It is because of the transportation by the winds and the ocean currents of (*a*) the ice itself, and (*b*) the surrounding chilled thaw water that distant temperate climes are actually affected. As previously stated, the pack ice, drifting southward in two prominent streams, the East Greenland and Labrador, seems to exhibit a marked tendency to remain inside the continental edges of the respective shelves. This may be partly illusory, for any ice which drifts out across into warmer oceanic water quickly disappears through rapid melting, but only partly so, for both the ice and the current trend on shore. Icebergs are no exception to the above rule, despite the phenomenal drifts occasionally reported, the number of such are comparatively few. This coastal concentration of icebergs in Davis Strait was strikingly emphasized in the summer of 1928 when out of a total of over 2,000 bergs sighted by the *Marion* expedition, only 3 were observed outside the 500-fathom contour (see fig. 93, p. 142). Thus only a very small percentage of the northern ice actually floats out into the Atlantic basin to melt directly in contact with the ocean water itself. Consequently, while the ice melts in the surface layers of the coastal zone, it cools oceanic waters only indirectly by the mixing of the two waters, and the extent of this mixing, considering the ocean generally, is surprisingly small.

We found (see p. 190) that glacial ice in the Davis Strait-Baffin Bay region is only about 2 per cent of the volume of sea ice. In this connection the ice area and the melting area of these regions were calculated as 467,300 square miles. (See fig. 121, p. 200.) Employing the foregoing data we are now ready to determine quantitatively the magnitude of the chilling effect of ice melting on the water masses of the North Atlantic. The fact that melting area and ice area are approximately equal (see p. 190) allows a direct comparison between the relative thickness of the ice and the depth of the surface layer through which the chilling effect of ice melting directly permeates. We have taken 80 feet as the depth of the surface layer which is continually kept in turbulence by waves and local currents. Assuming that the melting of 1 cubic foot of ice will cool 80 cubic



REPRESENTATIVE TEMPERATURE CURVES OF WATER COLUMNS IN DAVIS STRAIT AND BAFFIN BAY

FIGURE 120.—The degree of solar warming of the surface layers, to a depth of 150 meters or less, in various parts of Davis Strait and Baffin Bay, is shown on the shaded areas above. The temperature Centigrade is recorded as the abcissa and the depth in meters as the ordinate. The greatest increase in temperature, as naturally should be expected, occurs in the southern stations of this region, such as 1125 and 355, while the least affected are Baffin Bay stations, such as 100 and 1015. The geographical position of the stations is shown on Figure 121, p. 200.

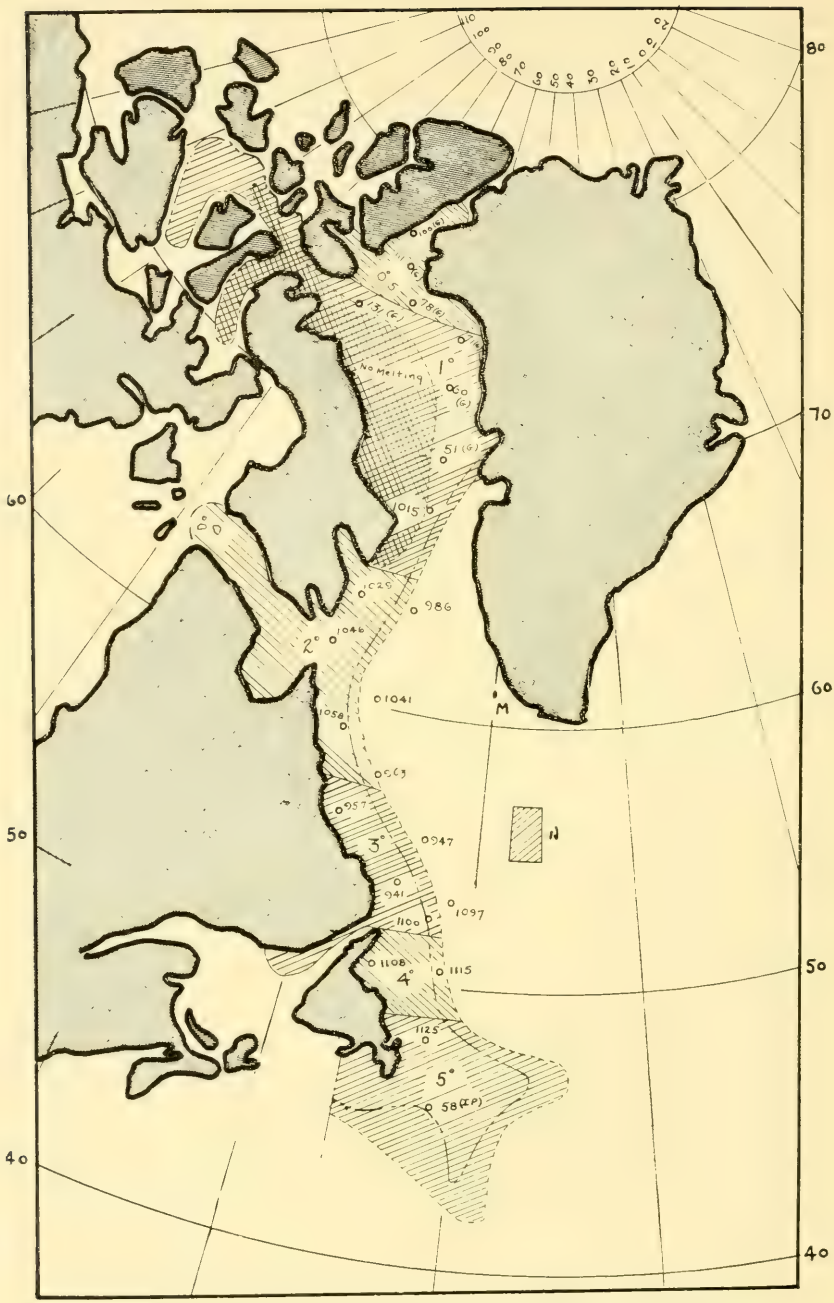
feet of water 1° F., we have, in the case given, 80 feet of water to 6 of ice, or the 467 cubic miles of pack ice in the western North Atlantic (shown shaded in fig. 121), will counteract the warming of the 80-foot surface layer of the sea a total of 6° F.

