VOLUME III
to
TECHNICAL REPORT SRC-71-TR-N3501
15 August 1971

ARCTIC OPERATIONS TECHNICAL NEEDS (U)

Appendix B
ENVIRONMENT OF THE ARCTIC

by
J.E. Sater
THE ARCTIC INSTITUTE OF NORTH AMERICA

prepared for
Director, Naval Analysis Programs
Naval Applications and Analysis Division
Office of Naval Research
Arlington, Virginia 22217

IN ACCORDANCE WITH CONTRACT No. N00014-71-C-0305

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FOREWORD

This report describes the environment and character of the Arctic Basin and its near-shore areas as well as those of the arctic sectors of the North Atlantic and North Pacific and their shores. It is drawn from the available literature and represents a synthesis of the writings of many experts in many related fields. The material presented is at best indicative of the conditions as they are known to exist. Subjects which lend to small scale-large area treatment are presented at length. Those which require large scale-small area treatment are presented more briefly and the interested reader should refer to the materials listed in the bibliographies. In addition to these materials, the reader is directed to works such as the Arctic Bibliography, the CRREL Bibliography and U. S. Government Research and Development Reports as well as journals such as Journal of Geophysical Research, Polar Record and Arctic.

Throughout the report, all distances stated as miles are nautical miles, not statute miles.

The material in this report volume will be published in book form at a later date by the Arctic Institute of North America. Recipients of this report will receive copies of the book at that time.

This report is an update of a previous report, Environment of the Arctic Ocean and the Rimland, Volume III, 18 December 1970, by J. E. Sater for Systems Research Corporation under Contract N00014-70-C-0323.

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B.1 INTRODUCTION

B.1.1 Definitions

In order to analyze the environment of the Arctic Basin it is first necessary to delineate its physical form or boundary. However, as a result both of remoteness from most population centers and of adverse climatic conditions, knowledge of the northern portion of our planet is still limited and has only partially been collated and verified. This circumstance has led to the discrepancies within the vocabulary which denotes and delimits various aspects of the region. A single expression often may have widely differing meanings. An excellent example is the phrase *Arctic Basin* (often referred to as *Polar Basin*). To many persons, especially to sailors and oceanographers, the term refers to the deep basin which is filled by the Arctic Ocean (also called Polar Ocean or Polar Sea), whereas to others it indicates the drainage basin in which water flows northward into the Arctic Ocean.

At the same time, the forces which act upon the Arctic Basin and which modify its internal conditions must also be comprehended, because a number of these, e.g., winds and currents, have their origins external to the Basin. Thus, the discussion will at times expand to include as well the Arctic, or the lands immediately surrounding the Arctic Basin. It follows that there must also be some objective discussion about the meaning of the word *arctic* and about other associated terms, because unlike the Arctic Basin, the Arctic does not appear to have an immutable boundary.

One certain characteristic of the Arctic is the cold. The term cold, unfortunately, may also be interpreted in a number of ways: it may mean extremely low temperatures (-40°F or lower) at which machines become relatively difficult to operate; or it may mean physiological cold - the combination of wind and low temperature which produces rapid heat loss from the body, generally referred to as wind-chill. It also may refer to the temperature at which plant growth ceases (near 32°F), or it may mean the temperature which permits the growth of only the relatively impoverished tundra vegetation as compared with the more complex plant societies of the boreal forest. The upper limit of this temperature band forms a boundary line of considerable cultural significance as well because it marks the limit of agriculture, even the restricted agriculture of the taiga. Here, in fact, cold actually means only one thing - the heat deficit state of the atmosphere.

Climatologists have advanced a variety of climatic definitions to establish boundaries for the Arctic. These definitions are based for the most part on the relationship between atmospheric temperatures and the effects temperatures have on the ecology or, in some cases, the geomorphology. By establishing as arctic regions those areas in which the average temperature of the warmest month is below 50°F, Koppen suggested that the boundary of the Arctic coincides closely with the tree line. Furthermore, tundra regions were classified by him as those in which the average temperature of the coldest month is 32°F or below. In a second system, 43°F is a more significant plant growth temperature than is 32°F. On the basis of this criterion, the barren land or arctic regions were defined as those in which the mean monthly temperature throughout the year is less than 43°F while the tundra was taken to be that area in which the mean temperature is above 43°F for one to two months.

Nordenskjold's formulation of an arctic limit - also intended to coincide with the tree line - included those regions in which the average temperature of the warmest month plus one-tenth of the mean temperature of the coldest month is 51.4°F, which suggested that the winter temperatures also have an effect, although slight, on the position of the tree line. However, it should be noted when considering these definitions that the tree line itself is in a state of flux and is not an ideal reference.

In terms of military operations, the U.S. Army defined the Arctic Operations Area as the area in which temperatures can be expected to reach -40°F. According to such a delimitation - which includes most of Canada and some parts of the United States - the significant characteristic of the so-called arctic area is its low temperature, whether of long or short duration. The Army's definition of the Arctic is: that portion of the northern hemisphere characterized by having an average temperature of the coldest month of less than 32°F and an average temperature of the warmest month of less than 50°F. The subarctic is defined as that portion of the northern hemisphere characterized by having an average temperature of less than 32°F during the coldest month and average temperature above 50°F for one to three of the warmest months.

Geomorphologically it is possible to approximate an arctic limit using the permafrost zone as a criterion. Permafrost, or perennially frozen subsoil, is a phenomenon closely associated with the Arctic and is of particular ecological interest because of the relationship of low soil temperature to plant growth. Geologists and geomorphologists have described the Arctic as the area of continuous permafrost, an area somewhat broader, particularly in western Canada and eastern Siberia, than the region delineated by Koppen.

Because it is a peculiarly polar phenomenon, the auroral zone has also been advanced as a criterion for the geophysical definition of the Arctic. The aurora borealis, observed in high northern latitudes, is a luminous circumpolar phenomenon of the upper atmosphere. The center of auroral activity is located near Etah, Greenland, the point at which the geomagnetic pole intersects the earth's surface. From maps indicating the frequency and location of auroral observations, it appears that a specific line (isochasm) might easily be drawn to delineate the auroral or ionospheric Arctic.

Oceanographers consider the Arctic to be that region in which only pure arctic water (the temperature of which is at or near 32°F and the salinity approximately 30°/oo) is found at the surface. Arctic water is formed in the Arctic Ocean by a combination of: (1) water from the Atlantic and Pacific oceans, (2) water drained from the surrounding land areas, and (3) water resulting from the melting of sea ice. Using these criteria, one can draw a line north of which all surface waters can be considered as arctic waters and which can therefore be held to encompass the Arctic itself. That region, however, would be considerably less inclusive than those delimited on the basis of temperature, vegetation, permafrost, or ionospheric criteria in that it excludes land areas (Fig. I-1).

Thus, the problem of a definitive boundary for the Arctic is a complex one. In its symbolic aspect, the Arctic coincides with that region which surrounds the North Pole; in its literal sense, it refers to the area encompassed by the Arctic Circle. Its outer boundaries can, however, be established only within a specific context, and even then the delimitation is often arbitrary. In their specific forms the various delineations overlap and interplay. and rigorous definition is pointless. While its geographic and environmental features are peculiarly distinctive, the Arctic has no boundary that is immutable and it cannot, therefore, be conclusively defined.

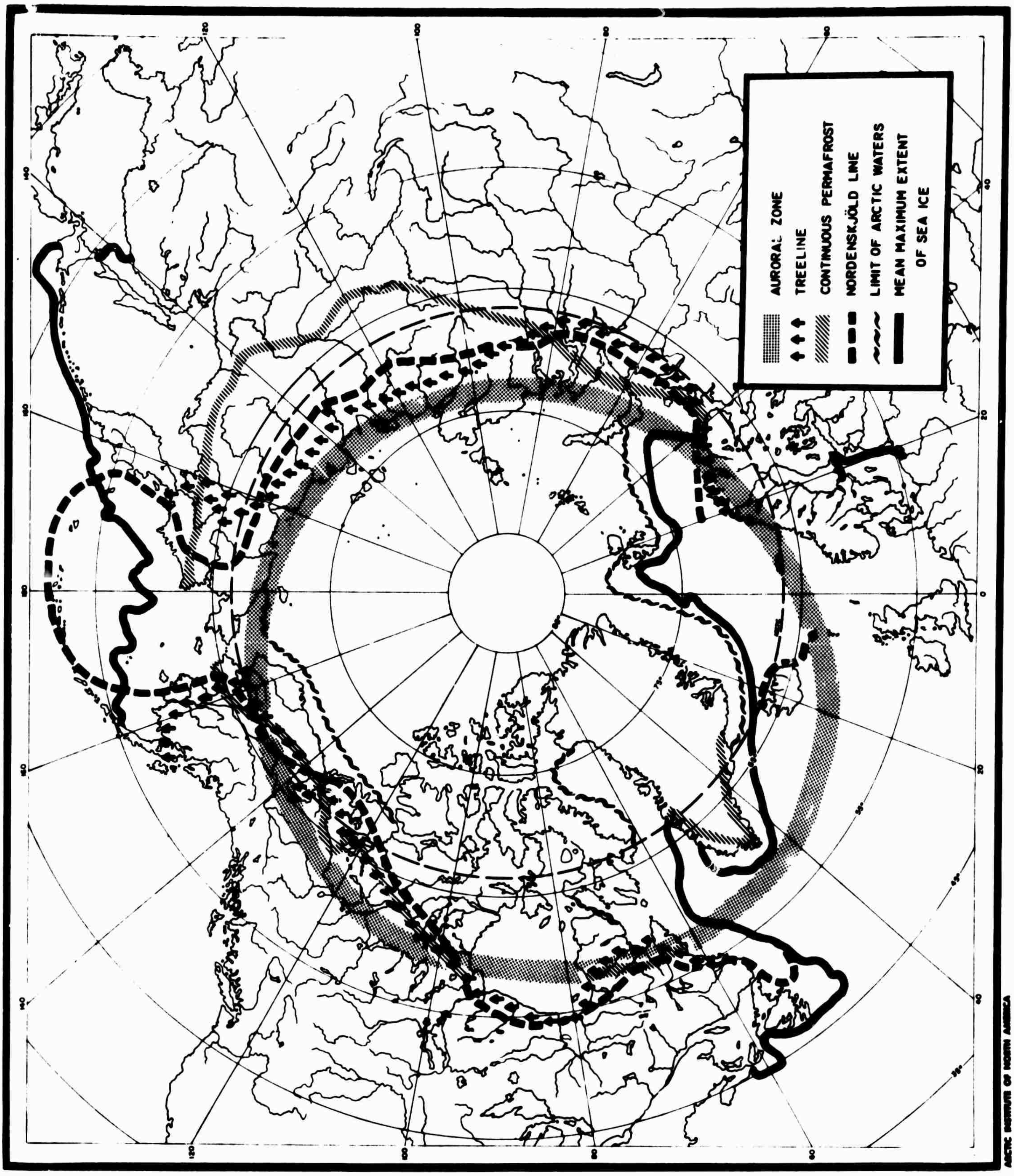


Figure I-1 ARCTIC REGIONS

In view of this, the Arctic must be considered as a region, not an area. For the purposes of this study it will be defined as the portion of the northern hemisphere in which sea ice may be encountered together with the coasts which experience it. The term *Arctic Basin*, in contrast, will be used to denote that basin which is filled by the Arctic Ocean and its peripheral seas, whereas the more inclusive drainage basin within which water flows northward into the Arctic Ocean will be designated the *Arctic Drainage Basin*.

B.1.2 Sovereignty

Today nearly all international misunderstandings and doubts regarding sovereignty over arctic land areas have been resolved. The so-called "sector principle" proposed in Canada and promulgated by the USSR in 1926, while neither renounced by them nor ever tested in litigation, is of doubtful validity and of no practical significance when accepted criteria of sovereignty already obtain, as they do in most instances. This principle would permit a nation bordering on the Arctic Ocean to claim sovereignty over lands within that segment of the ocean lying between the easternmost and the westernmost meridians bounding its territory to the point where those meridians meet at the North Pole. The principle has served *de facto* rather than *de jure*, for a short period, to delimit areas of sovereignty of the respective countries over land areas north of the mainland. However, Canada now claims sovereignty on the basis of exploration, occupancy, and administration; and her claim is recognized over all islands except Greenland north of the mainland of North America between longitude 62°W and 141°W. Under the same pattern such islands and island groups as Wrangel, Zemlya Frantsa Iosifa, Novaya Zemlya, and Novosibirskiye Ostrova fall in the Soviet sector and are not disputed. Overall, the Soviet Union controls approximately the same amount of the territory bordering the Arctic Ocean as do the United States, Canada, Norway, and Denmark combined.

In 1926 the Soviet government published a decree claiming all land, including ice formations that are more or less immovable, and all islands in the triangle that lay between the meridians of 32°4'34"E and 168°49'30"W and had as its apex the North Pole. The eastern islands of the Svalbard Archipelago and Little Diomede that lie within this triangle were accepted as belonging to Norway and the United States of America, respectively. In articles in 1928 and again in 1950, the Soviet press advanced claims to the open polar seas, including drifting ice; and the entire triangle has long been shown on USSR maps as constituting Soviet territory. While the USSR has not made her official position wholly clear as regards the arctic seas north of her mainland, it is significant that she has not demonstrated a proprietary interest in the area defined in the decree of 1926 despite numerous incursions into the area by other nation's vessels and aircraft. In a separate claim the USSR defined as territorial waters a 12-mile zone off her entire coast, including islands.

Greenland became an integral part of the Danish State in 1953. Jan Mayen was occupied and claimed by Norway in 1929. Svalbard was awarded to Norway in 1925 by an international treaty, which at the same time restricted her sovereignty in several respects. Military installations are forbidden, and in the case of Vestspitsbergen there is a somewhat complicated dual-occupancy arrangement with the USSR for the exploitation of the Longyearbyen coal.

Two other situations should be considered: ice floating on the sea and ice on the sea that is attached to land. It has now been demonstrated clearly by both the USSR and the United States that it is possible to establish relatively large camps for long periods both on the sea ice of the Arctic Ocean and on the much thicker ice of the ice islands that are broken from land-fast ice and are adrift in the ocean. In spite of the ability to operate from these floes and islands, such operations have not led to the assertion of claims to sovereignty over any area of ice or over any specific geographic location. They appear, rather, to be considered by their occupants to be analogous to ships drifting without control. A similar position seems to have gained general acceptance in regard to shelf ice, the ice that is attached to the land and extends to sea beyond the generally accepted limit of territorial waters.

Because of the possibility of petroleum deposits and other mineral resources lying under the continental shelves of the Arctic Ocean and for other reasons as well, there is considerable interest in learning more about these areas. In this connection the agreements reached a decade ago at the International Conference on the Law of the Sea at Geneva are of interest. The right of a nation to develop minerals on the bordering continental shelf to the outer practical limit of development was not held to be of sufficient practical importance to justify special regulations. Thus the ownership of these resources is still not established and the question is likely to be raised in future international discussions.

B.1.3 Daylight

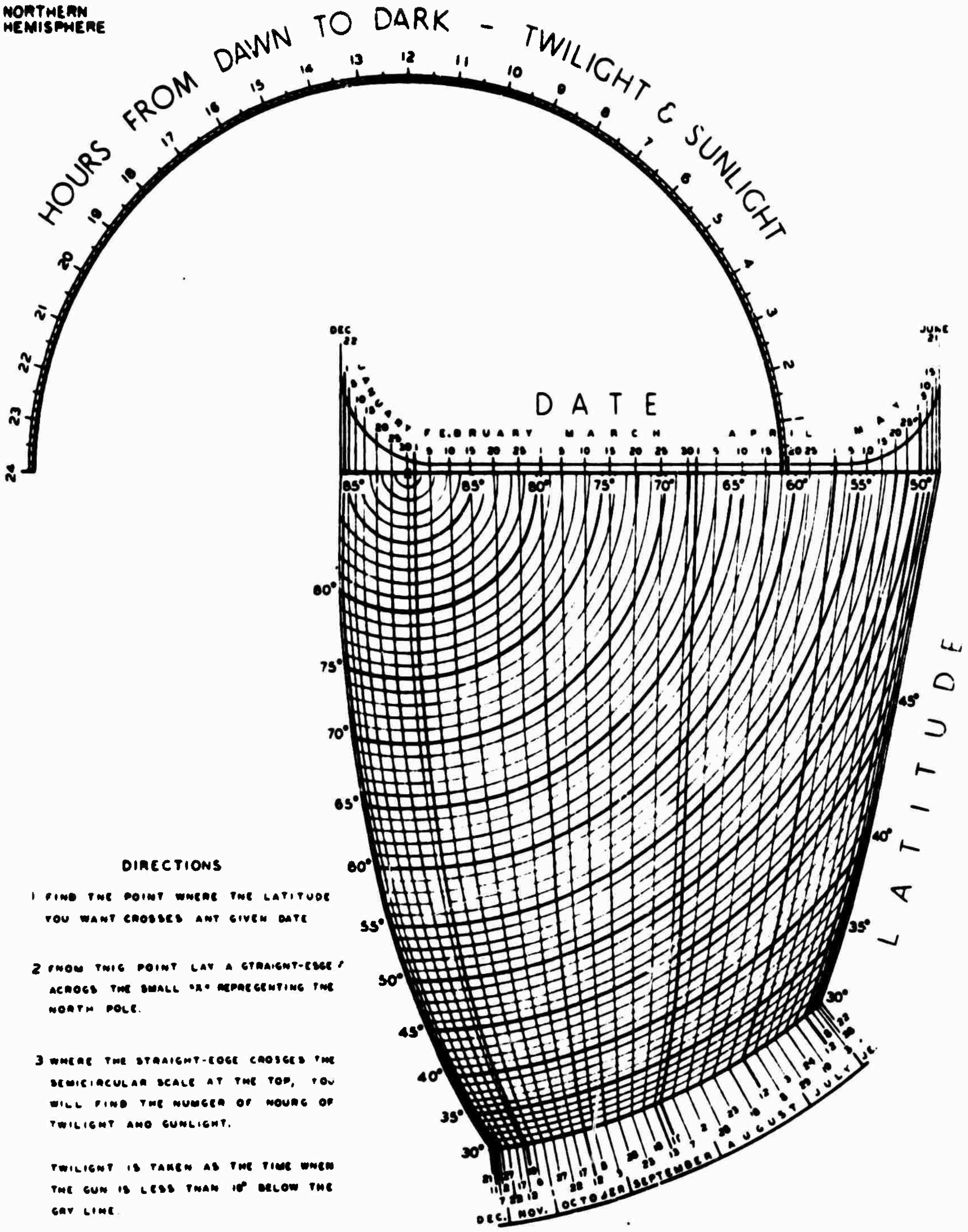
The "long darkness of winter", more than anything else except perhaps the fear of cold, has prevented people from going to the Arctic. Actually, it is only at the Pole that the extreme condition exists; and even there the sun shines for about 28 weeks each year and remains invisible for the remaining 24. Moreover, the season of "useful light" is much longer than this would indicate. If daylight is defined as the amount of light sufficient to enable one to read newspaper print out-of-doors under a clear sky, there are 32 weeks of continuous daylight at the Pole and over 8 weeks more during which there is at least some twilight all the time. This leaves only about 80 days of real night. At the

North Pole there are actually about 140 more hours of sunlight each year than at the equator. On the Arctic Circle there are about 230 more. The following figures (I-2 and 3) may be used to determine the various periods of light experienced in the Arctic.

B.1.4 Gravity

Observations of the strength of the earth's gravitational field are used both to determine the precise shape and size of the geoid and to establish the nature and thickness of the crustal materials. The principal goal of geodesy has been to determine the parameters of an ellipsoid of revolution that best fits the figure of the earth as a whole. The resulting determinations are now converging to a point beyond which there will be little change, and attention is being directed to determination of the earth's actual shape, or the relationship between the ellipsoid and the geoid. Different types of geodetic analysis may be used. A classical method involves astronomic observations for latitude and longitude which, in connection with geodetic positions, determine the deflection of the vertical or the slope of the geoid at the point in question. A second method involves observations of gravity at more or less uniformly spaced points over the area of concern. Many gravity measurements have been made in the principal countries of the world and across the oceans, particularly in the middle latitudes, but there is a paucity of such data from the Arctic.

In Alaska a continuous scheme of triangulation extends from the British Columbia-Alaska boundary through southeastern and southwestern Alaska, to the western extremity of the Aleutian Islands, to the shoreline and the islands of the Bering Sea to the Arctic Ocean, to the international boundary near the Demarcation Point. The elevations above mean sea level of established bench marks have been determined by lines of precise levels along the Alaska Highway



DIRECTIONS

1 FIND THE POINT WHERE THE LATITUDE YOU WANT CROSSES ANY GIVEN DATE

2 FROM THIS POINT LAY A STRAIGHT-EDGE / ACROSS THE SMALL "X" REPRESENTING THE NORTH POLE.

3 WHERE THE STRAIGHT-EDGE CROSSES THE SEMICIRCULAR SCALE AT THE TOP, YOU WILL FIND THE NUMBER OF HOURS OF TWILIGHT AND SUNLIGHT.

TWILIGHT IS TAKEN AS THE TIME WHEN THE SUN IS LESS THAN 18° BELOW THE SKY LINE.

Figure I-2 HOURS OF TWILIGHT AND SUNLIGHT

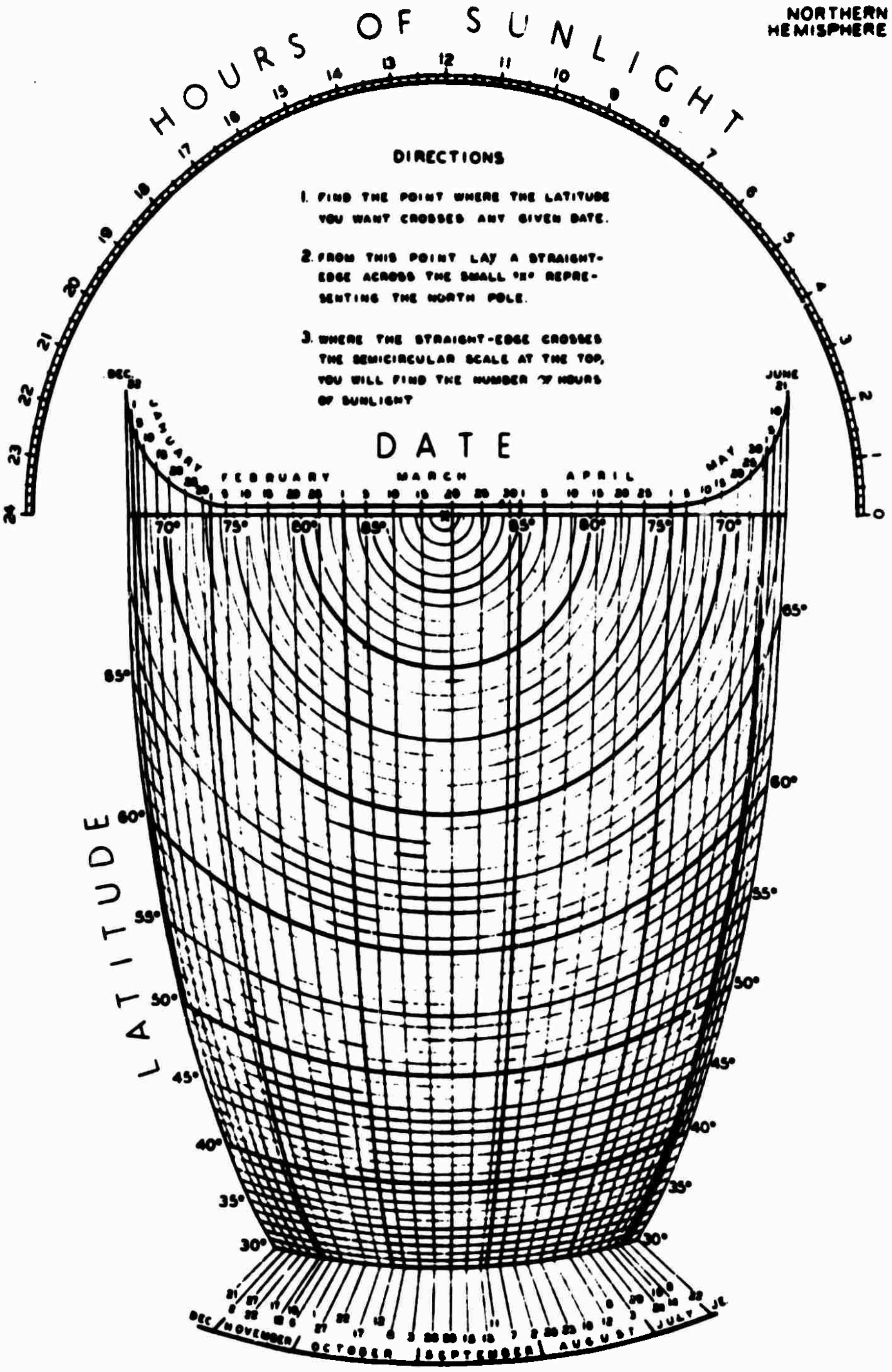


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Figure I-3 HOURS OF SUNLIGHT

from Whitehorse to Fairbanks, on the Richardson Highway from Fairbanks to Valdez, and on the Steese Highway to Circle.