Let us now investigate the amount of solar warmth that is absorbed by the water masses of the melting area in the western North Atlantic, beginning with the first rise in temperature of the southern section in early spring and continuing until the entire area attains its maximum heat absorption in late August or early September. The most accurate method of determining the amount of thermal absorption and the depth to which it extends is obtained by examining several temperature curves (see fig. 120, p. 198) for the water column of stations scattered throughout the melting area. *Marion* stations 941, 947, 957, 963, 1058, 1062, 1041, 1046, 986, 1029, 1015, 1100, 1097, 1108, 1115, and 1125 have been employed. (See Smith, 1931, for station table data of the *Marion* expedition.) Other stations examined were the *Godthaab's* Nos. 51, 60, 71, 78, 131, 111, and 100. (See Annually 1929.) Two ice-patrol stations were selected, 58 (see Nightingale, 1915, p. 54) and 355 (see Smith, 1924, p. 100). The position of these stations at which subsurface observations were made are plotted on Figure 121, p. 200.

The shape of the temperature curves (fig. 120) show that the sun during the summer months warms the uppermost 15 to 20 meters, the greatest and passing from that depth downward the seasonal gain in temperature decreases directly with the depth until the 150-meter (approximately 480 feet) level is reached, where more or less constant year-round conditions prevail. A comparison of the temperature curves, one with another, also shows that the farther northward we proceed the thinner and less heated is the sun-warmed surface strata. The thickness of the isolation layer has been taken as the limit to which the sun's heat has penetrated, or a layer having the horizontal dimensions of the melting area and a uniform thickness of 150 meters. The melting area as drawn on Figure 121 has been divided into six subdivisions, viz, A, B, C, D, E, and F. Division A being the southernmost area is most noticeably affected by the sun and the oceanographic stations in this region record a seasonal increase in temperature equivalent to 5° F. throughout the 480-foot layer. Division B warms up equivalent to 4° F. surface to a depth of 480 feet. Division C, 3°; division D, 2°; division E, 1°; and the most northern one in Smith Sound only 0.5° F. Combining these values of the degree of summer heat in the 480-foot layer, and spread uniformly over the melting area, we obtain a mean temperature rise of 2.1° F.

The annual melting of pack ice was found to counteract the warming of an 80-foot surface layer 6° F. The same agency will oppose the warming of a 480-foot layer 1° F., or, in other words, *the chilling of the northwestern North Atlantic as a result of ice melting amounts to approximately one-half the solar heating* this same region receives during summer.[91]

[91] Ricketts (1930, p. 109) estimates that the total chilling effect of 1,350 icebergs that drifted south of Newfoundland in 1929 was not sufficient to nullify more than two hours of the average vernal warming to these regions. The relative insignificance of icebergs cooling the North Atlantic agrees well with the very small proportion that annual volume of glacial ice bears to annual volume of sea ice.