The main triangulation net of arctic Canada is tied to the Alaskan network at the boundary near the Alaska Highway. A loop has been completed north of 60°N from the vicinity of Great Slave Lake to east of Fort Reliance. From the latter point an arc has been completed to close the loop east of Lake Athabasca. Another loop has been extended north from Great Slave Lake to a connection with Victoria Island. A triangulation network using short-range navigation by radar (SHORAN) covers all of northern Canada. SHORAN points have been established from the Alaska boundary to the Atlantic coast and north over the Canadian Arctic Archipelago to Ellesmere Island and northwestern Greenland. A tie between the geodetic surveys in North America and Europe was completed by HIRAN trilateration across the North Atlantic. This network extends from Baffin Island, across southern Greenland to Iceland, to the northern British Isles and from there to the coast of Norway. The European data were thus carried from the three Norwegian triangulation stations through the trilateration net to Canada: the North America data were, at the same time, carried from the Canadian SHORAN stations to Norway. The astronomic positions of the points in the HIRAN net have been determined.

It is known that the European data have been carried through Czechoslovakia on to the USSR (the Krassowsky triaxial ellipsoid) and it can be assumed that in this manner the Alaska-Canada-Greenland-Norway triangulation and trilateration net has been carried to its arctic coast. There is no tie across the Bering Strait. After the worldwide geodetic satellite triangulation network is completed, these various surveys in high latitudes can be interconnected. Even without the geometric results, considerable geoidal information has been obtained from dynamic studies of satellites with Doppler techniques.

Gravity observations can be obtained rapidly with airborne gravity meters. The data collected by satellites over areas not readily accessible by other means are of a generalized nature, however, and can determine only the long-period undulations of the geoid, not the detailed undulations that may be arrived at through work on the surface of the earth. A sea-gravity meter has been developed for observations from a surface vessel. In dealing with the problems of the Arctic, there is a need for a greater density of gravity observations, both on land and sea; such observations should be accomplished with the support of aircraft and surface vessels or submarines. For the proper interpretation of these observations, geodetic positions to about one kilometer and elevations to about five meters must be known.

The published data on arctic gravimitry, such as that compiled at the University of Wisconsin, are at best spotty (see Figure I-4). Other western nations have published results of traverses over parts of the oceans dealt with here but a complete picture is lacking. Presumably the most extensive records to date are those collected on the cruises of the U.S. Navy's nuclear submarines across the Arctic Ocean, but the data are not published. A lately published report discusses the most recent studies of the Wisconsin group. The observations of gravimetric forces continues on Fletcher's Ice Island, T-3, but the station has been relatively stationary for some time.

B.1.5 Magnetic Phenomena

The magnetic field measured on the earth's surface originates from three sources: (1) the self-exciting dynamo action within the earth's liquid core that produces the main magnetic field (2) the magnetic induction of rocks within the earth's crust and (3) an external and rapidly changing field generated by solar effects that disturb the ionosphere and give rise to a number of related

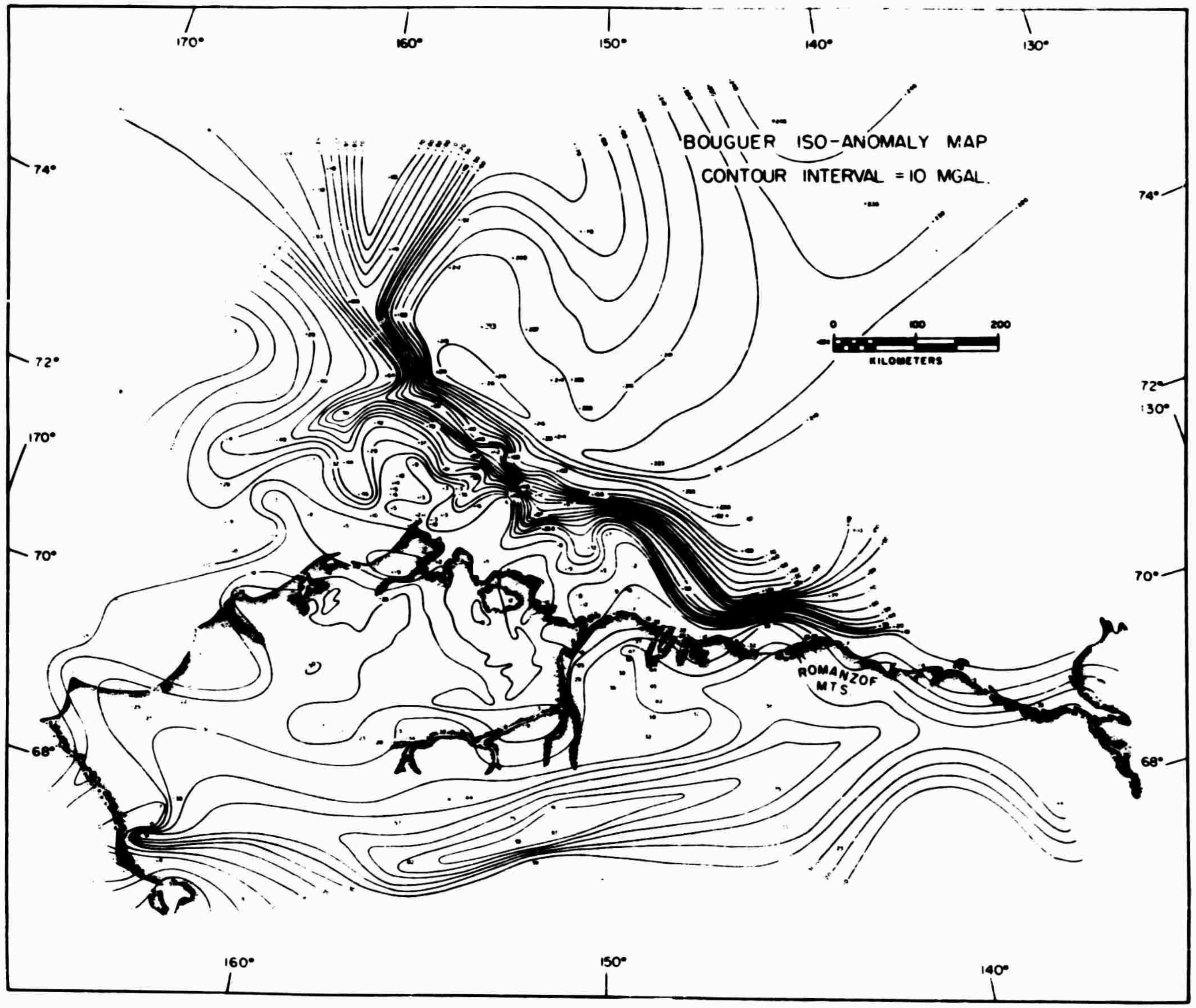

Figure I-4 BEAUFORT SEA GRAVITY

upper-atmosphere phenomena such as magnetic storms and aurorae. All of these magnetic fluxes have important consequences for operations in the Arctic.

The earth's magnetic field is almost entirely of internal origin. It can be fairly closely represented by a small number of harmonics; and the first-order harmonics, much the most important, have a physically simple description. They correspond to the potential of a small magnet or dipole located at the center of the earth and inclined at a small angle to the axis of rotation, and they produce the main dipole field. This dipole field is a useful simplification of the mathematically described geomagnetic field, which is computed by a harmonic analysis of the magnetic intensity measured at many locations on the earth's surface.

The poles of the dipole are called the geomagnetic (or dipole) poles. They should not be confused with the magnetic (or dip) poles, whose changing geographic locations on the earth's surface are influenced by the composite of all magnetic sources. Even though the dipole field differs from the magnetic field actually observed at the earth's surface, the difference diminishes rapidly with elevation. Thus those concerned with operations on or near the earth's surface must concern themselves with the complex total magnetic field, whereas those studying and working in the earth's outer environment are well served by the dipole field. In correlation studies and theoretical analyses the magnetic field has not been widely used, but its acceptance will grow now that the electronic computer makes the use of the actual field coordinates practical. In addition, new coordinates, such as the magnetic shell parameter, have been invented. They describe the motion of particles trapped in the magnetic field, and their principal use is in studies of the interaction of particles in the ionosphere.

Being mathematically determined, the dipole poles are diametrically opposed and their projections to the earth's surface are at approximately 78.5°N 69°W, near Thule, Greenland, and 78.5°S 111°E, near Vostok, Antarctica. The geocentric dipoles are inclined at an angle of 11.5° to the earth's rotational axis, an angle sufficiently small for the earth's rotation to be considered to play a dominant role in the creation of this magnetic field. Contrastingly, the magnetic (dip) poles are not diametrically opposed, each being about 800 miles from the point antipodal to the other. The line joining the two dip poles misses passing through the center of the earth by about 400 miles. The horizontal intensity can also vanish at other points within small areas on the earth's surface because of the magnetic susceptibility of crustal rocks. Such local north and south poles are not to be confused with the main dip poles that affect the compass on a global scale.

The earth's magnetic field is a vector quantity, having both magnitude and direction. The orientation of the total intensity vector (F) is defined by the angles of inclination or dip (I) and declination (D). Inclination is measured relative to horizontal, i.e., tangent to the earth's surface, and is downward in the northern hemisphere and upward in the southern hemisphere. Declination is measured in degrees of arc east or west of geographic or true north. Mariners commonly refer to declination as a variation of the compass. This value constitutes an important correction for navigation. The horizontal and vertical components of F are denoted by H and Z, respectively. H is always considered positive, whatever its direction, whereas Z is considered positive downward and thus has the same sign convention as I. The various components of the total field vector are geometrically relatable by the equations

$$H = F \cos I, \quad Z = F \sin I, \quad \text{and } \tan I = Z/H.$$

By definition the dip poles are at the sites where, on the surface of the earth, the total field vector is vertical ($H = 0$, $I = \pm 90°$). At these sites a freely suspended compass needle points straight down; hence they are commonly referred to as dip poles. The magnetic (dip) equator is the circumglobal line along which F is tangent to the earth's surface ($Z = 0$, $I = 0°$). Strictly speaking, the magnetic field strength is measured in fractions of an oersted but is commonly referred to in terms of the gauss, which is the unit of magnetic induction. Another frequently used and often more convenient unit is the gamma, which is 100,000 times smaller than an oersted. The maximum value of the earth's total field is just over 0.7 oersted and occurs in a region that includes the southern dip pole. The field at the northern dip pole is considerably smaller, being less than 0.59 oersted.

The northern dip pole is now located at approximately 75°N 98°W in northern Canada. From it a strong ridge of high horizontal magnetic intensity reaches across the Arctic Basin into Siberia, seriously distorting the magnetic meridians in this region (see Figure 1-5). The maximum value of the total field intensity (over 0.6 oersted) occurs not at the principal dip pole, but at a secondary focus in Siberia. Locally over the earth's surface (e.g., above some outcropping iron ore bodies in central and northern Sweden and in places in Norway) the magnetic intensity is as great as 1 to 4 oersted, but these are exceptional conditions. Elsewhere the varying rocks of the earth's crust can cause local distortions of the main magnetic field. These distortions, or anomalies, can amount to as much as 10% of the normal field strength. Also the oceans, because they are floored with basalt, have an associated magnetic field that is particularly disturbed and may produce rather large anomalies.

There is a slow change in the magnetic field with time, now called the secular variation, which is observed in all the components. If successive annual mean values of a magnetic component are obtained for a particular station, it is found that the changes are in the same sense over a long period of time, although

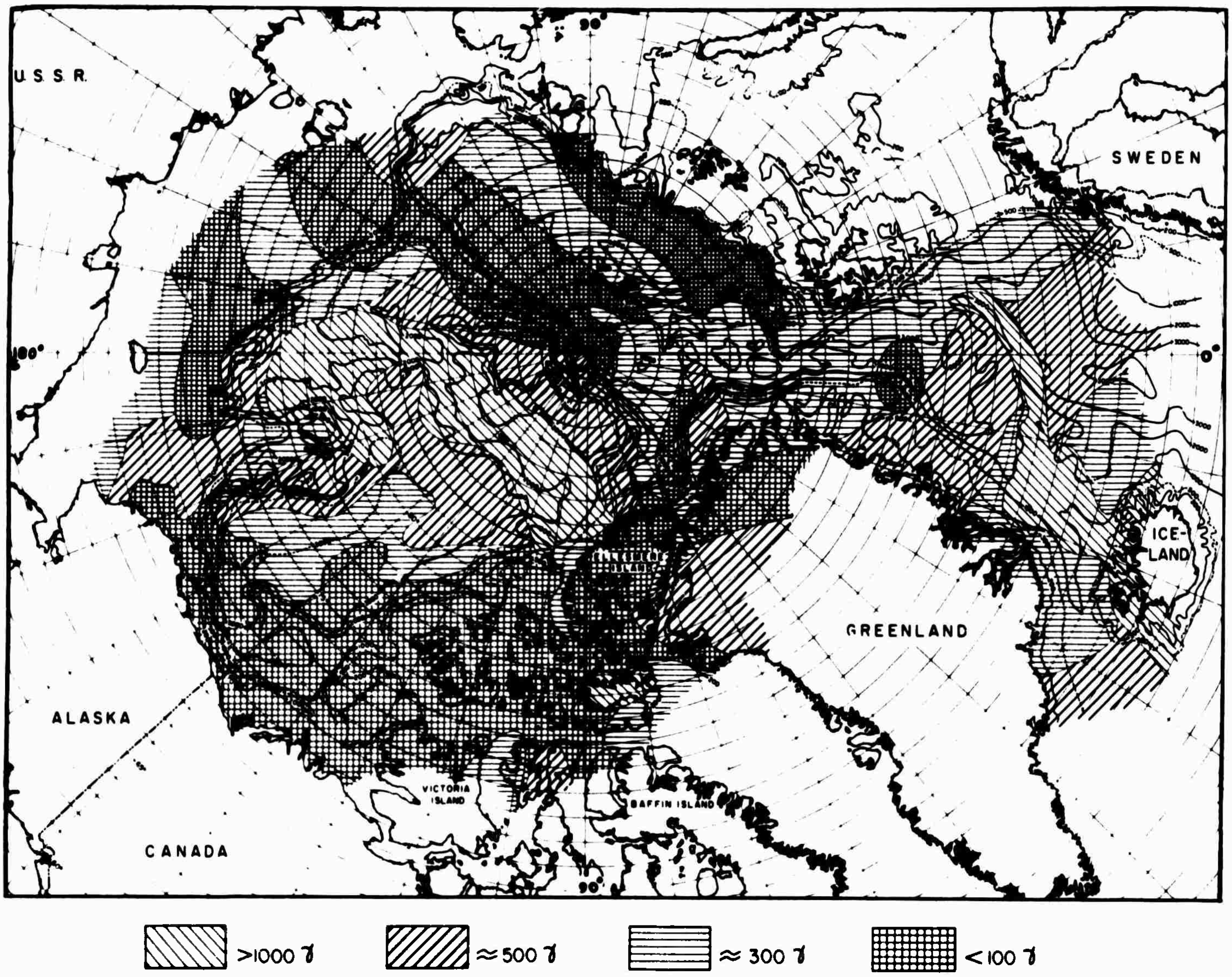

Figure I-5 MAGNETIC FIELD OF THE ARCTIC BASIN the rate of change is not usually constant. Over a period of years this change may be considerable. Thus H at Cape Town has decreased by 21% in the hundred years following the first observation in 1843. The secular drift of the northern dip pole since 1831 is shown in Figure 1-6.

The northern and southern dip poles have wandered at different rates and in different directions. The northern dip pole is now moving in a northeasterly direction at about 6 mi/yr whereas the southern dip pole is moving in a northwesterly direction at approximately 7 mi/yr. During the past century, secular variation has decreased the movement of the geomagnetic dipole at a rate of 0.05% a year and caused a westward precessional rotation of the dipole of 0.05°/yr, plus a northward shifting of the dipole by 1 mi/yr.

Lines of equal secular changes (isopors), in any component, form sets of ovals centering on points of local maximum change (isoporic foci). Considerable changes take place in the general distribution of isopors in periods as small as 10 or 20 years. The secular variation is a regional rather than a planetary phenomenon and is anomalously large and complicated over and around Antarctica.

Superimposed on the growth and decline and random movement of the isoporic foci pattern is an overall average motion of the whole pattern from east to west at an average rate of about 0.2 degree of longitude per year. The angular velocity of this westward drift appears to be independent of latitude but is closely related to the irregularities in the observed rate of rotation of the earth, after allowance has been made for the regular slowing-down due to tidal friction. Secular variations have been related to convective motion within the earth's core.

Within the past few years the aeromagnetic surveys of the Arctic Basin made by the Naval Oceanographic Office, the University of Wisconsin, and Dominion

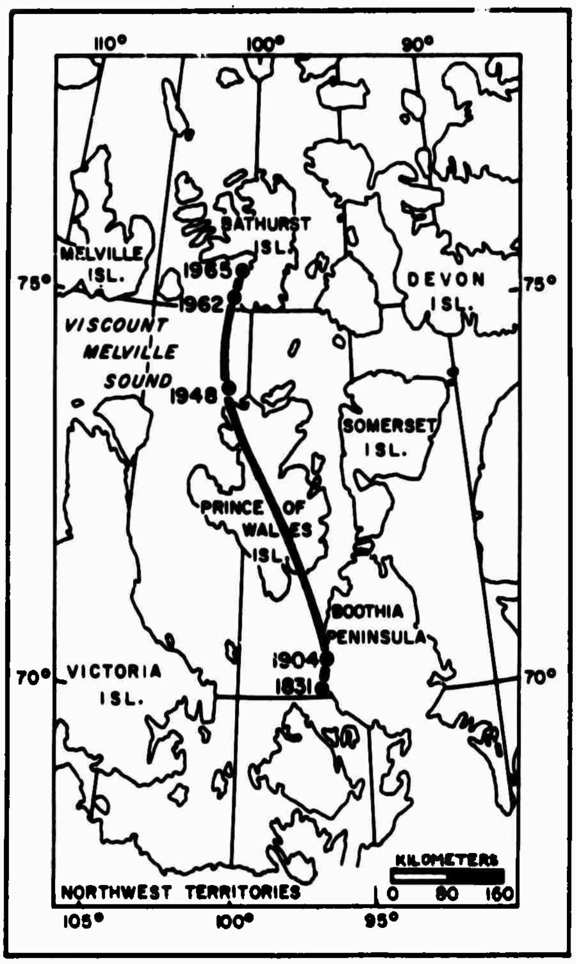

Figure I-6 MIGRATION OF THE NORTH MAGNETIC (DIP) POLE BETWEEN 1831 AND 1965

Observatory of Canada have shown the degree of disturbance of the magnetic field to be regionally variable in reflection of the complex geologic structure of the Arctic Basin (Fig. I-5). Over large areas the field is virtually undisturbed, whereas in other areas anomalies exceeding 1,000 gamma are common and gradients as steep as 2 gamma a meter are encountered.

Other salient observations are: (1) the magnetic field is much more disturbed over the Amerasia Basin than over the Eurasia Basin; (2) intense magnetic disturbance characterizes the Alpha Cordillera and suggests that it is composed of crystalline rock of high magnetic susceptibility; (3) contrastingly, the Lomonosov Ridge produces but little magnetic disturbance; and (4) the northward projection of the Mid-Atlantic Ridge into the Arctic Basin is indicated by a band of relatively intense magnetic disturbance, but its anomalies are not nearly so great as, and neither do they have the orderly arrangement of, those associated with the Mid-Atlantic Ridge proper.

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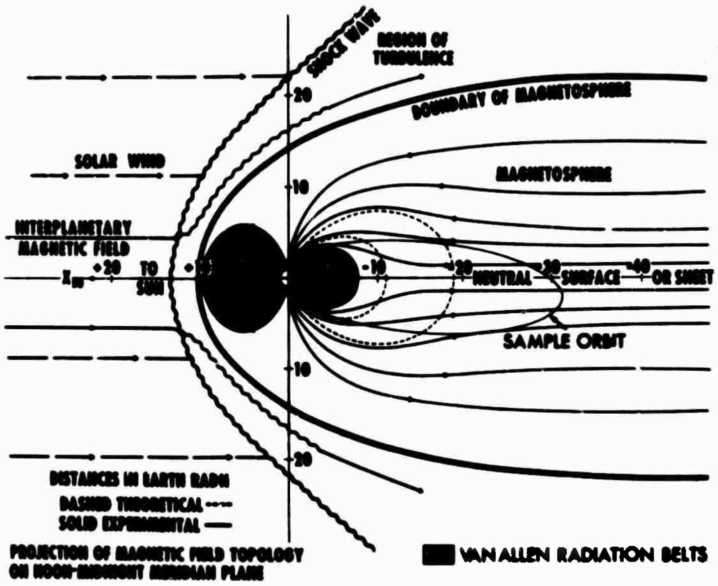
B.2 ATMOSPHERICS

B.2.1 Aeronomy

The atmosphere above the stratopause is of operational importance in the Arctic, particularly as regards of communications, radiation backgrounds, and transient magnetic variations. The time and space variations of electron densities in the ionosphere reach their widest limits over the polar (ionospheric) cap. The aurora is occasionally as bright as the full moon. Aurorae occur in both polar regions, the polar aurora of the northern hemisphere being termed aurora borealis and that of the southern hemisphere, aurora australis.

The magnetic fields in the space immediately surrounding the earth are contained within the magnetosphere. Figure II-1 illustrates our present knowledge, which is based on a large number of satellite observations. The solar wind, a stream of fully ionized hydrogen blowing continuously with varying intensity, compresses the sunward side of the magnetosphere to a minimum thickness of approximately 32,500 miles. In the direction away from the sun, however, the earth's magnetic field is drawn out far behind the earth in a magnetic tail that extends to at least 1,000 earth radii. The magnetic field lines terminating on the day and night sides over the polar caps go into the magnetic tail where they may or may not connect. However, some of the high-latitude magnetic field may be connected, at times, directly with the interplanetary magnetic field. Satellite observations have established the existence of a neutral sheet inside the magnetic tail and also a shock front toward the sunward side of the magnetosphere that can be understood by using the methods of hypersonic gas dynamics.

The "normal polar ionosphere" is unique because of two independent factors found nowhere else on the globe. The first is the presence of the dip poles



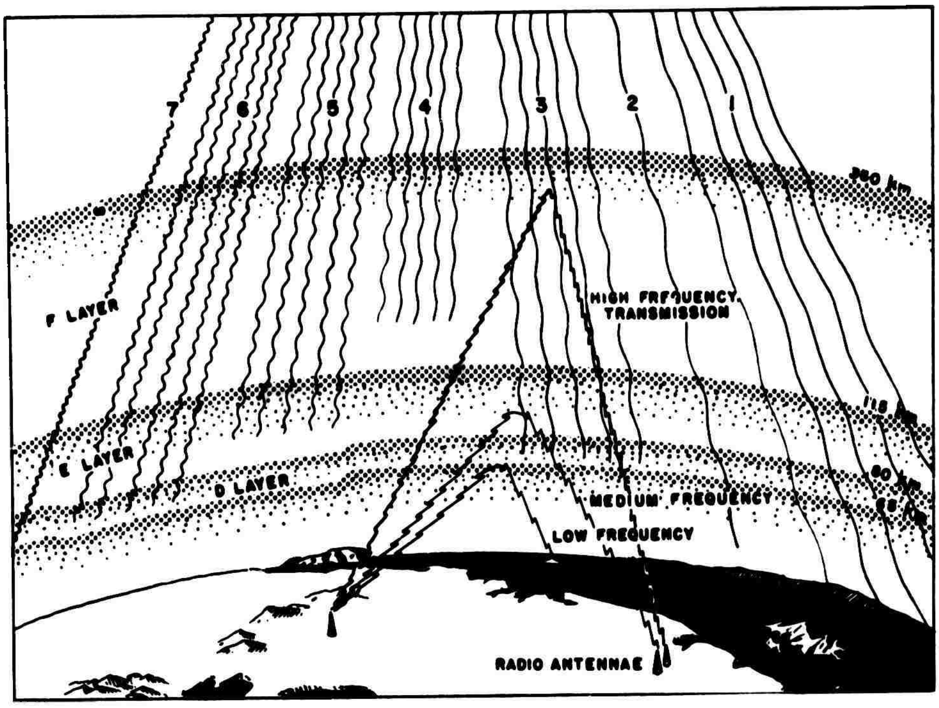
Magnetic-field topology within the magnetosphere in the noon-midnight meridian plane. The relative positions of the neutral surface or sheet in the earth's magnetic tail and the co-rotating magnetic field lines supporting trapped-particle motion are indicated. These include the classical Van Allen radiation belts. A cylindrical symmetry about the earth-sun line has been assumed for the boundary of the earth's magnetic tail in this presentation. A major problem for conjugate-point studies is determining from observations whether the field line connecting a conjugate pair is connected simply (as in the shaded region) or in a complex manner, for example with the field lines going out into the tail region.

Figure II-1 EARTH'S MAGNETIC FIELD

within the polar regions and may be considered accidental. The second factor, arising from sun-earth geometry and the tilt of the earth's rotational axis, produces the seasons and long periods when the sun is continuously absent above the polar circles. Even at ionospheric altitudes, a given layer, depending upon its altitude and latitude, may be sunlit or dark for periods of months. Thus, the polar ionic layers enjoy the constant presence or absence of solar ionizing radiations for periods of months; and they encounter a situation appreciably different from that presented to the lower-latitude ionosphere.

Figure II-2 shows the behavior of incoming solar radiation, which creates the layers of the ionosphere. Infrared radiation of relatively long wave length (1 in the Figure) is perceived as heat. Intermediate to this and the ultraviolet (2) is the visible radiation, light. The lethal ultraviolet would profoundly influence life on earth were it not absorbed at altitudes above 12 mi. Other ultraviolet radiations (3 and 4) of shorter wave length create the D and F layers (or regions) of the ionosphere by ionizing some of its constituent gasses. Of still shorter wave length are X-rays (5, 6, and 7), the longest of which (5) ionize gasses to form the E layer. Medium-length X-rays (6) appear during solar flares and cause depressions of the D and E layers that severely restrict radio telecommunications. Ultrashort X-rays (7) do not reach the earth's surface.

Under quiet conditions, the E and F_1 layers of the polar ionosphere are well-behaved and conform to the general predictions of the Chapman theory. The diurnal variation of electron density at the maximum of the layers is fairly symmetrical around local noon; the layers appear continuously during summer when the sun remains constantly above the horizon and are absent during the polar night when no solar ionizing radiation is present. Average values of noon median critical



The behavior of incoming solar radiation.

Figure II-2 THE BEHAVIOR OF INCOMING SOLAR RADIATION frequencies, determined over one sunspot cycle range at a number of North American arctic sites from 1.6 to 3.6 mHz for the E layer and 3 to 5.5 mHz for the F_1 layer.

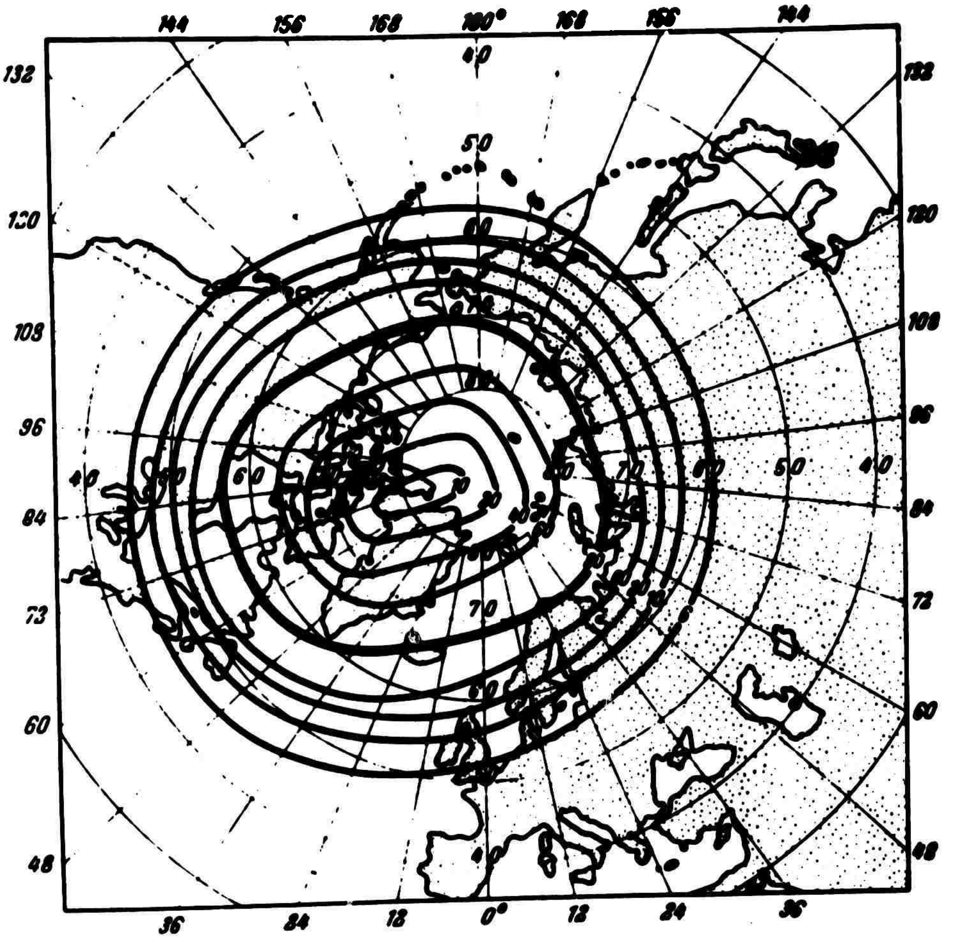
The F_2 layer is highly irregular in many respects. It does not follow the simple Chapman theory requiring a dependence of electron density on the solar zenith angle. At many places the maximum of ionization does not occur, for example, near local noon. During summer, when the sun constantly illuminates the layer, the critical frequency is almost constant near 5 to 6 mHz during sunspot maxima and 4 mHz during sunspot minima; and it displays a very small variation throughout the 24-hour period. In winter, the diurnal critical frequency increases in range markedly, attaining its greatest amplitudes of 8 to 10 mHz at the geographic pole during the maximum of solar activity. The amplitudes decrease with decreasing solar activity to about half the above-stated values.

In the absence of sunspots, the average values of the noon F_2 layer median frequency during winter portray the non-solar contribution to the layer's electron density. The same type of data gathered during the summer reveals the solar influence. The maintenance of the polar ionosphere in winter indicated the dominant role played by non-solar electron-ion production processes. These processes include diffusion that is due to density gradients and gravity; horizontal and vertical drifts caused by electric fields; movements of ionization induced by neutral air winds; and the redistribution of ionization caused by changes in ion or electron temperature. All these processes in the F region are not yet well-understood.

Because it is immediately evident to the unaided eye, the visual aurora has been observed and described for millennia. It probably results from the bombardment of the upper atmosphere by solar protons and, more importantly, by energetic electrons and protons accelerated by processes still unknown that probably occur in the tail of the magnetosphere. The bombarding particles excite or ionize the atmospheric gasses. Aurora activity occurs most frequently in zones about 23° from the geomagnetic poles. The average annual overhead occurrence of aurorae in the northern hemisphere is shown in Figure II-3. The zone of maximal auroral activity covers a latitude band called the auroral oval. This zone corresponds to the outer boundary of the region of electrons trapped in the magnetic field. During magnetic activity this zone moves southward by several degrees.

By far the most common altitude of the lower border of the aurora is 50 mi, but lower heights have been reported. The maximum altitude has been found to be over 550 mi. Visual auroral forms include homogeneous arcs and bands, arcs and bands with ray structure, pulsating arcs, diffuse and pulsating surfaces, glows, draperies, rays, coronae, and flaming coronae. Homogeneous arcs and some other forms frequently lie along, or at a small angle to, the dipole latitude.

The luminous aurora, in the usual connotation, includes the emissions resulting from interactions between precipitating electrons and protons and the atmospheric constituents. These excitations include not only the visual aurora (that detected by the naked eye) but also those other emissions detectable by spectroscopes in the near-ultraviolet and infrared regions. Some of the collisional interactions produce not only excitation but ionization as well. In many cases the resulting electron densities are sufficiently large and persistent to allow their examination by high-frequency (HF), very high-frequency (VHF), and ultrahigh-frequency (UHF) probings.



Isoaurores in the Northern Hemisphere. Note the low number of overhead auroras in the vicinity of the dipole pole.

Figure II-3 ISOAURORES IN THE NORTHERN HEMISPHERE

A radio aurora is an ionization that gives rise to certain characteristic types of radio reflection. If both the luminous and ionized (radio) aurorae are caused by the same stream of bombarding particles, they should exhibit a correspondence in time and space. This correspondence, however, is not always found; and probably this is true for at least two reasons. First, since the cross section for excitation or ionization is energy-dependent, an incoming stream may on some occasions produce excitation at one location and ionization at another, or it could produce ionization without excitation, etc. Secondly, a radio wave that is probing auroral ionization is sensitive to the geometry of the ray path involved - that from sender to aurora to receiver. Roughly speaking, a reflection from an ionized aurora will be detectable in the radar receiver if the incident ray is closely perpendicular to the ionized auroral sheath. If this aspect sensitivity in detecting the ionized aurora is neglected, the conclusion could be reached that no relationship exists between the luminous and ionized aurorae.

On reflection from the ionized aurora, radio waves usually acquire a random fading rate (100 to 400 Hz at frequencies of 50 to 150 mHz) that shifts their spectrum to either side of the carrier frequency. This aurorally induced modulation may restrict the information bandpass of the ionized aurora to about 1 kHz or even less, depending upon the fading present. Bandwidths of this size prohibit radiotelephone communications by seriously garbling the signal; nonetheless, CW, radioteletype, and similar narrow bandwidth systems may be received with little loss of intelligibility.

The ionized aurora is generally considered to lie along the earth's magnetic lines of force, which in polar regions may be inclined about 20 to 30° to the vertical. If Snell's law is observed, the reflection of radio waves with respect to the earth is then different on the poleward and equatorward sides of the aurora.

On a number of occasions, reports have been issued on the reception of radio noise from the ionized aurora. Thus, increased noise levels have been noted (at distances of 300 miles from aurorae) at frequencies of 46 and 72 mHz and also at 3,000 mHz. Noise bursts at 3,000 mHz have been found to correlate with geomagnetic disturbances.

Fluctuations in magnetic recordings may usually be associated with strong outbursts of solar activity; and they may be time-related to ionospheric storminess, polar-cap absorption, auroral occurrences, and cosmic-ray events. While these correlations also may be noted at lower latitudes, the variations occur with greatest frequency in the polar regions where the interdependence is more marked.

Small-scale variations of the magnetic field occur nearly all the time. The regularly repeated diurnal variation of magnetic activity shows a dependence on magnetic time with maxima occurring around magnetic noon and midnight within the auroral zone during winter and only around noon during summer. South of the auroral zone the magnetic noon maxima disappear. The seasonal variation has a major maximum during the vernal equinox, a minor maximum in the autumnal equinox, and a minimum during the solstices.

Large changes of the magnetic field are called magnetic storms. They are caused by solar plasma ejected from the sun during solar flares impacting on the magnetosphere. Such an impact causes the magnetosphere first to become compressed, then to expand and vibrate. A part of the plasma energy enters deep into the magnetosphere and creates a storm-time belt, the ring current. Its effect on the magnetosphere has been expressed as an inflation. Another part of the plasma energy is converted into the energy of energetic electrons, which by interacting with the polar (neutral) atmosphere causes the oval band-shaped light of the aurora around the dipole and an intense magnetic current, the polar electrojet.

Most storms begin with a very abrupt rise in H intensity, usually of about 10 to 20 gamma, which lasts only a fraction of a minute and is observed simultaneously throughout the world. This feature is called the "sudden commencement." Following a sudden commencement the field remains at an enhanced level for two to six hours, exhibiting a slightly pulsating character. This is the "enhancement phase." The field then begins to weaken while undergoing a series of large erratic pulsations. These pulses are spectacularly large and erratic both in amplitude and period. This "main phase" lasts from a few hours to a whole day, after which the pulsations diminish in amplitude and the field slowly rises back to its normal level in an exponential manner. This "recovery phase" generally takes one to two days, although complete normality may not be reached for a week or more.

Since the incident particles that cause severe disturbances in the lower ionosphere are charged, they are affected by the existence of the magnetic field, resulting in a particular precipitation pattern over the polar cap. Measurements from scientific satellites close to the earth as well as several earth's radii away, together with ground-based observations of ionospheric and geomagnetic parameters, continue to provide information about the interaction between the solar-flare emission, variations in the solar wind, and the expansion and contraction of the magnetosphere resulting in ionospheric and magnetic variations over the polar regions.

The normal daytime radio-wave absorption in the middle and high latitudes is approximately proportional to some power (0.75) of the cosine of the solar zenith angle. In high latitudes several types of abnormal absorption are encountered; their existence and prolonged prevalence has long been known to communicators employing high frequencies. It became known more recently that this absorption can also cause communication outages on high-power, VHF forward-scatter circuits. Roughly, it is usually considered that radio-wave absorption will occur during magnetically active periods for those radio circuits whose great circle paths are tangent to or intersect the auroral zone. Abnormal polar radio-wave absorption may be categorized in three groups: sudden ionospheric disturbances (SID), polar-cap absorption (PCA), and auroral absorption (AA). The first two types correlate with solar-flare events.

Sudden ionospheric disturbances are associated with visible solar flares. They are confined to daylight hours and occur not only in polar regions, but simultaneously throughout the entire sunlit hemisphere. They are caused by solar ultraviolet radiation that is absorbed in the D region (45 mi) and creates large ionization. Since the magnitude of the absorption depends on the zenith distance of the sun, it is normally not very important in the polar regions. It starts within minutes of the beginning of a solar flare and lasts only several minutes to about one hour.

Polar-cap absorption is the most severe absorption event. It starts one to several hours after the solar flare well inside the polar region around the magnetic pole. Its geographic extent increases with time until it covers the polar regions inside the auroral zones, arctic and antarctic. The absorption is much stronger over the sunlit portion than over the dark portion. Therefore, the magnitude and geographic extent of a PCA event change over the polar cap with the change of illumination during the course of a day. Solar protons with energies greater than 10 MeV can enter the polar cap to be absorbed in the D region without any worldwide magnetic disturbances, which start much later. The absorption of HF wave energy can be so strong that HF communication becomes impossible; this is called polar blackout. Such a blackout lasts for several hours, or several days in extreme cases, especially when two strong solar flares and therefore two PCA's occur within a day or two, which happens during periods of sunspot maxima. The mechanism by which solar protons enter the polar cap near the magnetic pole at the beginning of a PCA event is known, but the physical processes that lead to the geographic spreading of absorption during the following phase is not yet understood.

Those absorption events are called auroral absorption (AA) when they occur at latitudes above 50° in either hemisphere and are measurable by radio techniques. They are short-lived, i.e. lasting mostly only a few minutes and rarely up to a few hours; and are not related to PCA's and SID's. They are mostly associated with magnetospheric substorms (magnetic "bays") and aurorae. They display a very pronounced maximum in occurrence around dipole latitude 65°, with the magnitude of the absorption usually largest in winter (in both hemispheres) and their diurnal variation peaks a few hours before local magnetic noon. The average absorption in winter is twice that in summer, but from the zone of maximum occurrence their amplitude decreases toward the pole. The absorption has a well-defined peak a few degrees south of the visual auroral oval. Some events cover relatively small areas and others very wide areas. Some events occur more or less simultaneously in both hemispheres, but others do not.

Auroral absorption is only one of the many manifestations of precipitating particles (electrons and at times protons) coming from the magnetosphere into the ionosphere. For example, other effects are: visible aurorae, magnetic substorms, micropulsations, VLF hiss and chorus, sporadic E, and spread F. Recent satellite measurements have shown that precipitating electrons with energies of about 1 to 1,000 keV and protons above 100 keV are related to auroral absorption events. A discrete event is intense, is limited spatially, shows rapid and deep fluctuations, and in general can be thought of as a "splash." The diffuse event, on the other hand, can be associated with a particle influx that is steady and widespread and that can be aptly described as a "drizzle." Most of the evidence suggests that the discrete event arises from soft electron fluxes and the diffuse from relatively hard fluxes. The discrete event occurs roughly along the auroral oval, whereas the diffuse event occurs in a separate ring several degrees farther south.

B.2.2 Radio Propagation

Radio systems in the Arctic employ transfers of energy by most of the normal propagating modes. Those modes involving contact with the earth's surface or the ionosphere will, in many cases, behave quite differently in the Arctic than in the temperate zone. Those portions of the arctic land surface that are covered with snow and ice can introduce great losses of surface waves. Since the arctic ionosphere has ionization profiles that differ greatly from those in temperate regions, there are times when high-frequency ionospheric losses in the Arctic are much greater than at low latitudes. At these same disturbed times, low-frequency ionospheric losses are often less than at low latitudes. A listing of the types of propagating modes that may be important for arctic paths together with the significant earth parameters is given in Table 2-I.

Once a given communication path has been established and its distance determined, there are usually several types of propagation mode that can be employed and a range of frequencies for each mode. The choice of frequency and propagating mode will be influenced by the specific path-terrain and the data

Table 2-I PROPAGATING MODES USED IN THE ARCTIC AND GEOPHYSICAL PARAMETERS THAT MAY INFLUENCE THE PROPAGATING WAVE

	Earth's Surface			
Propagating mode	near antenna	along path	Ionosphere	Troposphere
Ground wave	x	x		
Sky wave	x		x	
Earth-ionosphere duct	x	x	x	
Ionospheric scatter			x	
Tropospheric scatter				x
Auroral scatter			x	
Meteor scatter			x	
Space wave				*
between elevated points				*
airborne relay				*
satellite relay			x	*
reflection (passive relay)				*
refraction (mountain type)		x		*

* Only at some frequencies

rates and reliability required. For example, in the VLF range, as in the LF range with a ground-wave mode, propagation is much more effective over arctic sea ice than over an ice cap or a permafrost region. In addition, line-of-sight systems have ranges that vary greatly depending upon terminal and on-path elevations. Some of the types of propagation mode and associated frequency ranges that are useful for arctic propagation paths of given lengths are summarized in Table 2-II.

Since most communication systems or sorts of equipment are designed to operate in specific frequency ranges (Table 2-III), anticipated propagation conditions are described in the following sections for specific frequency ranges.

VLF communication systems are primarily for broadcasting over a large area with path lengths up to about 6,000 miles. Because of the noise levels at these frequencies, the transmitting installations employ large antennae with transmitter powers ranging from 100 kW to 2 MW. These systems carry only 10 to 100 words a minute. Because of the cost of such installations, VLF is not used for point-to-point communications in the Arctic; but since the fields exist throughout the Arctic, VHF communication does provide a useful broadcast service as well as a means of navigation.

Arctic ionospheric conditions modify VLF propagation characteristics slightly. For most applications other than precise navigation, this effect is rather negligible. The primary modifier is the conductivity of the lower portion of the earth-ionosphere wave guide. Around 15 kHz, the normal attenuation rate over temperate-zone sea water ranges from 1 to 2-1/2 db/600 mi. When the path is over normal land, an attenuation of about 1 to 2 db/600 mi must be added to the sea water rate, and when the path is over a region of deep permafrost or an

Table 2-II TYPES OF PROPAGATION AND FREQUENCY RANGES USEFUL IN THE ARCTIC FOR VARIOUS PATH LENGTHS

Path Lengths (miles)	Propagation Modes	Frequency Ranges
1 to 5	ground wave (surface)	MF, HF
	direct (line of site)	VHF, UHF
5 to 50	ground wave (surface)	MF, HF
	point to point and refraction	VHF, UHF
50 to 500	ground wave (surface)	LF
	sky wave	LF, HF
	tropo-scatter	UHF
	iono-scatter	VHF
	line of site (with relay)	VHF, UHF SHF
500 to 5,000	ducted wave	VLF, LF
	sky wave	LF, HF
	auroral scatter	HF, VHF
	meteor scatter	VHF
	line of site (satellite relay)	UHF, SHF

Table 2-III FREQUENCIES AND THEIR USES

Longline: (Open wire line or multiple pair cable)	Very short inter-building and inter-site (less than 10 miles)
Very Low Frequency: (3 to 30 kHz)	Standard frequency measurements WWVL 20 kHz, Boulder
Low Frequency: (30 to 300 kHz)	Radio beacons; radionavigation WWVB 60 kHz, Boulder
Medium Frequency: (300 to 3,000 kHz)	Radionavigation; station-to-station mobile
High Frequency: (3 to 30 mHz)	Long-distance skip; station-to-station mobile; distress, emergency
Very High Frequency: (30 to 300 mHz)	Ionospheric scatter; station-to-station mobile; radionavigation
Ultra High Frequency: (300 to 2,000 mHz)	Tropospheric scatter; station-to-station mobile; microwave
Super High Frequency: (3 to 30 gHz)	Milimeter wave

ice cap, an attenuation of about 6 to 20 db/600 mi must be added. The attenuation increase is usually much greater during the day than at night. This is particularly true at the higher frequencies (near 30 kHz) where during the night most of the energy is contained high in the earth-ionosphere guide ; and as a result, a loss-prone surface at the lower boundary does not appreciably affect the attenuation rate.