THE ICE MELTING AREA IN THE WESTERN NORTH ATLANTIC

FIGURE 121.—This figure has been employed to express quantitatively a number of comparisons between the ice-melting effect and other related phenomena. The shaded area bounded by the full line is the normal pack-ice area. The dotted line boundary marks the mixing zone. Stations 1125, 1097, 1041, etc., are selected points of subsurface temperature observations mentioned in the text and shown graphically in Figure 120. The entire melting area, with a uniform thickness of 150 meters, is divided into six parts; the southernmost in summer being heated an average of 5° F., the northernmost only 0.5° F. The spot "M" off Cape Farewell represents the annual crop of glacial ice expressed in the same scale as the pack ice and as one large berg. The shaded area "N" represents the total annual discharge of glacial ice into Baffin Bay, expressed to scale and in terms of pack ice 6 feet thick.

The heat from the sun during the warmer months of the year heats a greater mass of water than is shown in the case cited above, because the cold current is continually bringing cold water from a far-northern unheated source. It is assumed that the current has an average rate of 7 miles per day, so that during the period of summer the surface layers of the northern half of the melting area will have been completely replaced, partly by water from west Greenland and partly by still colder masses from the waterways of the Arctic Archipelago. The total amount of heat received by the melting area might, therefore, be conservatively increased 50 per cent more than what is shown.

The fact that nearly all of the ice tends to remain in coastal waters from formation to dissipation also allows expression of this cooling factor of the North Atlantic Ocean masses in terms of the two cold currents—the east Greenland and the Labrador. The low temperature of these is due chiefly to (*a*) back radiation to the air, accompanying the diminution of solar radiation during winter and (*b*) the chilling effect of melting ice.[92] If we could determine the respective volumes of ice and water that are carried annually out into the North Atlantic by the cold currents, we would obtain an approximate quantitative estimate of the relative importance of each factor. The processes of mixing between the frigid coastal waters and the much more temperate oceanic masses, takes place along the entire continental edge, but in some places to a greater degree than others. The only sector for which quantitative information is yet available is the discharge of the Labrador current as it flows southwestward past the Tail of the Grand Bank into the side of the Gulf Stream.

According to the data collected by the international ice patrol, the volume of this discharge at its maximum is about 62.4 cubic miles per day (5,000,000 tons per second). The basis for the stated value of 62.4 cubic miles is the observations collected by the ice patrol, stations 195 to 211, and 217, and 223 to 228, inclusive, May 3–30, 1922, around the Tail of the Grand Bank. (See Smith, 1923, pp. 74–80.) After calculating the dynamic potentials of these 23 stations according to Bjerknes' formulæ, a map showing the topography of the sea surface was drawn. (See Smith, 1926, p. 39.) At the time and place the Labrador current measured 68 miles in width from its inside edge in on the Tail to latitude 41° 55′, in longitude 50° 19′, and had a mean depth of about 600 meters. Its velocity was then the greatest on the surface and in its middle, least along its

[92] The chilling effects of melting snow and evaporating surface water are agencies which also withdraw significant amounts of heat energy from the northern waters. It is also well known that a great deal of evaporation takes place directly from the ice and snow, the latent heat per unit volume being about eight times that in the case of melting ice. Probably 75 per cent of the total annual snow that falls on northern coastal shelves and seas goes to cover the sea ice which has already spread over these regions, but the remaining 25 per cent is believed to fall in the water and thereby directly to cool the latter. In the Gulf of Maine, by Bigelow's (1927, p. 696) calculations the chilling effect of melting snow is equivalent to that of an ice cover 2 inches in thickness. In the Arctic, with roughly equal snowfall, its effect equals only about 2 or 3 per cent that of melting ice. Evaporation is most rapid with low atmospheric pressure, low humidity, high temperature, both of water and of air, and with high winds—conditions which in combination seldom obtain in northern regions—therefore, evaporation as a chilling factor is of small importance there as compared with lower latitudes. Even so, however, the effect is greater than that of melting snow, for in the Gulf of Maine, according to Bigelow's calculations, cooling by evaporation in the course of a year is equivalent to the cooling effect of an ice cover 2 feet in thickness. Thus, assuming equal evaporation in the North, the chilling effect of evaporation would then equal about 30 per cent of that of the melting of ice.