LF systems are used for both broadcast and point-to-point communications, and also for navigation in the Arctic. The very low conductivity of some of the land considerably modifies the design procedures for antenna installations as well as the overall system design. Although in many cases the low-conductivity regions present additional problems, very efficient transmitting antenna systems are feasible in spite of the low conductivity at the surface. In addition, there are some advantages to be gained by using horizontal wire antennae of the Beverage type for transmitting and receiving. The horizontal and vertical positioning available with these antennae makes the installation relatively economical in many cases. Low-frequency systems also have a low data rate of 100 to 500 words a minute. The transmitting antenna installations can be large (330 to 1,300 ft high). Fortunately, the receiving antennae can be small (less than 3 ft high), although in some cases Beverage receiving antennae over a mile long are employed.

At LF, the propagation mode involved is primarily the ground wave or a combination of ground and sky waves. Typical ground-wave curves show that the fields produced for ice-cap paths are very highly attenuated compared with fields over sea water. In addition, when sky-wave propagation is involved, the conductivity for a few wave lengths in the foreground of the antenna can become very important. This factor must be very carefully considered in the design of any

LF system in the Arctic. At present there are installations where a relocation of several miles would increase the effective radiated power for the sky-wave paths involved by factors of 10 to 20 db. Typical transmitter powers range from 10 to 50 kW with useful path lengths ranging from 150 to 1,500 mi. Ionospheric disturbances that frequently interrupt normal HF communications in the Arctic have very little effect upon LF and, in fact, in the lower portion of this band the field strength may actually be enhanced during such disturbed periods.

The MF portion of the frequency spectrum is of limited usefulness in the Arctic because of the rapid attenuation of the ground-wave mode as well as the high attenuations on reflections from the ionosphere. The upper portion of the MF band is useful during undisturbed nighttime conditions and can also be employed for communications along paths that are primarily over sea water. The MF band has the advantage over VLF and LF bands in that the transmitting antennae are much smaller and the atmospheric noise levels are lower. The ranges available, however, are also considerably smaller; but for some short paths, usually 60 mi or less, MF systems may be the most economical solution. There are some short-range mobile systems operating in this band. Other uses include navigation systems such as A-N beacons and LORAN.

At high frequencies, it is possible to obtain good communications over long distance (up to several thousand miles under good conditions) with much lower transmitter powers than at the lower frequencies. The antennae required are relatively small (about 6 to 65 ft long), and very simple configurations are useful. For sea-water paths, the ground-wave propagation mode provides useful communication for up to several hundred miles. Over poor arctic soil or ice caps, the useful range is much less - in some cases less than 6 mi. The sky-wave mode, useful in the region from 60 to 1,000 mi, makes use of ionospheric reflections from the

E and F layers. In this frequency band, an appreciable amount of sky-wave energy is absorbed as the rays traverse the D layer. Since this absorbing layer has to be traversed twice for each sky-wave hop and the absorption for each traverse becomes quite high in polar regions, HF communication links are highly susceptible to blackouts due to auroral and other corpuscular-associated ionospheric disturbances. Since this type of attenuation decreases with increasing frequency, it is usually desirable to use as high a frequency as will effectively be reflected from the E or F layer under disturbed conditions.

The sky-wave mode also suffers two additional disadvantages. First, multipath scattering, particularly that occurring during aurorae, can result in severe fading with rates roughly proportional to frequency. The rather large time delays can be so severe as to make voice signals unintelligible. Second, depending upon ionospheric conditions and the length of the path, the radio wave may penetrate the E and F layers completely instead of being reflected. This tends to limit operations to that part of the HF band below about 15 mHz. The upper limit of 15 mHz can be exceeded a large percentage of the time over transmission paths to the south. Occasionally, particularly during the summer months, sporadic-E propagation permits the use of higher-than-normal frequencies. This is often the case during ionospheric disturbances. The maximum usable frequency will depend, of course, on the time of day, the season, the terminal locations, and the phase of the sunspot cycle. Contrary to conditions at VLF, where the atmospheric noise levels are rather low, the antenna noise levels in the Arctic can be high, primarily because of cosmic radio noise that has penetrated the ionosphere, although in some cases atmospherics and manmade transmissions from distant stations may be the limiting factor.

The rather simple installations required, the low radiated powers and appreciable bandwidth (typically several Hz), and the wide range of path lengths that can be employed make HF systems a fairly useful means of communication in the Arctic. The disadvantages include frequent communication failures that are due to excessive absorption, multi-path, or maximum usable frequency failure. The possible means of reducing these failures include: (1) decreasing the receiver bandwidth at the expense of the data rate, possibly causing frequency stability problems; (2) using vertical polarization, which would be significant, however, only for frequencies at or near the gyro-frequency, i.e. between 1 and 3 mHz; (3) using care in the selection of frequencies; (4) careful siting of antennae to permit high-conductivity paths where possible for short ranges, or launching over a short section of high-conductivity soil; (5) using receivers with very low noise figures; and (6) SSB modulation for voice and data systems.

The application of real-time propagation information operationally to communications circuits by oblique-incidence ionospheric sounding of the transmission path removes the unknown variable from high-frequency propagation. In general, where a choice of operating frequencies is available, it is preferable to use the highest frequency during blackout or near-blackout conditions. This is due to the (approximately) inverse frequency-squared dependence of the ionospheric absorption, which causes the lower portion of the HF band to be more highly absorbed than the higher portion. Maximum-usable frequencies tend to be increased in the Arctic during disturbances, and for this reason it may be beneficial to try a higher frequency at times when a lower frequency has failed during disturbed conditions.

For radio communications a technique of forecasting disturbed ionospheric conditions by electronic computer has been developed and tested and should become operational within the next few years. However, the relationship of sun-spot activity to magnetic storms and the usual orderly development of the storms makes it possible to forecast the quality of radio propagation with reasonable accuracy. Short-term radio-propagation forecasts are broadcast on each of the standard radio carrier frequencies of WWV (at 19.5 and 49.5 minutes past each hour) and WWVH (at 9.4 and 39.4 minutes past each hour). Those forecasts (at 1700 and 2300 GMT for WWV; 0600 and 1800 GMT for WWVH) predict communication conditions expected for the succeeding six or more hours. The WWV announcement is for the North Atlantic region and the WWVH announcement is for the North Pacific. In the near future it may be possible to establish a mathematical model of the arctic ionosphere which will permit still better forecasts.

A type of propagating mode called ionospheric scatter makes use of scattering by irregularities in the ionized D region. The effective forward-scattering coefficient at VHF is quite low and such systems require large directional antennae and powerful transmitters. For this reason, ionospheric forward-scattering is not an effective communication system for small mobile groups or ships. Its advantages include long path lengths - roughly up to about 1,000 miles - and a number of data links. Because both absorption and scatter are critically dependent upon frequency, the optimum frequency range in the Arctic is fairly narrow and lies somewhere between 50 and 70 mHz. Because of the rather high cost and inflexibility of such systems, their use is quite limited. VHF propagation may also involve the scattering of radio waves from meteor trails. In this technique, transmitters operate continuously on closely spaced frequencies at each end of the circuit until a trail of meteor ionization occurs with a position and orientation that permit a predetermined energy level to be exceeded at the receiving terminal. The receipt of such a signal is used as an indication that good propagation is available, and the message is immediately transmitted at high speed until the adjoining receiver shows that the propagation mode has faded below some preset level. This intermittent communication system has the advantage of lower radiated power requirements than the standard ionospheric forward-scatter system. Also, because of the character of the propagation, it is relatively private. Path lengths of up to 1,200 mi may be used. It has been found that, like the VHF ionospheric forward-scatter mode, meteor-burst communication links are much less affected by high-latitude ionospheric disturbances than are HF communication links.

Both types of VHF ionospheric scatter can be disturbed by the major polar cap events. Such events have occurred several times each year during sunspot maxima and are likely to cause severe attenuation or even complete failure of the link during daylight hours for periods of up to two to four days. The obvious defect of meteor-burst communication systems is the intermittent nature of the propagation path. The delay time is dependent upon the sensitivity of the system which is a function of the operating frequency, the radiated power, the bandwidth, and the sensitivity of the receiving system. Under some circumstances, when instantaneous communication is not required, meteor-burst communication is acceptable.

Another form of VHF communications employs the direct or space wave with some intermediated high-elevation relay - an airborne relay, passive reflectors, or diffraction over a mountain or mountain range. The last mode is, of course, subject to multi-path and to some tropospheric fading, and its bandwidth capacity is limited to a few hertz. It is relatively little-used except under unique circumstances, such as when a high mountain is surrounded by a large plain. The airborne relay is useful for distances of up to 300 mi. A suitably located aircraft can be used for short-term relay work to provide communications along a very broad band with rather low ambient noise levels. When high-gain antennae are used on the ground, the transmitter power requirements are fairly low.

One additional method of arctic VHF propagation is scatter from auroral ionization. This form of scatter decreases rapidly in intensity with increasing frequency; even so, the high frequency of 3,000 mHz has been observed. As a communication technique it has numerous disadvantages, primarily those of a rapid fade rate and a very intermittent and unpredictable occurrence of the propagation mode. Under certain conditions, in the lower portion of the VHF band (30 to 60 mHz), quite strong auroral-scatter signals can be expected during ionospheric disturbances and then such a propagation mode can be of particular value as a HF circuit back-up. Any such link must take into account the pronounced directional sensitivity and limited height of the auroral ionization. Such ionization tends to act rather like a metallic mirror inclined parallel to the magnetic field lines and is most common at heights of about 60 miles. In general, east-west propagation in the Arctic would be via auroral forms to the north of the midpoint. North-south auroral-scatter propagation would involve scatter from auroral ionization several hundred miles to the north of the northern station. In other words, for north-south auroral-scatter modes, both antennae should be directed northward at low angles of elevation toward a common scattering column in which the line of sight is approximately perpendicular to the earth's magnetic field lines.

UHF frequencies are useful for short-range line-of-sight systems where the antennae are small and the powers low enough that mobile or hand-carried equipment can be employed. For conditions where the line of sight does not obtain between the transmitter and receiver terminals, it is possible to use some type of relay mode, such as passive reflectors, mountain diffraction, airborne relays, and satellite relays. Most of these modes, with the possible exception of mountain diffraction, have very wide bandwidth capabilities. At the greater ranges directional antennae are usually employed, but for moderate gains, their size fortunately is not great.

An additional useful type of propagation mode in this frequency range is tropospheric scatter, which has found wide application in the Arctic. When properly engineered, such systems are very reliable, are not subject to failure because of ionospheric disturbances, and may have an unusually wide bandwidth capacity (up to a few megahertz) depending upon the system parameters. The major disadvantage is the high cost of the large antennae and powerful transmitters needed to overcome the rather high losses of the tropo-scattering mode.

Although it is possible to use the SHF region for short-range direct and line-of-sight communications, it is usually more economical to use VHF or UHF systems for such applications. The primary present-day application for SHF is in satellite relays although some of the satellite systems are also active in the UHF region. In the future, however, use of SHF may expand vastly, as it offers the broadest bandwidth yet to accommodate the proliferating need.

B.2.3 Bibliography

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B.3 SEA ICE

B.3.1 Physical Properties

Sea ice is the dominant element in surface operations in the Arctic Basin. Its presence or absence determines the areas where conventional shipping may operate, and the nature of its upper surface limits the types of airborne vehicles which may land on it. The properties of sea ice are derived initially from the presence of dissolved salts in seawaters. The composition of these salts is sufficiently uniform throughout the world's oceans that it can be described by a single parameter known as the salinity (S). Salinity is the weight of the salts contained in 1 kilogram of seawater expressed as grams per kilogram (°/oo) and varies from 36°/oo or more in tropical waters to 31°/oo or less in arctic waters.

Pure water reaches its maximum density at 39.2°F and expands on further cooling. The effect of this inversion in density on the formation of an ice cover over fresh water is well known. For seawater, both the freezing temperature and the inversion temperature decrease linearly with increasing salinity, until at a salinity of 24.7°/oo each of these temperatures is equal to 29.7°F. This critical salinity may conveniently be taken as the division between brackish water and true seawater. Seawater in the Arctic usually has a salinity at the surface of between 28 and 32°/oo, and it freezes at about 29°F. This is well below the critical salinity and temperature so that the density of arctic seawater increases continually with decreasing temperature, down to the onset of freezing. As a result, a seawater column of uniform salinity, cooled from the top, will develop a vertical circulation so that the entire column must cool to the freezing point before freezing begins. The column of uniform salinity may extend to the bottom of the sea or may terminate at a thermocline or halocline where a denser layer underlies the surface water. As an example, in the fall following the summer melt period, the surface water is of low salinity and will begin to freeze without turnover to any appreciable depth. As the freezing continues, the surface waters are enriched in salt, and turnover penetrates to greater depths, reaching 160 feet by the end of winter over most of the Arctic Basin. Immediately after open water begins to freeze, there is usually a rapid buildup of the first foot or so of ice. This occurs because ice reflects the incoming radiation better than water but, at the same time, continues to emit radiant energy with about the same efficiency as water so that the net heat loss is initially rapid. Both the temperature of the ice surface and that of the overlying air drop rapidly. As the ice cover thickens, its own low thermal conductivity, together with any snow cover that may accrue, slows the rate of heat loss from the sea and limits the thickness reached.

The crystal structure of ice is intolerant of impurities such as the inorganic salts which make up most of the dissolved material in seawater. If seawater is frozen very slowly, the result is a layer of pure ice from which the impurities have been rejected into the seawater below it. Natural freezing is never this slow, however, and sea ice always contains a certain amount of entrapped salts. The actual amount is dependent upon the freezing rate, but new sea ice generally has a salinity of 4 to 6°/oo. Thus freezing of seawater in nature is a refining process with about 80 to 90% of the salts being excluded from the ice cover.

The presence of the salt in sea ice is of such importance that the physical properties of sea ice are almost entirely determined by the salt content. The peculiar structure which results dominates the physical parameters to the extent that one can relate any of them to brine volume with little regard to the properties of pure ice, except in the limiting case where the brine volume is equal to zero. Brine volume is defined as that portion of the volume of sea ice occupied by fluid, either liquid brine or air bubbles.

It is known that the atoms in ice are arranged with hexagonal symmetry. This atomic-scale symmetry appears microscopically in the hexagonal pattern of most snowflakes. The unit cell is in the form of a prism whose cross section is a regular hexagon which has one main axis of symmetry along the length of the prism. This is known as the *c*-axis and any plane perpendicular to this axis is called a basal plane. Frazil ice is a single crystal with its *c*-axis accurately perpendicular to its areal extent. From this we deduce that ice growth is much more rapid along a basal plane than along the *c*-axis.

When seawater is cooled to its freezing point and more heat is then removed, the initial ice forms in very thin disks or platelets known as frazil ice. On the average, these platelets are about 1 in wide and about 0.02 in thick. Their shapes vary considerably, ranging from hexagonal dendrites through irregular shapes to almost square. Frazil ice generation is not confined to the surface but may occur anywhere within the top few inches of water. The frazil crystals are pure ice so they float to the surface and form a slush which gives the water a slightly oily appearance. With continued cooling, the crystals consolidate into a solid but extraordinarily flexible surface ice cover about one inch thick. The bow wave of a ship may cause transverse waves in this cover which will not fracture the ice until an amplitude of several inches is exceeded.

When frazil ice first forms, if there is no wind or current, each crystal will float to the surface and lie with its *c*-axis vertical because of buoyancy. Such conditions are rare in nature, and normally the frazil crystals are jammed together by wind and waves so that the first crust is consolidated from masses of frazil crystals with a great variety of orientations. As growth continues, the frazil particles act as seed crystals and some of them grow. The growth potential of an ice crystal depends critically on the angle between its *c*-axis and the vertical. The larger this angle, the more rapid the growth of the crystal. This geometrical selection results in the bulk of a sea ice cover consisting of crystals with nearly horizontal *c*-axes.

A single crystal of sea ice is defined as one which shows a uniform brightness in polarized light. A crystal has a pronounced internal structure and consists of a large number of thin parallel platelets of pure ice separated by rows of cells of liquid brine. Typical platelet thickness and brine cell diameter are of the order of 0.02 and 0.002 in, respectively. Each platelet may be an inch or more in length. The platelets grow downward into the water and thicken near the ice cover. When the brine film separating the platelets becomes thin enough, surface tension causes it to break into individual vertical columns interrupted by ice bridges linking the platelets. The sea ice thus solidifies into a compact mass threaded through by very large numbers of brine cells. Almost all the brine entrapped in sea ice is situated in these brine cells within crystals.

The average salinity of the oceans is about 35°/oo which is taken as the "standard" salinity of seawater. The three principal dissolved salts in the order of their concentration are $NaCl$ (23.48°/oo), $MgCl_2$ (4.98°/oo) and Na_2SO_4 (3.92°/oo). The phase diagram of seawater below its normal freezing point is quite accurately known down to a temperature of -13 to -22°F. At any given temperature there is a unique salinity at which ice and brine can remain in equilibrium with each other. This is the freezing point for this brine. Thus, ice at 28.6°F can remain in equilibrium with standard seawater, the amount of ice neither increasing nor decreasing with time. When seawater trapped in ice is cooled, in the range from its freezing point down to 17.3°F, all of the salt is in solution in the brine. At 17.3°F sodium sulfate starts to precipitate as a hydrated solid, $Na_2SO_4.10\ H_2O$. On further cooling, the amount of precipitated sodium sulfate increases, but no other salt is precipitated until a temperature of -9.2°F is reached. This temperature marks the onset of precipitation of solid $NaCl.2H_2O$ and the apparent initiation of ice strength reinforcement by this solid salt. Below -9.4°F the characteristics of sea ice change notably - the ice is whitish because of the sodium chloride crystals, more brittle but harder to drill or chisel, and significantly stronger.

Ice is a visco-elastic material that flows readily under sustained stress, so that its strength varies with the rate of stress application. The following discussion of the various ultimate strengths is valid only for a rapid application of stress. Experimentally it is found that ice strengths appear to be independent of stress rate for loads applied at rates greater than approximately 3 $lb/in^2/sec$. The table shows the average results obtained at 23°F for such loads on freshwater ice samples with a minimum dimension of 2 inches (test samples smaller than this tend to give higher values). The word average must be emphasized. Individual test results may easily differ from these averages by a factor of two or even three. Although interest here is with sea ice, the table can serve as a basis for discussing the corresponding strength in sea ice, which cannot be tabulated in this way because of the dependence on brine volume. The most extensively studied property is the ultimate tensile strength, and the commonest measurement is the ring-tensile test in which a hollow cylinder of ice is broken by a load applied along the full length of the cylinder in a direction perpendicular to its axis. As a result of this load the cylinder fails in tension. The maximum tensile stress in this test is applied to a very small volume of ice, and the results observed are appreciably higher than those obtained by the conventional "simple" tensile tests. However, because of the convenience of the test it lends itself to mass production, and a very large number of tests have been made by a considerable number of observers. In general the values are in good agreement with the comparable value for sea ice.

Ultimate Strengths of Freshwater Ice
(lb/in^2)

Unconfined Compression	Shear	Tension		Flexure	
		Simple	Ring	Small beam	*In situ* beams and cantilevers
210	90	102	174	102	42

Two other principal experimental methods have been used. In the first, small ice beams (typically 20x3x3 inches) are sawn from the ice and, with their ends freely supported, are broken in flexure. The figure of 102 lb/in^2 for the flexural strength of pure ice comes from these tests. A number of large-scale, in-place cantilever beam tests have also been made. For such a test three vertical cuts are made through the ice cover, thereby leaving a rectangular beam of ice, still attached at its base to the cover but otherwise floating freely. The beam is then broken by pushing down or pulling up on the free end. Such a test, at first sight, seems ideal for determining the actual flexural strength and hence bearing capacity of an ice cover because there is a minimum disturbance of the temperature and salinity profiles of the ice and the effect of buoyancy is automatically included in the depression tests. There are difficulties, however. In theory, bending a cantilever beam produces stress concentrations in the fixed end. Fortunately, comparisons of *in situ* cantilevers with *in situ* simple beams suggest that the actual stress concentration is about 10% or less. Also the shape of the ice beam, particularly in thick ice, is far from the theoretical ideal (of width much greater than thickness), so that uncertainties enter into the calculations. The results of such tests always show much lower strength values than ring-tensile or small-beam tests, typically about 12 to 30 lb/in^2.

Unconfined compression and shear tests on sea ice are much less numerous than tensile or flexural tests. In general, the strength decreases with increasing salinity or temperature of the ice, although data are sparse. In the light of present knowledge the best estimate is that any ultimate strength of sea ice should be calculated from the equation

$$\sigma = \sigma_o (1 - 1.9 \nu)$$

using the appropriate value from the Table for σ_o. When calculating the bearing capacity of an ice cover, the σ_o determined from *in situ* beam and cantilever tests should be used. A safety factor should also be added by the person making the calculation.

The response of ice to a periodic force with a frequency greater than one cycle per second is elastic, provided, of course, that crushing does not occur. Sound transmissions, responses to moving vehicles, and impacts resulting from aircraft landings all fall within this category. Deflections caused by parked vehicles or aircraft do *not*. Almost the only reproducible values for Young's modulus E are found from seismic and acoustic measurements. On the basis of small-sample tests the best approximation is

$$E = (1.02 - 3.56\nu) \times 1.32 \times 10^6 \text{ lb/in}^2$$

where ν is the fractional brine volume. Less is known about Poisson's ratio, but it appears to be essentially constant for sea ice over a wide range of values of ν, with a value $\mu \sim 0.30$.

Ice subjected to a steady or slowly varying load flows plastically. This plastic flow has been studied extensively in connection with glaciers, but few data are available for sea ice. The most practical problem - that of parking or storing heavy loads on an ice cover - can as yet be handled only empirically. As an example, if the ice surface temperature is 14°F a thickness of 3 ft of sea ice is considered safe for parking of a DC-3 aircraft of about 25,000 lb weight. At this thickness the ice cover does flow slightly, i.e. it sags, but over such a large surface area that the increased buoyant force of the sea water supports the load. Caution must be exercised nevertheless.

Seismic methods can be used to measure the average thickness of an ice cover if E is known. A detonation at or above the ice surface generates several types of wave including an air-coupled flexural wave whose frequency is inversely proportional to the ice thickness. This wave can be used to measure the ice thickness from the air. A second method of making this measurement from the air, by electromagnetic waves, is under study by several groups. Considerable success has been achieved in measuring the thicknesses of glaciers and ice caps, but so far it has not been possible to obtain a satisfactory reflection from the interface of the sea ice and seawater.

Since sea ice in its natural state always contains cells of liquid brine, any change in the temperature of the ice will result in a phase change in a portion of the ice, either from water to ice or ice to water, depending on the sign of the temperature variation. For this reason the concepts of latent and specific heats are closely interrelated, and in reality a definite latent heat of fusion cannot be established since the phase change from solid to liquid is a continuous process. The quantity of heat required to raise the temperature of a given amount of sea ice is always greater than that needed to raise the temperature by the same amount of an equal mass of pure ice. While the difference is small for ice of low salinity at temperatures below 14°F, the specific heat of sea ice rises rapidly as the melting point is approached, especially as the salinity increases. Because the thermal properties of sea ice are controlled primarily by its salinity and because the salt content depends principally on the rate of freezing, the phase change from sea ice to seawater is not thermodynamically reversible. Consequently, if a sample of sea ice is melted and then cooled to its original temperature, there is very little chance that it will retain its original saline and thermal properties.

Pure ice has an essentially constant specific heat of 0.50 in the range 32 to 14°F, but the specific heat of sea ice rises very rapidly with increasing temperature, becoming infinite at the melting point. Consequently it is necessary to integrate the specific heat from any desired temperature up to the melting point to obtain the quantity of heat needed to melt sea ice. Another useful quantity is the latent heat of freezing, L_s. This is the quantity of heat which must be removed from unit mass of seawater of salinity S to produce sea ice of salinity σ. The empirical equation, which can also be justified theoretically, is

$$L_s = (1 - \frac{\sigma}{S})\ L_i$$

where L_i is the latent heat of pure ice (about 80 cal/gm). The other important thermal property of sea ice is its thermal conductivity, k. Because of the brine, k is always smaller for sea ice than for freshwater ice but it approaches the value for the latter at a temperature of +5 to -4°F. Either form of ice is a poor heat conductor, as can be seen in the relative thinness of ice covers on water.

Nowhere does the dominant role of the brine in sea ice show more strikingly than in the peculiar electrical behavior of this material. The first systematic observations of dielectric properties and electrical conductivity have appeared in the last few years. For sea ice of salinity 10°/oo at a temperature of -7.6°F, the dielectric coefficient is extremely large (of the order 10^6) at 20 Hz. It decreases approximately as the reciprocal of the frequency in the audio range and more slowly above 100 kHz, reaching a value of about 10 or less at 50 mHz. The effective electrical conductivity over the same range increases slowly with frequency, from about 8 micro-ohms/in. at 20 Hz to about 80 at 50 mHz.

Sea ice transmits acoustic energy readily, particularly at low audio frequencies. Furthermore, the skeleton-layer structure below an ice sheet helps match the acoustic impedances of sea ice and seawater so that little reflection occurs at the sea ice-water interface. This has been used to permit bathymetric observations through an ice cover. If a flat transducer of a conventional echo sounder is bonded to the surface of sea ice with a thin film of a non-freezing liquid, acoustic power can be transmitted through the ice with little attenuation and with the only significant reflection occurring at the bottom of the water. In one test, the frequency used was 37 kHz. Actual data on the acoustic absorption in sea ice are sparse. Other studies suggest that the attenuation rises from a negligible value at 20 kHz to about 5 or 6 db/m at 100 kHz. An abrupt increase in attenuation occurs between 100 and 200 kHz. This is to be expected since the longitudinal wave velocity in sea ice is about 2 mi/sec, corresponding to a wave length of 2 cm at 175 kHz. This is just about the mean minimum dimension of sea ice crystals, and the attenuation would be expected to be much larger for wave lengths equal to or less than the size of the obstructions. Above 200 kHz a typical attenuation coefficient is of the order of 50 to 60 db/m.

The complete annual cycle of a cover of sea ice, restricted here to a discussion of a shorefast ice cover which remains in place throughout its history, is of value. The maximum thickness of ice in a given locality is quite constant from season to season. The first attempt to relate ice thickness to the temperature regimen of a particular area was made in 1891. The analysis showed that

$$h = \left[\frac{2k}{L_s p} \cdot E_t \right]^{1/2}$$

where h is the thickness of the ice cover with density p; L_s the latent heat of freezing of salt water of salinity S; and E_t the freezing exposure in the accumulated degree-days of ice surface temperature below the freezing point of seawater. This analysis has several weaknesses in that it ignores the specific heat of the ice, which cools after it is formed and thereby releases a considerable amount of heat, and it is based on the generally unknown surface temperature of the ice. In practice it is generally necessary to use the air temperature, which may be considerably lower than the ice temperature because of the insulating effect of snow cover and other parameters. Nevertheless, the equation works well. If E_t is calculated using air temperatures, h is usually found to vary quite closely with $\sqrt{E_t}$, and most forecasts of ice thickness are based on climatological records and a modification of the equation. The simplest modification is to substitute an empirical coefficient for the theoretical coefficient E_t. The appropriate coefficient depends on the area of the world in question and reflects the prevailing local conditions such as snow cover. A number of suggested coefficients have been computed.

A more valid but more difficult approach to the problem of ice growth and decay is to study the micrometeorology of the sea ice-air boundary and to attempt to measure the quantities of heat transferred across this boundary. Many such studies have been made; and one interesting conclusion is that in the Arctic at least, radiation is the almost completely dominant factor in determining the ice surface temperature. The ice cover will continue to grow until some time in spring when the increasing value of incoming solar radiation changes the heat budget of the ice cover to positive - that is, it starts to gain heat from above rather than lose it. This usually happens before the air temperature rises above the melting point of the snow cover. Pure white snow will reflect as much as 90% of the radiation incident on it, and the quantity of heat absorbed by the snow and ice cover is at first quite small but does have the effect of modifying the crystal structure of the snow cover and of starting to raise the temperature of the ice. After a few weeks the air temperature reaches the melting point and the snow surface suffers an abrupt drop in albedo, to about 40%. The result is particularly dramatic in the high Arctic where at this season the sun is above the horizon 24 hours a day and the sky is normally cloudless in the spring. As a consequence of the large amount of incident radiation and the high percentage of absorption, the snow cover melts rapidly (usually in one day or less) producing a "flash" flood. To be caught on the ice at this time is unpleasant, although the ice cover itself is still intact and strong.

The next stage is the deterioration of the ice cover. The heat absorbed gradually raises the temperature of the ice cover to the melting point, and flaws and cracks develop through which much of the surface water drains away, leaving dry hummocks of white ice separated by ponds and streams of water. When the ice has this appearance, operations on it must be carried out with caution. The bearing strength is uncertain and a strong wind may cause it to break into floes of various sizes.

In sufficiently low latitudes, the floes finally melt and the cycle is completed. Farther north the summer is too brief for complete melting to take place, and by late August or September the ice floes start to grow again. A floe formed in one year which has survived the following summer differs chemically and physically from ice that is less than one year old. Over the course of a year, most of the salt drains out of sea ice so that the typical salinity of perennial or polar ice (more than one year old) is about 0.5 to 1°/oo. Thus melted polar ice is quite potable. The crystal structure of polar ice (particularly when several years old) is less regular than that of annual ice. Polar ice crystals are smaller and somewhat more rounded, and their *c*-axes are no longer uniformly horizontal. The most important result is that polar ice is extraordinarily resilliant. Standard icebreakers would experience extreme difficulty in attempting to "break" polar floes, even in summer. It can be done, but there is an excellent chance of breaking the plating on the icebreaker as well.

In summer, polar floes may be distinguished from annual ice by their color. The meltwater pools on a polar floe have a characteristic pale blue color which persists after the pools freeze. Annual ice is generally grey in summer. Also, annual ice is comparatively smooth, except perhaps for occasional remnants of pressure ridges, while polar ice typically has a hummocky appearance with gently rounded hummocks, each approximately 3 ft in height, spaced at intervals of about 100 to 150 ft. The origin of this relief is the difference between the high albedo of white ice and the low albedo of meltwater pools. Differential melting increases the height contrast each year. Polar floes vary considerably in thickness but have an equilibrium value of about 10 ft in the Arctic Ocean. Each summer about 3 ft of the surface melts and drains off, and each winter about 3 ft freezes onto the bottom.

B.3.2 Occurrence and Classification

Sea ice rarely forms in the open ocean below 60°N but is important in more enclosed bays, rivers, and seas farther south, such as Hudson Bay, the Gulf of

St. Lawrence, and the Sea of Okhotsk. It may also drift south from higher latitudes. Between about 60°N and 75°N, the occurrence of sea ice is a seasonal matter; and there is usually a period during the year when the water is ice-free. Above 75°N there is a more or less permanent ice cover. It must be recognized that generalizations such as these are subject to modifications because of local conditions and seasonal variations. As an example, the Parry Channel (Northwest Passage) cuts through the Canadian Arctic Archipelago at 74 to 75°N. Its eastern end is usually ice-free for 3 to 5 weeks each summer. Its western end is almost invariably ten-tenths covered with ice the year around, partly because of drifting Arctic Ocean ice entering through M'Clure Strait and partly due to local ice. Other examples of waters which are nearly ice-free at some season of the year are Davis Strait and Baffin Bay, the Norwegian Sea, and the near-shore portion of the Barents, Beaufort, Kara, and Laptev seas. The Arctic Ocean is never totally ice-covered. One estimate, based on infrared temperature measurements, is that even during winter as much as 10% of the area of the ocean is either open water or a thin ice cover over refreezing leads.

Most northern countries have developed elaborate terminology and codes for reporting sea ice. The World Meteorological Organization has attempted, without complete success, to standardize the terminology. Most of the terms defined below are in fairly general use. Sea ice is classified by age into young ice (less than 6 in thick), annual (or 1-year ice), and perennial or polar ice. The size of unbroken pieces determine its classification into ice fields (more than 1 mi across), ice floes (33 ft to 1 mi across), ice cakes (6 to 33 ft across), and brash (pieces less than 6 ft in diameter). Other terms relate to the nature of the surface which may be smooth, ridged (when ice pressure has forced ice to buckle upward), or rafted (when one floe has overridden another). Few of the terms above relate specifically to thickness. Some orders of magnitude for typical regions may be useful.

The average maximum thicknesses of unbroken, unrafted floes are as follows: approximately 6 ft in Parry Channel; approximately 8 ft in the northern Canadian archipelago; approximately 10 ft in the Arctic Ocean; and approximately 6 ft in Eurasian arctic seas.

Pressure ridges and rafted ice may be much thicker than these typical figures. There are few accurate data on the height and frequency of pressure ridges, but in the Arctic Ocean heights of 26 to 33 ft are occasionally seen and heights of 13 to 16 ft above the average ice surface are common. To date it has not been possible to determine a roughness coefficient, although the work undertaken by Project AIDJEX should produce many valuable insights. The number of pressure ridges per mile may vary from zero to 60 or more. Because of its density, uniform sea ice floats with 87% of its volume below sea level. When a pressure ridge is formed it tends to sag or slump because of the plasticity of ice. Hydrostatic equilibrium would be reached when 87% of its volume was below water level. Thus, at equilibrium a ridge reaching 13 ft above the water surface should extend 85 ft below the surface. Pack ice is usually too active for equilibrium to be reached but submarine observations show that the bottom relief is similar to that of the top surface except that it is magnified in scale by a factor of 3 to 5.

The date on which freeze-up commences depends on both the oceanographic and meteorologic regimens encountered in an area, but only rarely is the knowledge of ocean dynamics in a particular region extensive enough to forecast the values of the relevant parameters accurately. Consequently, in important areas, it is the practice to dispatch a ship in the fall to take a number of oceanographic stations to determine the depth of the thermocline (if any) and the temperature and salinity profiles above it. From these data may be calculated the quantity of heat per unit area which must be removed for ice to form. The climatological records of the area together with theories on the rate of heat transfer from the ocean to the atmosphere are then used to forecast the date of freeze-up. This is the ice potential method introduced by Zubov in 1938.

While it is possible to forecast with fair accuracy the thickness at any time after freeze-up and the maximum thickness which will be attained, it is much more difficult to predict the rate at which ice will decay, break up into floes, and melt. In most areas affected by sea ice, the earliest date at which a ship can navigate safely varies greatly from year to year as well as from place to place. The approximate opening and closing dates of some arctic ports for conventional ships with some icebreaker support are:

Ambarchik	Late July to late September
Amderma	Late June to late October
Anadyr	Late June to mid-October
Churchill	Mid-July to late October
Dikson	Early July to mid-October
Dudinka	Mid-July to mid-October
Frobisher	Early August to early October
Igarka	Mid-July to mid-October
Kozhevnikova	Mid-July to early October
Pevek	Mid-July to late September
Provideniya	Late June to mid-October
Pt. Barrow	Early August to mid-October
Resolute	Late July to early October
Tiksi	Mid-July to early October
Thule	Mid-July to early October
Tuktoyaktuk	Mid-July to early October

Icebergs are fragments of ice which originated on land. Their average size is the largest of that category - in the Arctic, they are about as big as a city block. Smaller fragments are bergy-bits (about the size of a small cottage) and growlers (about the size of a piano). The greatest number of icebergs in the northern hemisphere is found in Baffin Bay and the Labrador Sea. Predominantly of irregular shape, they are calved principally from the glaciers of Greenland and consequently never reach the central portions of the Arctic Ocean; nevertheless they are of considerable importance because of their movement into the North Atlantic shipping lanes. The principal sources of icebergs within the Arctic Ocean itself are Svalbard, Zemlya Frantsa Iosifa, Novaya Zemlyz and Severnaya Zemlya, but these bergs are relatively small in number and size. Their limited frequency and smallness is attributed to the size of the parent ice caps characterizing the Eurasian arctic archipelagos, which are considerably smaller than the Greenland ice sheet, and also to the relative shallowness of the shelves surrounding those archipelagos.

The largest individual pieces of freshwater ice circulating in the central Arctic Ocean itself are tabular bergs known as ice islands. The source of ice islands is believed to be an ice shelf, or series of ice shelves, off the northern coast of Ellesmere Island. As recently as 1961-62 the largest of these, the Ward Hunt Ice Shelf, lost an area of 300 mi^2 by the breaking away of five large ice islands and a great many fragments. Ice islands first became known when the USAF identified and followed the drift of three of them in 1946 and the years following.

Perhaps the most significant characteristics of both kinds of floating ice are their vast extent and the extreme short- and long-term variability of their distribution. Mainly on the basis of material contained in the U.S. Naval
in thousands of miles
Oceanographic Office's atlases, a recent estimate/of the areal extent and volume of ice of land and sea origin is:

	Icebergs		Sea Ice			
	Volume		Volume		Areal extent, ignoring varying concentrations	
	Maximum	Minimum	Maximum	Minimum	Maximum	Minimum
N. Hemisphere	-	-	8	4	7,000	4,000
S. Hemisphere	-	-	10	2	12,000	6,500
Total	4	3	18	6	19,000	10,000

The maximum and minimum extent of sea ice in the northern hemisphere is shown on the map in Figure III-1. The drift and deformation of ice are primarily determined by the vertical and horizontal transfer of momentum. This is an area of study that has been neglected, perhaps because of the complexity of the problems to be solved as well as the inadequacies of environmental data coverage. The forces involved in the drift of ice are those represented by (1) the stress imparted by the wind to the ice-air interface, (2) the water stress at the water-ice interface, (3) the Coriolis effect, (4) the pressure-gradient force that is due to the tilting of the sea surface on which the ice floats (or "permanent current" effect), and (5) the internal ice resistance. The most comprehensive models for sea ice drift which have been published still do not permit a definite solution. Less sophisticated and considerably simplified studies are generally used in formulating practical forecasts of drift and movement. Although displaying fair reliability, these forecasts do leave much to be desired according to the few evaluations published to date. In the subjective forecasts, drift and deformation are expressed as empirically derived functions of the geostrophic wind, the isobaric gradient, the permanent currents, the quantity of ridging in the pack ice, and the concentration of the ice. The last two parameters are determined from recent aerial observations.

From data obtained by the U.S. Navy's Project BIRD'S EYE it has been found that the arctic pack ice has no consistent thickness; and this is confirmed by the

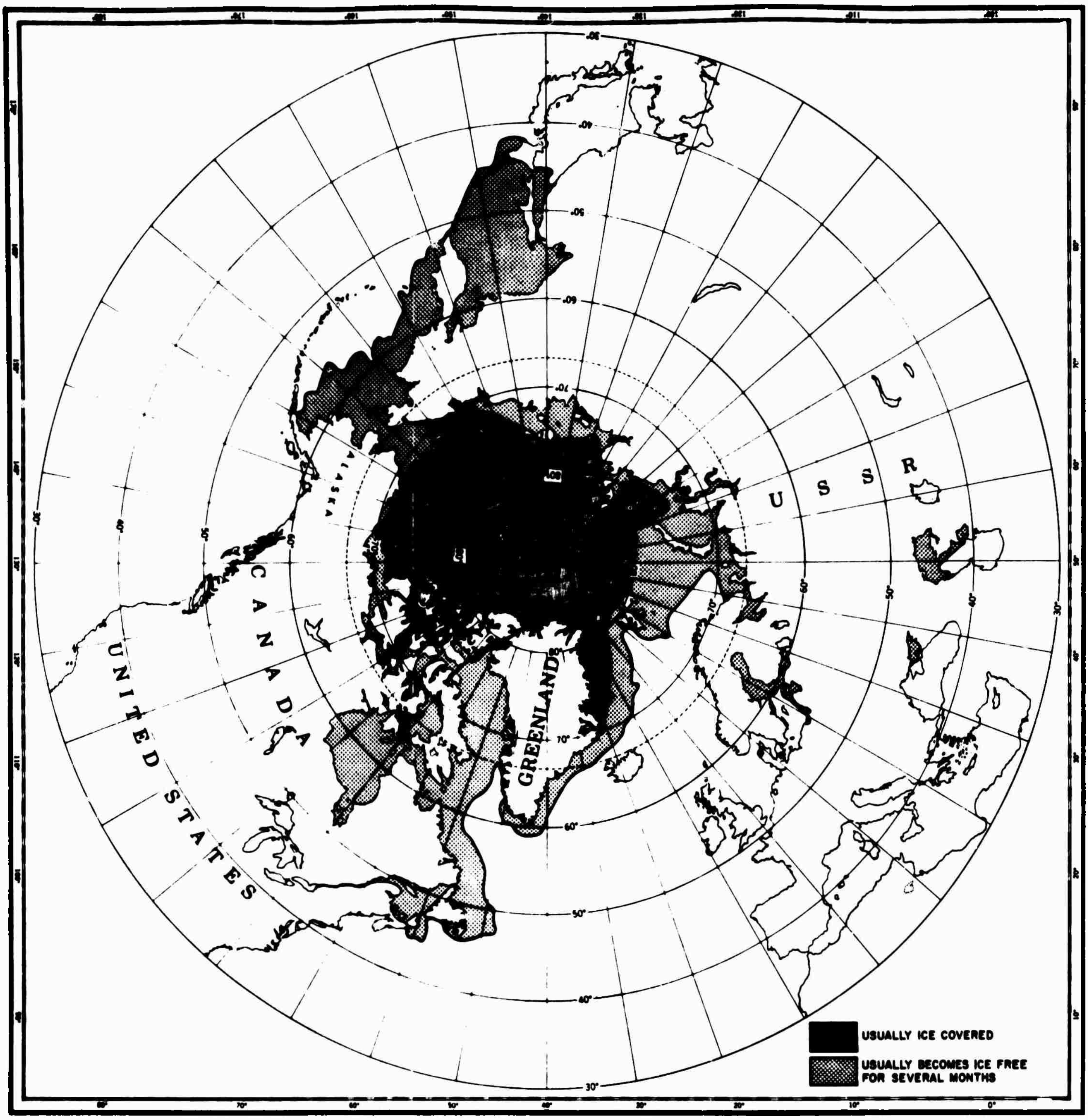

Figure III-1 EXTENT OF SEA ICE IN NORTHERN HEMISPHERE upward-beamed acoustic data obtained during the 1957 through 1962 under-ice cruises of submarines. The constant fracturing, diverging, and compacting motions result in a spectrum of thicknesses ranging from a few inches to more than 15 ft. The data further indicate that the ridges and hummocks - whose drafts may average 40 ft in the summer and 55 ft in the longer winter period - contribute significantly to the total ice volume. Indeed on rare occasions, sea ice ridges may attain a total thickness of 200 ft.

Zones of Open water

Coastal configurations, especially geographic constrictions occurring in the path of a current or wind-driven stream of moving ice, profoundly affect sea ice distribution. The famous North Water, which in mid-winter occupies an area of up to 14,000 mi^2 is found in northern Baffin Bay and Smith Sound. It has long been known by whalers as a prime example of an area of reduced ice concentration that can occur near arctic coasts or coastal regions faced with pack ice. Common smaller-scale coastal phenomena of this type are flaw leads, or zones of navigable size between the fast ice (ice attached to the shore) and the moving pack ice. Both the flaw leads and the large-scale zones such as the North Water may be open or covered with thin ice. They may also contain fragments of thicker drift ice commonly called clutter. These phenomena are not completely understood but are believed to be caused mainly by coastal configurations, wind, and current-induced drift. Tidal openings, however, are also an important contributing cause in regions whose tidal ranges are high or where tidal currents attain a considerable magnitude. Perhaps the waters of Frobisher Bay and Foxe Basin exemplify most outstandingly the considerable effects of tidal parameters on ice conditions and behavior.

B.3.3 Reporting and Forecasting

The practical or operational basis for the need to understand sea ice behavior is the desire for an ability to forecast ice conditions. Three types of prediction are involved: general climatic predictions of probable conditions at a given time of year and place, long-term forecasts of conditions at a specific time and place, and short-term operational forecasts. General climatic predictions are provided by the various available ice atlases, but the quality of the atlases still suffers from the insufficiency of accumulated data.

Denmark produced the first annual survey of arctic ice conditions in 1901 and subsequently issued one each year without interruption, except for the years of World War II, until 1956. Since then she has confined her interest to the regions surrounding Greenland, where U.S. groups are now also making visual aerial observations.

The U.S. Naval Oceanographic Office and the Canadian Meteorological Service have been issuing both long- and short-term ice forecasts for some years, and they have had considerable success in rather generalized forecasts. The U. S. Navy is continuing aerial surveillance of the Bering, Chukchi, and Labrador seas and of Baffin Bay. Canada has an active ice surveillance program in progress over the waters of the Candian archipelago and from the coast of Newfoundland to the Great Lakes. In the Sea of Okhotsk, the Japanese Hydrographic Office is conducting an aerial observation program. In addition, the British Meteorological Office summarizes and publishes hemispheric maps of ice conditions on both a monthly and a shorter-range basis.

While USSR scientists have produced considerable literature dealing with the general patterns of drift within the arctic regions as well as with the theory of ice drift, divergence, and behavior, they have not made available any regular reports of ice conditions or any systematic data concerning the areal distribution of openings in the ice, pressure-ridges, degree of melt, or other ice-water features. The USSR conducts an aerial program in the marginal seas of the

Eurasian Arctic and in the central Arctic where it probably overlaps with Project BIRD'S EYE. The USSR's high-latitude airborne expeditions are of considerable magnitude as a mechanism for the synoptic collection of sea ice data.

B.3.4 Reconnaissance Techniques

Ice reporting and forecasting techniques have been continually developed in the USSR since the 1930's. In the collection of near-synoptic distribution data by aircraft, and in the theory of what may be deduced from it, the USSR is presumed ahead of all other countries. The effect on forecasting theory of the data received from the drifting stations and from the radio beacons left on drifting ice since 1953 (291 up to 1967) has surely been great. But the effectiveness of the forecasts cannot be checked in the West, since neither forecasts nor detailed reports of actual conditions are published regularly. Far from new to the Western specialist, however, are the newest aids in the ice reporting system: the use of helicopters from icebreakers and radio facsimile transmission of ice charts. The USSR system has always relied principally on frequent reports from highly skilled observers. Forecasting techniques include the use of computers, but the results are as yet apparently far from infallible.

At present most ice observations are made visually by trained observers who record their observations with conventional signs on a map and make periodic entries on prepared forms for spot observations. This is a most unsatisfactory method, subject to a great deal of human error and dependent on good visibility. The use of aerial photography has been suggested, but the number of photographs required makes it an impractical aid for daily operational reconnaissance and it is also restricted by visibility conditions. For longer-term synoptic purposes, photography should not be discarded for reconnaissance without much more thorough investigation than it has so far received; in particular as a medium for the new techniques of remote-sensing imagery.