sides and near the bottom. If the current map of May, 1922, be compared with others constructed in 1926 and 1927 (see Smith, 1927 and 1927b), we find it representative of normal conditions during spring when the Labrador current is believed to be most voluminous. The opposite extreme, when there is only a weak flow of the cold current south of Newfoundland is represented by observations of the ice patrol made October 21–26, 1923 (see Smith, 1924, pp. 151–164); February, 1922 (see Smith, 1923, p. 94); and even June 9–25, 1927 (see Smith, 1927b, pp. 87–93). At such times it is estimated that the discharge past the Tail of the Bank may fall to 10 cubic miles or so per day (0.8 million tons per second). As pointed out on page 156, the flow and volume of the Labrador current shows wide fluctuations. The current may not only show a seasonal variation within the year, but at the same time alter irregularly in velocity and size over periods of a month or even less. With maximum and minimum rates, in terms of daily volume of about 70 and 10 cubic miles, respectively, a value of 40 cubic miles (3.2 million tons per second) may be assumed to be approximately the year round mean. According to this figure the total yearly discharge is, therefore, something like 14,600 cubic miles.

The mean temperature of the cold current in cross section, which we have calculated above, is 37° F., and it discharges south of Newfoundland into the North Atlantic at latitude 43°. The normal surface temperature for the latitude at which the Labrador current discharges and mixes with the ocean masses off the Tail of the Grand Bank is approximately 54° F.[93] The drop in temperature from the surface to a depth of 600 meters (the thickness of the circulatory layers) is about 9° F. The mean annual temperature of this 600-meter ocean layer is therefore approximately 47° F. This value not only includes the chilling effect of the Labrador current in the vicinity of the forty-ninth meridian but it also takes into consideration the warming effect of the Gulf Stream farther eastward in the same latitude and therefore the two effects are more or less counterbalanced in the temperature value of 47° F. Assuming the average temperature of the Labrador current is 37° F., the mean annual cooling effect of this discharge off the Tail of the Grand Bank is therefore 10° F. Consideration of the melting ice volume of 467 cubic miles and annual volume of Labrador current discharge of 14,600 cubic miles indicates that the cooling effect of the latter would counteract an amount of ice about four times as great as normally is melted in the northwestern North Atlantic; or to put it another way, the branch of *the Labrador current which discharges at the Tail of the Grand Bank has four times the cooling effect as that of the annual melting of Arctic ice for the western North Atlantic.*

The *Marion* expedition measured a section across the continental shelf opposite Hamilton Inlet and from a preliminary calculation of these data the cubical daily transport of the main trunk of the Labrador current is approximately 130 cubic miles (10.5 million tons per second). If we assume, therefore, that the branch of the Labrador current that passes the Tail of the Grand Bank amounts

[93] Krümmel (1907, p. 400) has calculated the mean annual surface of ocean water for both north and south latitudes. At 43° north latitude it is 12° C. (53.6° F.).

to about one-third of the cold discharge between Cape Race and Flemish Cap, then *the cooling effect of ice melting amounts to only 10 per cent of the total cooling that the Atlantic receives from the cold water masses of the Labrador current.*[94] The fact that the relative proportions of ice and water are roughly similar at the other principal point of discharge—the east Greenland current—makes this figure applicable to the entire North Atlantic. Obviously, the low temperature character of *the Labrador and east Greenland currents* is not due to the melting ice with which these streams are charged *in spring and summer. These ocean currents are cold because of the small amount of absorption of solar radiation at the earth's surface in the polar regions.*

The foregoing corroborates Campbell-Hepworth (1912, p. 4), that the ice in the western North Atlantic is not the agency chilling the surrounding ocean masses but it is largely the Labrador current. Helland-Hansen-Nansen (1920, p. 266) conclude that the variations in temperature noted in the unperiodic coolings of the North Atlantic Ocean are due to the changes in the distribution of air pressure, i. e., the northerly winds. Schott (1904, p. 279), in the course of remarks on the intensification of the Labrador current in 1903, adds that the cooling effect of Arctic ice in the Atlantic is negligible. While our own researches do not find that the ice-melting effect is negligible, they do prove it is of secondary magnitude.

It is interesting to note in the quantitative treatment of these northern seas phenomena that the cooling factors, viz. chilling by the winter atmosphere, ice melting, snowfall, and evaporation, when totaled, outweigh the solar warming of summer. We have found that the ice-melting effect counteracts about one-half that of seasonal solar warming, and mounts to approximately one-tenth the cooling effect of the annual discharge of the cold currents. The great major warming effect therefore tending to maintain more or less of a constant counterbalance over a long period is the warm currents from the Tropics. We conclude that *the major controls of the thermal equilibrium of the North Atlantic Ocean are the ocean currents, while the direct action of vernal warming in the ice areas themselves, and the opposite effect of chilling by Arctic ice, are secondary.*