The new advanced Vidicon, automatic picture taking (APT), and high-resolution infrared camera systems contained within the new ESSA and Nimbus polar-orbiting weather satellites promise ultimately to provide the necessary hemispheric - indeed global - sea ice data required not only for operational support by surface ships and submarines but also for a final evaluation of the sea ice mass balance. However, the resolution of the satellite imagery is not now sufficiently accurate to obviate the need for an arctic aerial reconnaissance program. Infrared-scanning imagery sequences now permit a more objective description of water and ridging features in periods of darkness as well as light. Side-looking radar imagery can penetrate cloud layers 20,000 feet thick and promises to provide data during periods of complete undercast when stresses on the ice are at a maximum. Also, a passive-microwave system is under development, the primary objective of which is to acquire imagery which will distinguish sea ice and icebergs from ship hulls.

The relatively wide use of remote sensors in studying sea ice is just beginning at this writing. Orvig's tabulation of areas of open water within pack ice (Table 3-I) which relied on remotely sensed data, is in general agreement with the earlier theoretical studies (roughly 1% open water) and the observations of the British Trans Arctic Expedition. Thus the 10% reported by the Birds-eye Program may not be representative. More accurate figures await a rigorous study of satellite-collected data and the more extensive use of microwave sensors in satellites.

Table 3-I OPEN WATER AREAS IN PACK ICE

ICE AREA (I) OF CONCENTRATION 1.0, AND OPEN WATER (W) WITHIN THE PACK ICE (km² · 10³)

Region	Jan.		Feb.		Mar.		Apr.		May		June		July		Aug.		Sept.		Oct.		Nov.		Dec.	
	I	W	I	W	I	W	I	W	I	W	I	W	I	W	I	W	I	W	I	W	I	W	I	W
Norwegian-Barents Sea	1,600	150	1,756	85	1,901	77	1,736	140	1,600	147	1,290	305	856	510	256	362	214	217	613	322	1,223	205	1,475	105
West Siberian coast	472	15	472	15	472	15	472	15	472	15	472	15	383	59	182	140	152	100	370	92	472	15	472	15
East Siberian coast	743	23	743	23	743	23	743	23	743	23	741	25	635	105	318	148	481	150	670	65	743	23	743	23
Beaufort Sea	407	13	407	13	407	13	407	13	407	13	382	38	357	21	151	89	134	78	349	41	407	13	407	13
Canadian Archipelago	941	29	941	29	941	29	941	29	932	38	915	47	871	68	665	175	607	115	886	69	937	33	941	29
Davis Strait-Baffin Bay	693	20	722	22	726	20	722	34	631	47	479	72	242	116	63	120	30	159	149	50	355	51	663	39
Central Polar Ocean	5,320	61	5,320	61	5,320	61	5,320	61	5,320	61	5,320	61	5,209	172	5,193	188	5,190	191	5,208	173	5,320	61	5,320	61
East Greenland Sea	757	31	807	21	908	31	907	41	733	48	718	44	499	117	230	194	242	103	334	45	391	44	527	33
Total Arctic Ocean	10,933	342	11,168	269	11.418	269	11,248	356	10,838	392	10.317	607	9,052	1,168	7,058	1.416	7,050	1,113	8,579	857	9,848	445	10.548	318

Both the Naval Oceanographic Office and the Cold Regions Research and Engineering Laboratory (CRREL) are analyzing the frequency, distribution, and orientation of pressure ridges from information collected with laser profilometers. The extreme accuracy of measurement of this instrument offers considerable promise of providing quantitative, rather than qualitative, data concerning these parameters. The reports available are preliminary in nature, however the studies are continuing.

Studies carried out on the surface of the ice by personnel from CRREL, the Johns Hopkins University and their contractor indicate that side-looking radar will prove to be a viable tool at this inclination as well as in the higher, airborne mode. Cross-sectional measurements have been made but the low grazing angle mutes pressure ridges and complicates the masking of a snow cover, especially at 94 GHz. The preferred ranges appear to be X-band and 35 GHz. The initial reports of these studies should be available in the summer of 1971.

Under-ice observations of ice distribution have been made for many years by the U. S. Navy but the data is difficult to obtain. Fortunately, the March 1971 cruise of the British nuclear submarine Dreadnaught from Svalboard to the Pole has resulted in a wealth of data for that sector. To quote Swithinbank, "A preliminary analysis of some of the data indicates that the roughest ice bottom topography was found between latitudes 86° and 88°N; this agrees with findings from earlier United States submarine voyages. A simple ice keel count between latitudes 80° and 90°N yields an overall average of 1.6^5 ice keels per km of linear track with drafts of 10 to 15 m, 0.3^2 keels per km with drafts of 15 to 20 m, 0.06 keels per km with drafts of 20 to 25 m, and 0.007 keels per km with drafts of 25 to 30 m. The deepest ice keel reached a depth of 30 m. About 5% of the total under-ice part of the track consisted of open water or young ice."

B.3.5 Behavior

Wind stress appears to be the most important factor in ice drift. The ice moves in a direction about 25° to 30° to the right of the wind direction, because of the Coriolis force. A rule of thumb is that the speed of ice drift is about 2% that of the average wind speed. For the Arctic Ocean, Zubov's rule is frequenly used. This purely empirical quotation is:

$$V = a_i \cdot d_p$$

where V is the drift vector of the ice in km per month, d_p is the magnitude of the pressure gradient in millibars per kilometer (calculated from the average monthly map), and a_i is the isobaric coefficient. The direction of V is assumed to be parallel to the direction of the isobars on the weather map - that is, at right angles to the direction of the pressure gradient. Zubov's original value for a_i, 13,000, was later modified by taking a_i = 9,100 for the period February to April and 12.9OC for the months from August to October. Zubov's rule appears to be applicable only to the Arctic Ocean itself, and cannot be used for the North Atlantic Ocean or Bering or Okhotsk seas.

The pattern of ice movement in the Arctic Ocean is shown in Figure III-2. The two main features are an east-to-west drift on the Soviet side of the North Pole and a clockwise circulation called the Beaufort Gyral in the area between Canada, Alaska, and the Pole. In the Soviet seas there are also local ice circulations which are too complex to show in detail in this figure. The existence of the central current was first established by the voyage of the *Fram* in 1893 to 96. This ship was deliberately frozen into the ice near the Novosibirskiye Ostrova and allowed to drift with the ice. Three years later she became free of the ice near Svalbard. This central current is the main discharge route for ice from the Arctic. After passing between Greenland and Svalbard, the ice follows the coast closely and is borne by the East Greenland current. Each year the ice of this current reaches and passes Kap Farvel, but usually it melts before penetrating far into the Labrador Sea or Davis Strait. However, this does not mean that Labrador and Newfoundland are ice-free. Sea ice formed in Baffin Bay, Davis Strait, and along the Labrador coast is carried southeastward by the Labrador Current, blocking the Strait of Belle Isle (and often the east coast of Newfoundland) until early summer. This is also a region of high incidence of icebergs.

The Beaufort Gyral is a quasi-permanent circulation driven largely by the arctic high pressure system. Ice on the outer edge of the gyral takes about 10 years to complete a circuit, whereas near the center the period of revolution may be as little as 3 years. Ice on the periphery of the gyral may get caught in the central arctic stream and be discharged from the Arctic Ocean, but ice nearer the center circulates indefinitely. This area therefore contains the oldest sea ice in the Arctic.

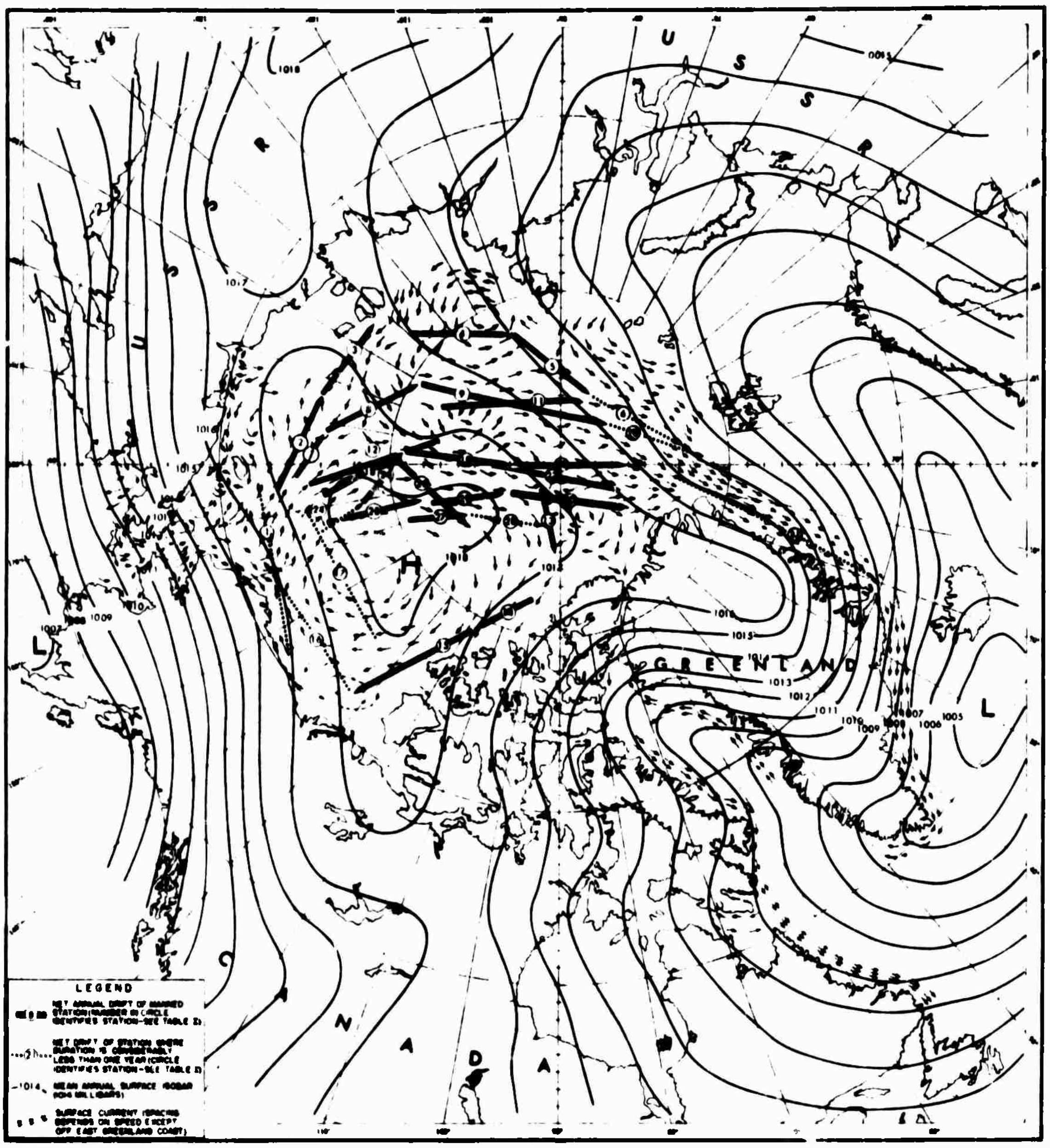

Figure III-2 GENERAL ICE DRIFT PATTERNS AS INDICATED BY MANNED DRIFT STATIONS, MEAN ANNUAL ISOBARS, AND SURFACE CURRENTS

In addition to the ice discharged in the East Greenland current, a certain amount of sea ice moves out of the Arctic Basin to the southeast through the channels of the Canadian archipelago. There is little data available to estimate the fraction of ice discharged by this route. Closely connected with the rate of discharge of ice from the Arctic Ocean is the question of the average age of the ice there. Again the data are too sparse for anything but a very rough estimate of about 10 years.

The ice in estuaries and near them may present more of a hazard to shipping than the usual sea ice. There are a number of major rivers discharging into the Arctic Ocean, including the Mackenzie in Canada and the Lena, Yenisey, and Ob in Siberia. Since settlements tend to occur on and near these rivers, it is a great inconvenience that the estuaries with their freshwater ice usually remain frozen after the nearby sea is open. The Russians have had some success with dusting the ice to hasten the breakup. In this technique any dark, local material is spread, usually by aircraft, on the ice to decrease its albedo and make as much use of solar radiation as possible.

Sea ice in the central arctic grows to a thickness of 6 ft during the first year. Subsequently, it grows more slowly, reaching an equilibrium thickness of 10 to 13 ft after a period of 5 to 8 years. The ice cover is marked by leads and pressure ridges which form as the wind and currents deform it. On the Eurasian side, the ice moves directly toward the opening between Svalbard and Greenland. The drift of ice station Arlis II (Fig. III-3) illustrates this pattern. The swiftest part of the drift occurred off eastern Greenland where steady speeds in excess of one knot were logged. Ice on the Canadian side moves circularly in a clockwise gyre. The drift of Fletcher's Ice Island, T-3, provides an example of this. The ice island made two orbits of the gyre in 17 years. In the vicinity

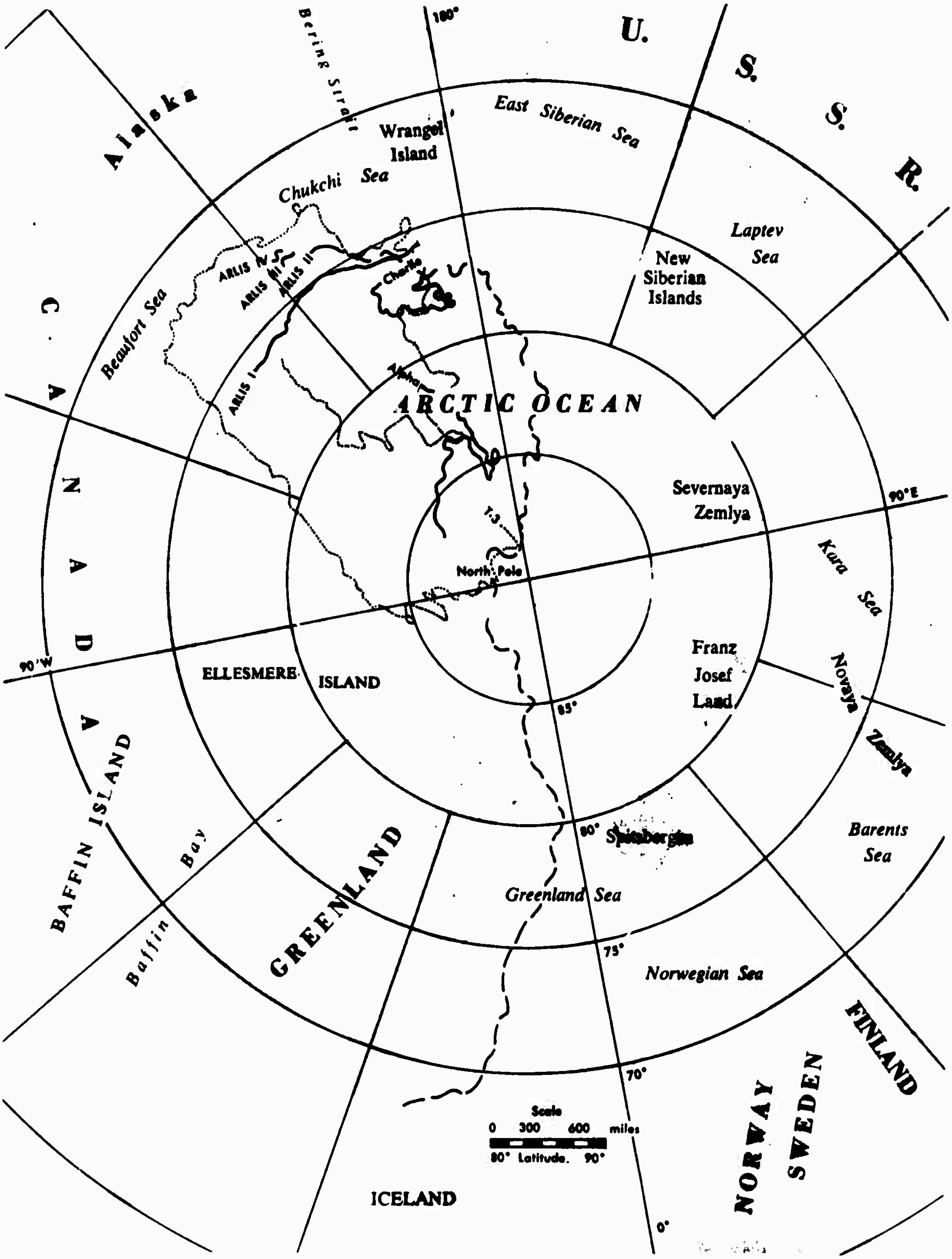

Figure III-3 ICE STATION DRIFTS

of the North Pole there is an ice divergence which separates the two patterns of drift. The exact location of the divergence varies from year to year. Even though the tracks of Arlis II and T-3 were similar on the Canadian side, one exited into the East Greenland Current and the other remains locked in the gyre.

Superimposed on these broad patterns are many irregularities of drift. The small-scale motions are due to wind and are particularly noticeable in the T-3 track. There is a still-smaller scale of ice motion with a period of about 12 hours which can only be detected with the more precise and frequent fixes obtained with satellites. The ice describes clockwise circles with a diameter of approximately one-half mile. These are inertial motions representing the transient response of the ice to changing wind conditions.

In many areas around the Arctic Basin there is, for greater or lesser parts of the year, a belt of ice which is fastened to the shore. This ice is commonly smooth and unbroken, but there may be deep snow drifts or large meltwater pools depending on the season. As the season progresses the shorefast ice breaks apart and either melts or joins the drifting pack. The amount of shorefast ice varies greatly. In most parts of the Arctic Basin some knowledge exists of where and when it has occurred over several years, where it is consistent and dependable, and where it is extremely variable from year to year. The most extensive areas of smooth ice, not strictly shorefast but included herein this general classification, are found in the Greenland fiords, especially in Peary Land, and on bays and sounds of the Canadian archipelago, Svalbard, and Zemlya Frantsa Iosifa.

B.3.6 Distribution

The geographical distribution of sea ice in the northeastward extension of the Atlantic Ocean is high asymmetrical at all seasons. The striking feature is the absence of ice in the southern and east sectors of this part of the ocean, the Norwegian Sea, resulting from the ameliorating effect of warm water and air. The most extreme example is northwest of Svaldbard where open water is normally found in winter as far north as 80°. This is over 800 mi closer to the North Pole than any other open circumpolar sea at this season.

The seas around Iceland occasionally were reached by pack ice before the climatic improvement of most of the present century. In an unusual year in the nineteenth century, ice appeared off the north coast in December 1887 and by the following June had spread clockwise round the island until only the west coast was ice-free. In the last thousand years the island has been surrounded by ice probably several times each century, although the variations from one century to another have been considerable. Since 1920 sea ice has appeared off the north coast of Iceland on only a few occasions, except for the last three years during which conditions have deteriorated considerably.

Jan Mayen and Bjornoya are normally reached by pack ice and remain surrounded for 2 to 4 months. Although as the ice margin is not far away during this period, conditions are variable.

The sea ice off East Greenland originates primarily in the Arctic Basin. Together with smaller amounts of ice that freeze locally and icebergs from glaciers, pack ice is carried south between Svalbard and Greenland by the East Greenland Current as a broad belt of ice that is about 270 mi wide north of Scoresby Sund (70°N) in late winter. North of this latitude there is sea ice at all times of the year, although the belt narrows and contains open water in late August. Beginning in September the southern and southeastern margin of the pack ice expands southward until by December or January a tongue of pack ice has reached as far as Kap Farvel and has effectively blocked the east coast. While the pack ice develops offshore, the fiords begin to freeze in October, and this ice remains until the following June or July. When the fiords clear of fast ice they are commonly filled with pack ice blown in by onshore winds. The maximum extent of East Greenland ice is reached in March and April, and in very heavy ice years there is continuous pack ice from southern Greenland to Svalbard and eastward to Novaya Zemlya. The margin of the ice off Greenland begins to withdraw in May by melting on the outer side and by the developing inshore leads between the coast and the open pack.

In the most favorable years, the limit of pack ice in August may be as far north as 75°, while in a bad year ice may still reach Kap Farvel at this time. Great fluctuations have been recognized in the extent of the East Greenland ice, and this is clearly related to the rate of transport by the current and amount of ice that exits from the Arctic Ocean. Although in general it appears to have been less in the 20th century than in the previous three centuries, forecasting the quantity and distribution of the East Greenland ice is still not possible.

Only three glacier systems in East Greenland are significant producers of icebergs. The most active glaciers calve into Scoresby Sund and may contribute 40% of the total iceberg production. Iceberg calving begins in July and continues through the latter part of the summer. Few icebergs actually reach the polar pack and move south with it as the majority become stranded in the fiords or trapped in the many offshore islands. The smaller the concentration of pack ice, the more icebergs move south because they are less apt to be blocked by the pack.

In Peary Land, fast ice remains in the fiords from September until the end of July and is rarely broken up by winds once it has formed. In unusually cold summers it may fail to melt. This ice may survive several decades and becomes fresh and re-crystallised; only with difficulty can it be differentiated from glacier ice. It is found in several fiords, including J. P. Koch's Fiord and Independence Fiord. On some coasts in North Greenland the fast ice is practically permanent, but the ice breaks up in most fiords and for a short period there is open water along the coast. Throughout the year heavy pack ice is present off the coast.

The west coast of Svalbard is free of ice or experiences only limited severe ice conditions, while the east and north coasts experience very heavy pack ice. Fast ice generally develops on the west coast of Vestspitsbergen in December, but in some years a continuous ice cover fails to form in winter. This inshore ice melts in April or May; it may be replaced by pack ice later in the summer in the south, but the area north of Isfjorden is normally ice-free. On the east coast, local fast ice, icebergs from glaciers, and pack ice that has been carried from the east block the coast for the first half of the year. Storfjorden and Hinlopenstredet between Vestspitsbergen and Nordaustlandet may remain blocked throughout the summer. When the ice begins to loosen up with summer melting, it is carried by currents round the southern tip of Vestspitsbergen (Sorkapp) and north along the west coast, in some years reaching Prince Karls Forland by the beginning of July. The northwest part and the western part of the north coast are generally the most accessible.

The opposing coasts of Baffin Bay and Davis Strait show startling contrasts in sea ice conditions. Due to the warm current the southwest coast of Greenland is commonly ice-free even in winter, except for the inner fiords, as far north as Disko Bugt. In summer the whole coast may be free of pack ice although in Melville Bugt icebergs occur. However, in bad years ice is carried round Kap Farvel from East Greenland in summer and blocks the southwest coast.

Ice conditions on the Canadian side are much less favorable. In winter heavy fast and pack ice blocks all coasts and moving pack stretches across Baffin Bay. Open water may occur, particularly in late winter in the extreme north of Baffin Bay and off Baffin Island between the moving pack and the fixed inshore ice. Condtions improve only slowly in early summer and heavy pack may continue to be carried south well into August. In favorable years the ice will disappear about this time except for icebergs. In bad ice years some pack remains inshore and heavy pack ice may survive in the middle and north of Baffin Bay.

The northern channels that lead to the Lincoln Sea have great ice variations from day to day and from one year to another. In general, Kane Basin, Hall Basin, and Robeson Channel provide greatest difficulty to ships. Robeson Channel in particular may be so jammed with ice throughout the summer that it remains impassable even to icebreakers.

Jones Sound which is the normal route for ships making for Eureka can be entered either through Glacier or Lady Ann straits. The ice begins to loosen in July and in good years will be clear from mid-August until the second half of September.

Parry Channel normally shows a range of ice conditions from heavy ice at the western end throughout the summer to navigable conditions in Lancaster Sound at the east end beginning in mid-June and lasting until the end of September. Summer conditions in Lancaster Sound are helped by the presence of open water early in the year and sometimes throughout the winter in Barrow Strait.

The ice in the coastal waters of Labrador has three sources: local freezing, Baffin Bay ice carried into the area by the Labrador Current, and icebergs which originated in northwest Greenland. The pack ice extends southward with the onset of winter and is off south Labrador by early December. A month later it is off Notre Dame Bay (Newfoundland), and by March when the sea ice is at its maximum the whole Labrador-Newfoundland coast as far as the Avalon Peninsula is infested. Breakup begins in the southern part in April and slowly extends northward in the next two months to Strait of Belle Isle and more rapidly by mid-July to Hudson Strait.

Freeze-up in the Gulf of St. Lawrence begins on the north shore in December and extends southward with the addition of ice that enters through the Strait of Belle Isle. By late January the Gulf is 80% ice-covered except for the southwest corner of Newfoundland which remains clear a few more weeks. There is sufficient open water throughout the winter for shipping to be possible through the Gulf, although it may be locally restricted by large ice fields, particularly in Cabot Strait. The ice cover decreases rapidly in early April and within a few weeks ceases to be a hazard except in the northeast corner of the Gulf.

Ice conditions in Hudson Strait are highly variable. In addition to that which develops *in situ*, ice is carried into Hudson Strait from Foxe Channel, Hudson Bay, and Davis Strait at different times of the year. At the beginning of the season, pack ice is often present at the west end of Hudson Strait where it has been brought in from Foxe Channel. Inshore ice on the Baffin coast begins to form in mid-October but there is not much growth until early November in the west and until the middle of the month at the east end. Much of this early fast ice breaks away to form the winter pack ice in the Strait. There is a heavy cover of moving pack ice between early December and the end of May. There is often appreciable open water, particularly on the north side, but navigation is impossible. The breakup is prolonged as the ice that moves out into the Atlantic is replaced from the west. In an average year the straits are open in the third or fourth week of July.

Sea ice usually begins to appear in Hudson Bay about the middle of October and reaches a thickness of about 5 ft by December. Most harbors on the west coast are normally blocked by late October, but Churchill may remain open into November. An ice-covered zone gradually builds up along the shores with moving ice fields in the middle of the Bay. The Bay is not covered until late December or January; and even then in the northern part of Hudson Bay temporary ice-free areas may develop after continuous northwesterly and northerly winds. Melting begins in May and open water increases rapidly in early June. Winds play an important part in slowing down or speeding up the removal of the ice; on the whole the prevailing winds tend to clear the northern part earlier. Ice fields are still present in many years during late July, but by August the ice has usually disappeared.

Ice conditions in Foxe Channel and Foxe Basin are generally severe. The ice is extremely rough and often dirty in appearance because of the incorporation of bottom muds as it forms in shallow water. Fast ice begins to develop in the east side of the Basin early in October. Pack ice in the center and west extends the ice cover until, by late December, Foxe Basin and Channel are completely covered. The pack on the west and south sides is on the move throughout the winter, and open water may be found at many points. The first significant clearing of sea ice occurs in late June in Frozen Strait and Foxe Channel. This is, however, temporary; and when the main pack becomes mobile in late July, areas that were formerly open are again closed. Ice conditions remain difficult even on the west side during the first half of August, and only by late August is part of the Basin (depending on winds) likely to be open.

The channels of the Queen Elizabeth Islands present ice conditions for surface navigation second in difficulty only to those of the Arctic Ocean. Throughout most of the straits there is a continuous ice cover at practically all times of the year, and only in the more favorable summers are there openings in the form of small leads. Ice conditions are best in the east. Parry Channel, which forms the southern margin of the area, has relatively clear summer conditions in Lancaster Sound, but these deteriorate further west.

At the west end of Barrow Strait the ice is at a minimum in mid- and late August, although generally a good deal of ice remains throughout the summer, particularly if strong westerly winds blow it eastward from Viscount Melville Sound. Farther west in Parry Channel, there are considerable quantities of ice at all times of the year; and M'Clure Strait has never been known to be ice-free, although in occasional and unusual years small inshore leads develop. Generally, conditions along the north side of Parry Channel are better than the south because of the tendency for northwesterly winds to blow the ice southward. In winter, and especially in spring, areas of open water may be found in the eastern part of Parry Channel, particularly at the west end of Barrow Strait.

The best approach to the inner channels is through either Wellington Channel or Jones Sound. The former has experienced several areas of open water in spring, and the expansion of these produces fairly early melting. Thus, by late June there may be considerable open water in the north of Wellington Channel and in the vicinity of Belcher Channel; however, the ice may not clear from the southern part until late in the summer. Norwegian Bay, between Axel Heiberg, southwestern Ellesmere, and Grinnel Peninsula, is normally ice-covered from late September to late July. Navigation may be possible during mid-August, and conditions improve northward toward Eureka Sound, where the strong currents contribute to an early breakup. Elsewhere the channels remain ice-covered throughout the year. The movement of ice islands through the channels from the north and northwest shows that the ice tends to move as a unit, although probably only in good years. From all practical points of view the remaining waters may be considered unnavigable by surface ships at all times.

The character and distribution of the sea ice in the southwestern portion of the archipelago is profoundly influenced by the low tidal range which permits fast ice to develop in all but a few localities during the winter. The absence of strong tidal currents except in a few of the narrower straits (e.g. Dease Strait) enables the ice to form quickly in the fall; and it is only where there are strong tidal currents that channels remain open as late as December. The comparative summer warmth of the southern channels and the presence of many large rivers flowing off warm land contribute to early and continuous melting of the sea ice. In contrast, the channels that open to the north become jammed with ice under the influence of the northwesterly winds, and some waters - notably M'Clintock Channel - may never be clear of ice. In the northerly channels it is often true that ice conditions are better in the early summer than later in the year, by which time much of the ice has broken up farther north and blown into them.

Along the mainland coast the first open water appears near the mouths of the large rivers. The southern part of Bathurst Inlet normally has extensive open water by the end of June, and in an average year much more open water is evident by the second or third week of July when, except in very bad years, navigation is possible all along the coast. Heavy pack ice may exist until well into August in Coronation Gulf and until even later at the east end of Queen Maud Gulf. The freeze-up in these areas begins in late September or early October and spreads comparatively quickly on the brackish water close to the large rivers. Hence, by the end of October the coasts are frozen again, and by early November there is once more continuous ice except in narrow channels.

In contrast to the ice conditions in mainland waters, the south sides of Viscount Melville Sound and M'Clintock Channel contain the most heavily ice-infested waters south of the Queen Elizabeth Islands. The winter ice in this area, except in the bays is rough and contains a proportion of perennial pack ice. There is some evidence that ice in the midde of M'Clintock Channel is on the move even in midwinter. Leads may develop in early August in these areas, particularly adjacent to the river mouths; and with favorable offshore winds, a shore lead may persist although it can never be relied upon. The larger bays (e.g. south Hadley Bay) begin to open at the head by the end of July and become ice-free toward the end of most summers, but there is always a danger that ice will be pushed back in. The conditions at the south end of M'Clintock Channel are made more difficult by the inability of the ice to escape from it. In the early 1960's an ice island from the Arctic Ocean moved south through M'Clintock Channel and became wedged in the Victoria Strait area.

Prince of Wales Strait has extremely variable ice conditions. In most summers passage is possible at some time, although long periods of waiting may be expected. It is likely that small craft can negotiate the straits in late August after the ice has melted in the northern part of the Strait and before it is filled from Viscount Melville Sound. Similar conditions are found in the east where Peel Sound has an extremely variable record. In Franklin Strait conditions are not as good; and it seems unlikely that this sector is ever completely clear of ice, although in late August a shallow-draft vessel may occasionally achieve a successful passage.

The sea ice conditions in the Beaufort Sea during the summer vary greatly and only general statements are possible. The clockwise current tends to force the permanent polar pack into the southeastern Beaufort Sea. However, this drift is counteracted effectively by the dominant southeasterly winds. Therefore, at freeze-up in late fall, the margin of the pack has retreated about 100 mi from the mainland near the Mackenzie Delta. Normally it is close off northern Banks Island and Point Barrow.

During the first week of June the only open water along the coast is found in Shallow Bay on the west side of the Mackenzie Delta. By the following week Kugmallit Bay opens on the east side, and Tuktoyaktuk is usually accessible from the Mackenzie by the end of the month. Along the Yukon coast the open water extends slowly westward, affording the possibility, in an average year, of reaching Herschel Island during the second week of July and Point Barrow by the third week. Large leads in Amundsen Gulf absorb solar radiation and contribute to an early breakup in this area. The Thesiger Bay coast is usually free of ice by July first. Summer ice conditions off Banks Island depend on the location of the polar pack. At the end of July ships can normally proceed north of Cape Kellett. Beyond Storkerson Bay, however, the sea is never free of ice, and in bad years only shore leads are open.

Between Herschel Island and Point Barrow the edge of the polar pack closely approaches the Alaskan coast, generally leaving a channel several miles wide in late summer that may be closed at any time if the pack surges south. The polar pack in this area has a clockwise movement and may contain large ice islands.

Drifting pack ice covers most of the Arctic Ocean. In the pack there are four areas where the ice is quite distinct from that in the other areas. In the southern Beaufort gyre where it impinges on Point Barrow and across the head of the Bering Strait, the ice, having come slowly down along the Canadian archipelago and across the Beaufort Sea, is relatively old. It is mostly thick ice blocks or floes which have been jostled and compressed by being squeezed by the northward trend of the Alaskan coast, and the thin ice has been pushed and pressed onto the thicker ice. It is fairly warm north of Alaska compared to many other parts of the Arctic, so melting of the thin ice takes place there. Nevertheless, the pack ice, from having been under compression, is rough-surfaced, and what is there is mainly fairly old ice. This ice remains almost unchanged as it goes past the northeastern coast of Siberia and drifts toward the North Pole. Through that area, there are many old ice floes. Quite often these floes are measured in fractions of a mile, but in most areas there are some that are measured in miles. The floes are separated by pressure ridges; by and large, the larger the floes, the larger are the pressure ridges between them. For most of the year this region is nearly completely ice-covered with only small areas of open water. This region gives us the classic picture of Arctic Ocean sea ice.

In the middle of the Beaufort gyre there are from several hundreds to thousands of square miles where the ice does not move as much and is not subjected to great shearing stresses. Usually turning more or less on its axis, the ice does move, of course, when subjected to the considerable stresses due to storms, but there is not a constant translatory motion. Here the ice floes are thicker and generally smaller than in the outer parts of the gyre; the pressure ridges are not so well defined and are mostly short and branching, running in various directions; and the general topography is rougher with not very much open water.

Still within this western gyre, there is another zone in the southeastern Beaufort Sea itself where the ice pack expands after having been compressed against the Canadian islands and is allowed to diverge and spread out as it moves on toward the delta of the Mackenzie River. This is also a warm area--the amount of sea ice melting is perhaps as great as anywhere in the whole of the Arctic - so every autumn along the shore there are large regions of open water that later freeze into young ice. As the ice passes Banks Island and turns the right-angle corner to go westward past Alaska, the flow pattern which has been generally in extension changes to a compressive pattern of movement. Because most of the pack consists of pieces of multi-year ice separated from one another with the spaces in between filled with one- or two-year-old ice, the result is a frozen mixture perhaps 6 ft thick of varying ice types. It is very difficult for a ship to get through this ice, although it is not as thick as ice elsewhere, and it has a fairly uniform surface roughness in comparison with the wide range of roughness in the middle of the ocean where there is a greater contrast between smooth floes and well-organized ridges.

On the Eurasian side of the Arctic Ocean, the U-shaped current pattern comes out of the Atlantic and goes back into it again. By the very nature of this pattern the ice does not stay long in the Arctic Ocean. The average ice thickness therefore is less than in the Beaufort gyre. On the basis of the observations that have been published, mostly by the USSR, the ice in a smooth floe is seldom more than 7 ft thick. It is difficult to determine the age of this ice, but extremely thick old floes are not a conspicuous part of the pack ice in this area. In the areas from about 85° latitude to the North Pole there are, depending on the local history, fairly smooth regions with well-organized ridges separating them. Sometimes it is possible to see a regional pattern in the organization of ridges through the area. Satellite photographs often give a good indication of the texture of this part of the pack and these seem to be fairly constant.

The marginal seas of the northern Pacific have no direct, unrestricted link with the Arctic Basin, and ice from the polar pack never reaches them. In the Bering Sea the local pack ice reaches its maximum extent in April when its margin lies roughly from the tip of the Alaskan Peninsula slightly south of the Pribilof Islands to the east coast of Kamchatka. Open pack ice present along the Kamchatka coast poses no problems to navigation at this time. The ice in Bristol Bay begins to break up in May. By early June, Bristol Bay is completely ice-free, and the ice has retreated to St. Lawrence Island. By mid-June Norton Sound is ice-free and by July the Bering Sea is completely free, the Siberian side clearing before the Alaskan. The first pack ice reforms in Bering Strait during the first two weeks in October.

Ice conditions in the Sea of Okhotsk closely resemble those in Hudson Bay. Ice usually forms in the northern bays in late October or early November. It covers the bay completely by December, and by this time there is floating ice throughout the northern half of the sea. Ice continues to form under the influence of the intensely cold continental air mass until April, and the strong winds carry it toward the Kuril Islands until the whole sea is covered. There is drift ice in the Kuril Straits from January until March, the amount depending on the strength and duration of the winter monsoon. Sometimes the ice moves well into the Pacific before melting completely. In the north the breakup comes in May, starting with the river meltwater which forces the landfast ice out to sea. The Shantar Islands are sometimes blocked until July, but in general the sea is ice-free in June.

Ice conditions are also variable in the Chuckchi Sea as in the Beaufort Sea. When the volume of water coming through Bering Strait in summer is large, it forces the margin of the polar pack ice north toward the edge of the continental shelf, at 72 to 73°N; but continuous strong northerly and easterly winds will drive the ice back to the coast near Point Barrow and into Proliv Longa. Ice begins to form in September in northern coastal parts of the sea while in sheltered inlets such as Kotzebue Sound, freeze-up occurs by mid- or late October. On open coasts fast ice may not form for another month. Offshore, roughly beyond the 80-ft submarine contour, heavy pack ice is on the move all winter. Breakup in coastal areas occurs during June, and the southern half of the sea is navigable from mid-July to mid- or late September. Farther north shipping is restricted until mid-August.

The ice conditions of the Laptev and the East Siberian seas are more severe than in any other of the peripheral seas off northern Asia. Sea ice forms unusually quickly as the water temperatures of the shallow seas drop rapidly in the autumn. A zone of hummocky ice as much as 30 ft high is common at the boundary of the fast ice and pack ice and is most strongly developed in the western Laptev Sea and in Proliv Longa. In the spring the ice melts first at the mouths of rivers and along the nearby coasts. The rivers freeze between October and December, and their flow throughout the winter either ceases or is quite low. Breakup is at the end of May or the beginning of June. On large rivers, with ice up to 8 ft thick, this involves violent movement, ice dams, and flooding. The ice on the smaller rivers melts in place. The navigation period for the lower reaches of the rivers lasts from 3 to 4 months.

In summer, Proliv Dmitriya Lapteva tends to have favorable ice conditions but Proliv Sannikova and Proliv Longa are rarely completely free of ice, and Proliv Vil'kitskogo has the worst ice conditions. This is caused by drift of ice with the current from the Kara Sea and by pack ice and icebergs that drift southeast from Severnaya Zemlya.

Breakup begins in June off the deltas of the Lena and Yana rivers. The eastern side of the Laptev Sea is the first to be free of ice because of the influx of river water and the action of winds blowing the ice offshore. Changes in the western part are slower because of the deeper water and the cold current that flows south along the east coast of Severnaya Zemlya. Once the ice has broken up in the west, the ice margin retreats at 10 to 25 mi a day. The southern part of the Laptev Sea is navigable in July, but there may not be access to it from the seas on either side. Heavy pack ice from the Arctic Basin does not extend far into the Laptev Sea (unlike the East Siberian Sea) and only small quantities find their way south. The navigation season is effectively August and September.

Freeze-up begins at Tiksi, in Guba Buorkhaya, in early October and spreads both east and west from there. By December the sea is completely frozen over. Ice conditions vary from the west to the east. In the west there is much hummocky and rafted ice caused by wind stress. There are often large fields of heavy sea ice off Severnaya Zemlya and Poluostrov Taimyr where heavy floes and icebergs may run aground as they are driven southward. In the eastern part, the ice is generally smoother and from 60 to 90 in thick. The fast ice is 10 to 15 mi wide off the northeastern parts of the Taimyr, 50 to 60 mi wide along the southern shores of the sea, and 250 mi wide from the mainland to north of the Novosibirskiye Ostrova.

In the East Siberian Sea the first open water appears in late June off the Kolyma delta. Open water expands rapidly west of the Kolyma and the ice edge moves to the north. In the eastern sector of the sea, the ice does not begin to clear for 1 to 2 weeks longer. The navigation season is 8 to 10 weeks long in the west and 6 to 8 weeks in the east. During this period, ice conditions in the western sector are usually favorable for navigation. Proliv Dmitriya Lapteva is clear of ice and in the late summer it is sometimes possible to pass north of

Novosibirskiye Ostrova. The Ostrova De Longa are usually surrounded by drifting ice and bergy bits that have calved from small glaciers on the islands. In the eastern section, conditions are generally neither as good nor as predictable. The heavy pack of the Arctic Basin approaches the coast more frequently here than anywhere else along the whole north Siberian shore. However, during the late summer, there is usually an open lead close to the shore up to 20 mi wide.

Freeze-up begins in the west and by late November there may be an unbroken surface extending 180 to 300 mi offshore in the west and narrowing to 10 to 20 mi off Mys Shmidta. Ice is constantly on the move in Proliv Longa.

Around Ostrov Vrangelya navigation is possible only during the last half of August and September. Proliv Dmitriya Lapteva is normally open for navigation from late July or early August through September. The seas surrounding the Novosibirskiye Ostrova are rarely free to pack ice, although the concentration is at its minimum in September.

There is floating ice all year in the Kara Sea, the amount and distribution depending largely on the strength and direction of the prevailing winds. Most of the ice originates in the Kara Sea itself, because currents and shallow water in the north act as barriers to the heavier pack ice of the Arctic Basin. Breakup advances from the south, where it is essentially open by July. Optimum navigation conditions are in August and September, but in good years they last from July to October. The worst conditions for navigation are usually encountered near Poluostrov Taimyr where ice collects in the narrow straits separating the islands of Arkhipelag Nordenshel'da and blocks the area immediately west of Proliv Vil'kitskogo. Conditions are quite variable elsewhere in the Kara Sea, but ice is apt to be heavy either in the northwest or in the southwest. Small icebergs may be found off northern Novaya Zemlya and Severnaya Zemlya.

Mainland rivers begin to freeze in October at the river mouths, and at the same time fast ice develops around their mouths. Fast ice forms quickly and, anchored by the many islands, remains all winter. Along eastern Novaya Zemlya fast ice is 3 to 5 mi wide, and north of the mainland it may exceed 15 mi in width. The pack in the central Kara Sea is 6 to 10 ft thick but considerably thicker where it is ridged or hummocked by pressure. Grounded hummocks, called 'Stamukhi', are common off Novaya Zemlya and Severnaya Zemlya. They reach heights of 30 ft and tend to retain the surrounding ice cover in position. Ice conditions around Severnaya Zemlya are variable. The navigation season is usually from mid-August to mid-September. East of the archipelago there is ice all year, including some icebergs. East winds drive much of this ice between the islands making those straits impassable.

The Barents Sea is exceptionally ice-free when compared with the other peripheral seas of northern U.S.S.R. This is a consequence of the warm Atlantic water which enters the sea. The inflow is sufficient to keep the southwestern sector free of sea ice all year, and it causes ice conditions in the rest of the sea to be much less severe than is normal for the latitude. Ice conditions may be predicted by measuring the flow of the North Atlantic Drift north of Scotland. Warmer water there is followed 2 years later by better ice conditions in the Barents Sea. Negligible quantities of ice enter the Barents Sea from the Kara Sea and the Arctic Ocean, and all sea ice may be considered local. Maximum ice cover is in April, when the sea is 75% covered. The mean southerly limit of unnavigable ice at this time runs from the southwest coast of Svalbard to Bjornoya and east to 73° 30'N, 40°E and the mainland at Mys Svatoy Nos. Significantly the Murman coast is always open. In the spring the ice margin recedes east and north, and by mid-June it can be navigated north of Bjornoya and east to 50°E. At Ostrov Kolguyev in the southeast of the Barents Sea, ice lasts until the end of July. Ice conditions are best in September when the mean ice limit runs from southern Svalbard to Zemlya Frantsa Iosifa and continues east-southeast to 40 mi north of Novaya Zemlya. Occasionally in late August the Barents Sea is entirely free of ice.

Freeze-up begins in October in the shallow waters around Zemlya Frantsa Iosifa, Novaya Zemlya, Svalbard, and the estuary of Pechora. During November the west coast of Novaya Zemlya freezes, and by December most of the sea north of 75°N is frozen. The ice margin moves gradually south and west as the winter progresses. Winds and currents have a strong effect on the ice conditions because there is much open water in the west. Between February and April, if southeast winds are unusually strong, the ice will be driven to the northeast, leaving much open water in the southwestern quadrant of the sea. Zemlya Frantsa Iosifa is rarely clear of ice. The southern portion of the archipelago may be ice-free for a few weeks but the narrower straits are icebound all year. Western Novaya Zemlya is clear of ice from the end of July until October. In January there is much loose pack ice to the west of the island. From February to April Novaya Zemlya and Zemlya Frantsa Iosifa are completely surrounded by landfast ice, with breakup beginning in May. Small icebergs may be found in the vicinity of both island groups since both have glaciers which reach the sea.

Ice conditions in the White Sea vary radically from year to year as a result of the strength and direction of prevailing winds. In October fast ice begins to spread from gulfs and bights. By January the sea is frozen over, but ice is very thin in the central part and the entrance to the sea is still ice-free. Maximum ice occurs in February and March and by then ice is thickest in the entrance which is still blocked in May. Grounded hummocks form a chain north of Ostrov Morzhovets. Breakup begins in late April and by June the sea is ice-free again. The ice is usually forced out into the Barents Sea to melt, and consequently Mys Kanin Nos is often surrounded by floating ice until the end of June.

B.3.7 Drifting Stations

Because of their relative durability and availability, ice floes and ice islands have been used for many years as platforms from which to study the arctic oceanic environment. Since the early 1950's over 20 such stations have been occupied, primarily by the Soviet Union, on a long-term basis (see Fig. III-4). Semi-permanent living and working quarters were erected and a varying number of vehicles provided. As many as 8 to 10 disciplinary studies may be conducted at a station and, in the Soviet case, helicopters and small fixed-wing aircraft may add immensely to the areal extent of the programs. The only limitation on the number of personnel is budgetary although the usual complement is from 20 to 40. If conditions permit, the station may be supplied by icebreaker, but usually a landing strip is required for long-range aircraft. In this the Soviets excel as witnessed by their high-latitude expeditions which are mounted each spring and involve tens of aircraft and hundreds of people.

Drawbacks to the use of such stations include their susceptibility to fracturing and the inability to control their course. There is, of course, no assurance that a suitable site will be avilable in a given area at a specific time, so planning must be flexible. Nevertheless, drifting ice stations provide a valuable base for selected operations in the Arctic Basin.