The fact that the latent heat of melting ice as a chilling agent amounts to only about one-tenth that of the cooling effect of the ocean currents from the north also proves conclusively that melting ice is far from being an important factor assisting to maintain the present scheme of oceanic circulation. According to Pettersson (1904, 1907), the melting northern ice, which we have shown takes place almost entirely in relatively shallow waters, sets up a certain system of oceanic circulation. The waters of the Grand Bank, however, even when no ice is present, have been found to exhibit a similar

[94] Huntsman (1930), in discussing the Arctic ice which is brought into the Gulf of St. Lawrence via the Strait of Belle Isle, states that in his opinion its melting is the chief cooling agent in the Gulf. He points out that if the branch of the Labrador current that enters the Gulf has a velocity of 1 knot per hour for the three winter months, enough ice might be carried in to cover half the Gulf of St. Lawrence. He also adds that were this ice to melt it would chill by 8° C. a 30-foot layer of water equal in area to the entire Gulf. As a matter of fact, however, the inflow through Belle Isle is not believed to maintain such a high rate as assumed in the above example. It is estimated moreover, that the Gulf of St. Lawrence normally cools 8° each year throughout a mean depth of about 300 feet, and if this be a fact, the chilling effect of the ice importation amounts to less than one-tenth that of the other agents. Here, apparently, as in the Atlantic, the cold-water masses are the deciding factors.

type of circulation. (Smith 1924a, p. 133.) A similar type of circulation has been observed by Mathews (1914, p. 32); by Huntsman (1924, p. 280); and by Bigelow (1927, p. 921); it is also in fact a characteristic feature of practically all of the coastal waters along those parts of the northeastern American shelf that are boreal in character but ice free at the time. We must conclude from the above that the oceanic system of currents, at least in the higher latitudes of the northern seas, characterized by offshore expansion in the surface layers and a salty indraft beneath is set in motion not by the ice but by the belts of water around the sides of the basin, oceanographically known as "coastal."

Perhaps it is because the ice catches the eye and the imagination more than do the coastal water masses with which it is ordinarily

THE COLD WALL OFF THE AMERICAN COAST

FIGURE 122.—A remarkable photograph taken March 27, 1922, south of Newfoundland, in latitude 41° 40′ N., longitude 51° 07′ W., looking westward along the meeting zone of the Gulf Stream and the Labrador current. The surface water on the cold side of the wall (to the right) with a temperature of 34° F. was smooth and glassy. The warm water (to the left) with a temperature of 56° F., choppy and rippled. There was a range of 22° F., therefore, in less than a ship's length. This important oceanographic boundary marks a critical point in the disintegration of the icebergs that drift out of Davis Strait into the western North Atlantic. (Official photograph, international ice patrol.)

associated that its relative importance in the picture of oceanographic circulation has been overemphasized.[95] The regional difference of density between coastal and oceanic waters is the main springs for the convectional currents. The winds, also, by their direct frictional effect, combined with the presence of coast lines or other hindrances, develop significant slope currents. In any case, however, we can not refer the impelling forces to a riverlike source; in fact, they work along the entire extent of flow.

The transition zones, i. e., the continental slopes, the continental edges, and the ridges, mark the belt of greatest energy, and in the

[95] Pettersson (1927, p. 7) has even stated that the force that deflects the Gulf Stream towards the north and west, against the effect of earth rotation, is to be explained as an effect of northern ice.

sea energy is synonymous with current. The smooth curving trend of the eastern North American slope, from Florida to Nantucket, for example, is responsible for the riverlike character of the Gulf Stream described in such striking terms by Maury. The Gulf Stream parallels the coast of Norway, because a dynamic gradient prevails between light coastal waters and the off-shore oceanic water there also. The energy that is imparted in moving the water particles of the Gulf Stream off Sognefjord, Norway, is just as vital for the survival of the current as is the impetus that was given to these same particles during their sojourn in the Caribbean. It has been estimated that the forced momentum of the current pouring northward out of the Florida Strait would die one-quarter the distance to Cape Hatteras were it not sustained by gradient forces. When the Gulf Stream, on its inner side, after passing the Grand Bank, bends to the northward (in toward Newfoundland), it does so because at that time and place there exists a contrast in the density of the oceanic masses. (See Defant 1929, p. 16.) Ice melting over the North American and east Greenland shelves helps to accentuate the contrasts between coastal and oceanic waters, thereby intensifying the currents, but emphatically it is not the main cause of propulsion nor is it even a necessary attribute thereof.

In conclusion the energizing forces resulting from melting ice are not adequate to produce the circulatory mechanism of the world's oceans; nor is the water directly chilled by melting ice—the chief source of the ocean's bottom water.

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