B.3.8 Bibliography

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Table III-4 DRIFT POSITIONS OF SHIPS AND DRIFTING STATIONS

VECTOR NO.	STATION	DATE		POSITION		SPEED (cm/sec)
		Begin	End	Begin	End	
1	*Maud*	24 Sept 22	24 Sept 23	73.0°N 173.3°W	75.5°N 163.7°E	2
2	*Maud*	24 Sept 23	8 Aug 24	75.5°N 163.7°W	76.5°N 143.2°E	2
3	*Fram*	25 Nov 93	24 Nov 94	78.6°N 139.1°W	82.0°N 112.0°E	2
4	*Fram*	24 Nov 94	22 Nov 95	82.0°N 112.0°W	85.8°N 64.2°E	2
5	NP-6	19 Apr 56	19 Apr 57	74.5°N 177.0°W	76.0°W 171.7°E	1
6	NP-6	19 Apr 57	8 Apr 58	76.0°N 171.7°E	80.9°N 150.2°E	2
7	NP-6	8 Apr 58	12 Apr 59	80.9°N 150.2°E	86.3°N 39.6°E	5
8	NP-5	16 Apr 55	16 Apr 56	82.2°N 156.2°E	86.3°N 89.3°E	3
9	T-3*	July 50	July 51	75.5°N 175.0°W	83.0°N 169.0°E	3
10	T-3	1 June 52	4 June 53	88.5°N 140.0°W	85.2°N 92.0°W	1
11	T-3	14 July 57	14 July 58	82.2°N 102.2°W	79.5°N 116.6°W	1
12	T-3	14 July 58	2 July 59	79.5°N 116.6°W	72.0°N 132.5°W	3
13	T-3	9 June 59	7 Mar 60	72.3°N 130.5°W	71.8°N 150.5°W	3
14	NP-4	8 Apr 54	17 Apr 55	75.8°N 171.6°W	80.6°N 176.1°E	2
15	NP-4	17 Apr 55	19 Apr 56	80.6°N 176.1°E	87.4°N 178.5°W	2
16	NP-4	19 Apr 56	20 Apr 57	87.4°N 178.5°W	85.9°N 0.0°	2
17	NP-3	9 Apr 54	20 Apr 55	86.0°N 178.0°W	86.4°N 24.0°W	3
18	NP-2	12 Apr 50	11 Apr 51	76.0°N 166.6°W	81.8°N 163.8°W	2
19	NP-8	15 July 59	15 July 60	78.1°W 168.0°W	80.4°N 177.8°E	1
20	NP-8	15 July 60	15 July 61	80.4°N 177.8°E	83.4°N 146.2°W	2
21	Alpha	8 June 57	10 June 58	80.9°W 159.5°W	84.1°N 149.7°W	1
22	NP-7	13 Apr 58	11 Apr 59	86.7°N 150.0°W	85.2°N 33.1°W	2
23	ARLIS II	Nov 63	Nov 64	87.8°N 13.5°W	84.6°N 10.5°W	2
24	ARLIS II	12 Nov 64	12 Feb 65	84.6°N 10.5°W	78.5°N 6.5°W	9
25	ARLIS II	12 Feb 64	16 Apr 64	78.5°N 6.5°W	69.3°N 20.0°W	20
26	WH-5	13 June 62	13 Dec 62	83.2°N 69.0°W	83.0°W 65.0°W	0.3
27	WH-5	29 July 63	29 Sept 62	80.4°N 67.9°W	77.5°N 76.5°W	7
28	WH-5	29 Sept 63	29 Nov 63	77.5°N 76.5°W	71.1°N 67.0°W	15
29	WH-5	29 Nov 63	25 Jan 64	71.1°N 67 °W	66.4°N 60.0°W	12

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*Fletcher's Ice Island

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B.4 THE OCEAN

B.4.1 The Arctic Basin

B.4.1.1 Structure

The Arctic Ocean surrounds the North Pole and is bordered by Scandinavia, Siberia, Alaska, Canada, and Greenland. Its area, approximately 4 million sq mi, is about 5 times that of the Mediterranean Sea, but its volume is only 3 times as great. Unlike its southern antipode, the polar region above 80°N is a basin, some of which is over 14,000 ft below sea level, while most of the immense expanse between 55 and 80°N is an area of northward-flowing drainage - the arctic drainage basin. In an oceanographic sense it is a single, large, nearly enclosed basin connected primarily with the Atlantic Ocean through 2 other major arctic seas, the Norwegian Sea and the Greenland Sea. The water characteristics can be traced in large part to North Atlantic water characteristics which have been modified through surface-acting processes associated with the unique climatic conditions of the Arctic. The characteristics of the peripheral continental arctic seas are highly variable and are discussed in greater detail below. The connection of the Arctic Ocean with the Pacific Ocean through Bering Strait is of relatively minor importance oceanographically.

The most recent depiction of the bathymetry is given in Figure IV-1. The continental shelf on the North American side of the Arctic Ocean is narrow (20 to 50 mi), whereas the half bounded by Europe and Siberia is very broad (up to 600 mi) and shallow with peninsulas and islands separating it into five marginal seas - the Barents, Kara, Laptev, East Siberian, and Chukchi. Even though these marginal seas occupy 36% of the area of the Arctic Ocean, they contain

Figure IV-1 ARCTIC OCEAN: BATHYMETRY

only 2% of its volume of water. With the exception of the Mackenzie River in northern Canada and the Yukon River, which enters via Bering Sea and Bering Strait, all the major continental rivers reaching the Arctic Ocean flow into these 5 seas. Thus, these shallow seas, which have a high ratio of exposed surface to total volume and a substantial input of fresh water in summer, greatly influence surface water conditions in the Arctic Ocean.

The major features of the floor of the Arctic Ocean and its seas are shown in figures IV-2 and IV-3. The continental shelf is indented by numerous submarine canyons. Notable canyons include the very large Svyataya Anna and Voronin troughs in the northern Kara Sea and the Hope Valley and Barrow Canyon in the Chukchi and Beaufort seas. The northern edge of the continental shelf, particularly in the Laptev and East Siberian seas, is poorly surveyed; and undoubtedly there are canyons yet to be defined. These canyons have an oceanographic importance as preferential pathways for the egress of water from the relatively warm Atlantic layer onto areas of the continental shelf, where it comes within the influence of strong mixing processes. Hence they can modify locally, but significantly, the surface waters and ice cover.

The Arctic Basin is a true ocean basin. Beneath its deepest parts, the thickness of the earth's crust is about 5 mi, a figure that is fairly typical of the other oceans. The relief of the Arctic Ocean, too, is much like that of the other oceans in that it has two predominant levels: the continental shelves with water depths of a few hundreds of feet and the deep basins with depths of thousands of feet. A sharp break in slope where the continental shelf meets the continental slope marks the boundary between the shelves and basins and defines the edge of the central Arctic Basin. In most oceans, the shelf break is at about 650 ft depth, but in the Arctic it is deeper in some places (1,300 to

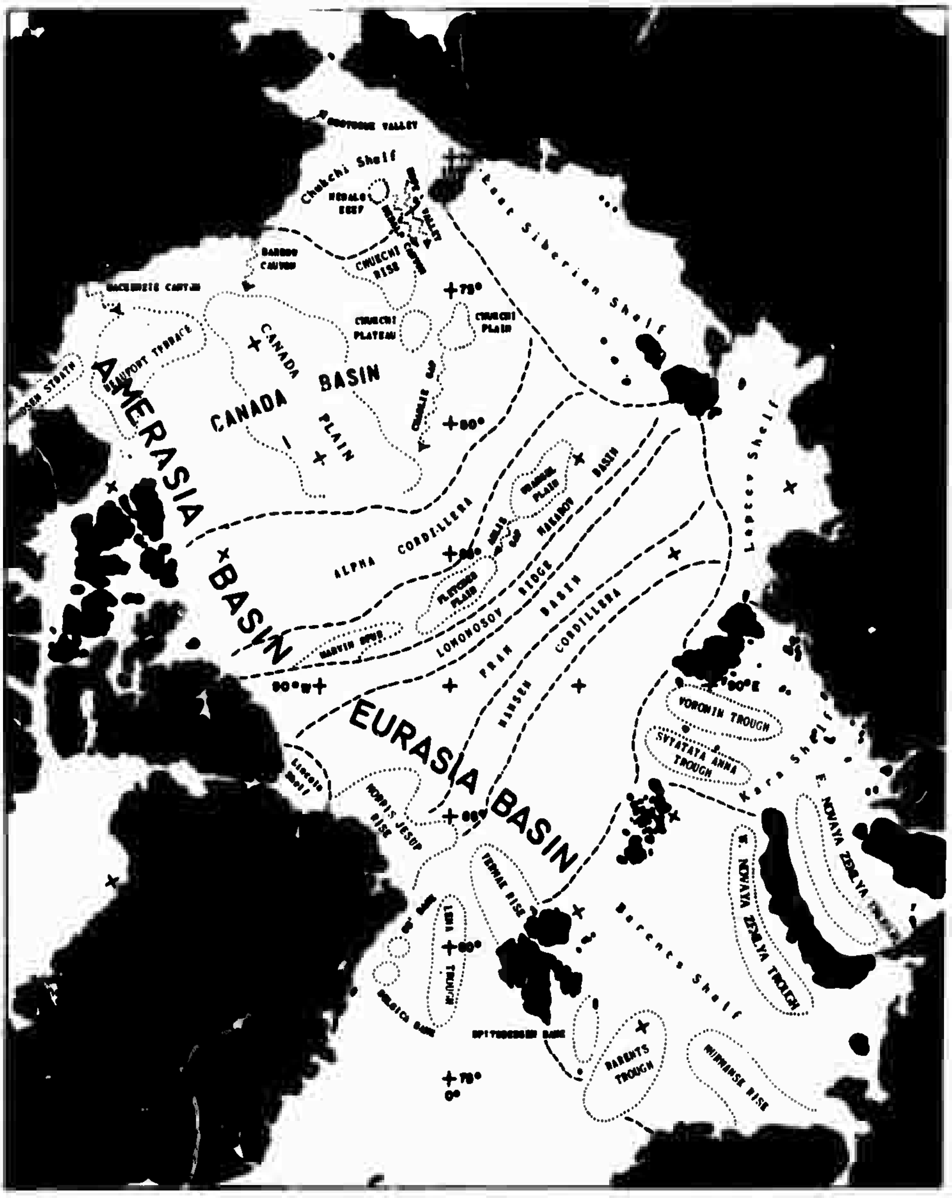

Figure IV-2 MAJOR FEATURES OF THE FLOOR OF THE ARCTIC OCEAN AND ITS SEAS

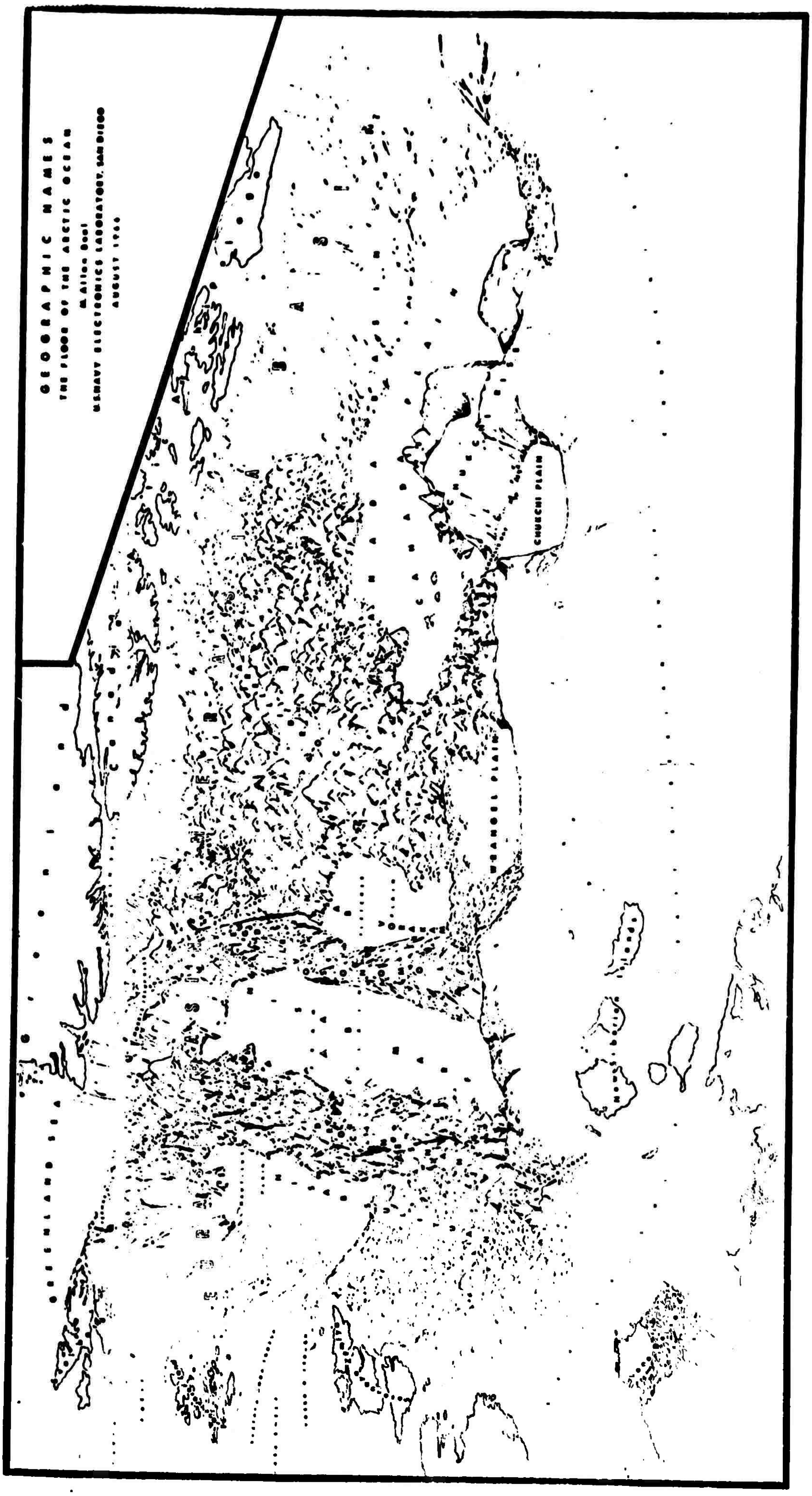

FIGURE IV-3 FLOOR OF THE ARCTIC OCEAN

1,600 ft).

The continental margin on the Eurasian side contains some of the broadest continental shelves in the world. The East Siberian Shelf is 250 to 450 mi wide, and the Barents and Kara shelves are 350 to 600 mi wide. In contrast, the shelves off Greenland and northwestern Canada are only 15 to 35 mi wide. North of Alaska the continental slope between the Novosibirskiye Ostrova and the Chukchi Sea is apparently formed of several level surfaces. The conti[illegible]al shelf along East Greenland, south to 77°N, is broad (150 mi) and contains [illegible] system of banks no deeper than 650 ft. South of 77°N the shelf narrows [illegible]il between 75°N and the Kap Farvel it is less than 50 mi wide. The shelf i[illegible] marked, particularly in the southern part, by several deep indentations. At about 79°N the Greenland Sea is separated from the Arctic Basin by a rise with a sill depth of about 9,200 ft. South of the rise the Greenland Sea contains 2 deep basins. At 71°N the Jan Mayen Ridge extends toward Greenland at a maximum depth of about 5,000 ft. In Denmark Strait, however, the sill depth is about 1,800 ft.

Around Baffin Bay the shelf is relatively narrow and never exceeds a width of 50 mi. The shelf is widest off central Baffin Island, Labrador, and West Greenland south of 78°N. Central Baffin Bay is a basin with depths exceeding 6,500 ft. The sill across Davis Strait appears to be about 2,000 ft deep. Baffin Bay is connected with the Arctic Ocean through a network of channels - Smith, Jones, and Lancaster sounds. The Jones Sound connection is quite restricted but the channels leading to Lancaster and Smith sounds are relatively deep with sill depths extending to 490 and 720 ft respectively.

A gently sloping continental rise usually lies at the foot of the continental slope. North of Alaska, this rise is 25 to 50 mi wide. Much greater widths of 150 to 250 mi occur in the continental rise off the Canadian Arctic Archipelago. Crossing the Canadian continental rise is a system of deepsea channels with relief of 15 ft. A marginal plateau is a level feature which borders the continental shelf at a greater depth and several are known in the Arctic Ocean. The Chukchi Rise is crowned at its outer end by the Chukchi Plateau which has a flat summit with a diameter of about 50 mi at the 1,000-ft isobath. The surface is marked by relief of 15 to 100 ft. Two submarine canyons indent the southwest side of the Plateau. Southeast of the Chukchi Plateau is an area of rough topography which has been described as a continental borderland. Within this area is another plateau, the Northwind Cap. Other marginal plateaus include the Beaufort Terrace, which on its outermost edge is similarly elevated above the saddle which connects to the continental shelf, and the Morris Jesup and the Yermak rises in the Greenland-Svalbard area.

The central Arctic Basin is crossed by three submarine mountain ranges. The ridges and rises are nearly parallel to one another and span the basin from the Eurasian side to the Canadian side. The Lomonosov Ridge, stretching 900 mi between the Novosibirskiye Ostrova and the Greenland-Ellesmere shelf, was discovered by Soviet scientists in 1948. It is a single continuous feature 50 to 100 mi wide. Echograms show a steep-sided ridge with a rather smooth profile. Its minimum depth reported is 3,100 ft, and saddles along the crest have depths of 5,000 to 5,250 ft. A flat surface near the crest of the ridge has been noted on two different crossings. An offshoot of the Lomonosov Ridge is known as the Marvin Spur.

The Alpha Cordillera is about the same length as the Lomonosov Ridge but much broader, its width ranging from 125 to 400 mi. The crest of the

Cordillera is 5,000 to 6,500 ft deep, and the topography is much rougher than that of the Lomonosov Ridge. The magnetic fields over these two features also differ. The field over the Alpha Cordillera is disturbed with numerous anomalies, many of them exceeding 1,000 gammas, while over the Lomonosov Ridge there is little disturbance at all. Neither of these ranges is now seismically active.

The Nansen Cordillera is an extension of the Mid-Atlantic Ridge into the Arctic Ocean. Where the topography has been sampled, it is rough, as it is on the Mid-Atlantic Ridge. The most distinctive characteristic of the Nansen Cordillera is the narrow earthquake belt along its crest. The belt of earthquake epicenters crosses Iceland and then abruptly changes direction just north of the island, where a large east-west fracture zone intersects the mid-oceanic ridge near Jan Mayen. Between northeastern Greenland and northern Siberia the earthquake belt is narrow and straight for a distance of over 1,000 mi, while within Siberia the earthquake zone spreads out and disappears. In the Atlantic a similar earthquake belt coincides with a central rift valley at the crest of the mid-oceanic ridge. Although this relationship might be expected to hold in the Arctic, insufficient evidence has been collected for positive verification.

The abyssal plains are the ultimate repositories for sediments which have been transported across the continental shelves and down the continental slopes via the submarine canyons. In the deep basins the sediments collect to form some of the globe's most extensive level surfaces, with gradients of 1:1,000 or less. Four of the abyssal plains in the Arctic Ocean are arranged in steplike pairs. Each pair is connected by an abyssal gap through which sediments are transported from the upper to the lower plain. The Canada and Chukchi abyssal plains are connected by the Charlie Gap. The complete route of sediment flow is from Herald Canyon to the Chukchi Abyssal Plain and then through Charlie Gap to the Canada Abyssal Plain. The Wrangel and Fletcher abyssal plains are connected through Arlis Gap.

Seismic reflection profiles show that a prominent sub-bottom basement ridge exists in the vicinity of the Arlis Gap. This ridge seems to have acted as a bedrock dam for the sediments of Wrangel Abyssal Plain, ponding them behind it until there was sufficient accumulation to spill over into Fletcher Abyssal Plain. Sediments move from the Siberian Shelf to Wrangel Abyssal Plain and then through Arlis Gap to Fletcher Abyssal Plain. A system of interplain channels funnels the flow across the plain and into the gap. The right (east) bank of these channels is higher than the left bank - a condition that is apparently due to the influence of the earth's rotation.

The most extensive of the plains in the Arctic Ocean is the Canada Abyssal Plain which covers an area of 80,000 sq mi. It is remarkably flat with depths ranging from about 12,400 ft in the north to 12,600 ft in the south. On its northern and western edges it is bounded by the scarps of the Alpha Cordillera and Chukchi Rise. The eastern and southern boundaries grade smoothly into the continental slope. The Pole Abyssal Plain is deeper than the four plains mentioned previously. In the neighborhood of the North Pole it is flat and smooth with a depth of 13,400 ft, but away from the Pole the depth of the plain increases to 15,000 ft.

B.4.1.2 Water Masses

Data from the Arctic Ocean taken over many years and during all seasons of the year show a remarkable regularity in the distribution of temperature and salinity at depth throughout the year as well as a repeating seasonal regularity in the surface waters. This nearly steady state in the observed distribution of temperature and salinity is apparently a result of continuing processes within the Arctic Basin. For many purposes the importance of taking data synoptically is thus reduced, and much may be learned about the geographic distribution of properties by comparing data from different seasons and years.

The most important processes conditioning and modifying the Arctic Ocean water are:

(1) addition of fresh water from the land, primarily from the large Siberian rivers;

(2) addition of fresh water locally through melting of ice;

(3) heat gain through absorption of solar radiation in areas that are not ice-covered;

(4) concentration of salt, and hence increase of density, of surface water as a result of freezing; and

(5) heat loss to the atmosphere at any open water surface, including leads in the pack ice.

The first three process lead to decreased density of the water affected, and they operate only from June to September. Consequently, the surface waters exhibit somewhat lower salt content in summer than in winter, and in ice-free areas (e.g. parts of Baffin Bay) surface temperatures may rise to a few degrees above freezing. In ice-covered areas the temperatures in summer remain near the freezing point because the heat is dissapated by melting the ice rather than warming the water.

The latter two processes lead to increases in density of the surface water. In some areas under certain conditions these modified waters sink to subsurface levels, 325 to 650 ft in depth, but in general in arctic waters these concentration mechanisms are not adequate to create water of sufficient density that it can replace waters lying deeper than 650 ft. Thus all deep arctic waters are advected into and out of the arctic seas from adjacent areas. As the connection with the Pacific Ocean through the Bering Strait is narrow (about 40 miles wide and about 165 ft deep), the deeper waters of the Arctic Ocean and probably Baffin Bay, essentially originate in the North Atlantic.

The similarity in vertical structure at diverse locations throughout the Arctic Ocean (Fig. IV-4) reflects the common North Atlantic origin of water deeper than 650 ft and the similarity of local processes modifying the surface water, principally the arctic climate factors. On the basis of temperature, three water masses may be defined:

(1) The surface layer (arctic water), from the surface down to about 650 ft, has varying characteristics. In ice-covered areas the water temperature is close to the freezing point for the salt content. In the usually ice-free areas (eastern Greenland Sea; along West Greenland north through Davis Strait) temperatures may be a few degrees above freezing as are those in areas that are ice-free in summer (Chukchi Sea; nearshore areas of other peripheral seas; most of Baffin Bay). Temperatures below the surface are typically always cold, except in the Canada Basin of the Arctic Ocean, where there may appear a small temperature maximum (-1.0°C) in the 250- to 325-ft layer. This is attributable to influx of Bering Sea water, which is transported around the Beaufort Sea gyre. The salinity of the surface layer may be uniform down to about 160 ft and then increase closer ot the surface. The surface salinity values also exhibit spatial variation (see below).

(2) The layer below the Arctic water, from about 650 to 3,000 ft, is known as the Atlantic layer. It has temperatures above 0°C, with a maximum at 1,300 to 1,600 ft (up to 3°C). Salinities continué to increase over the surface values until by 1,300 ft and in many instances shallower depths, they attain in

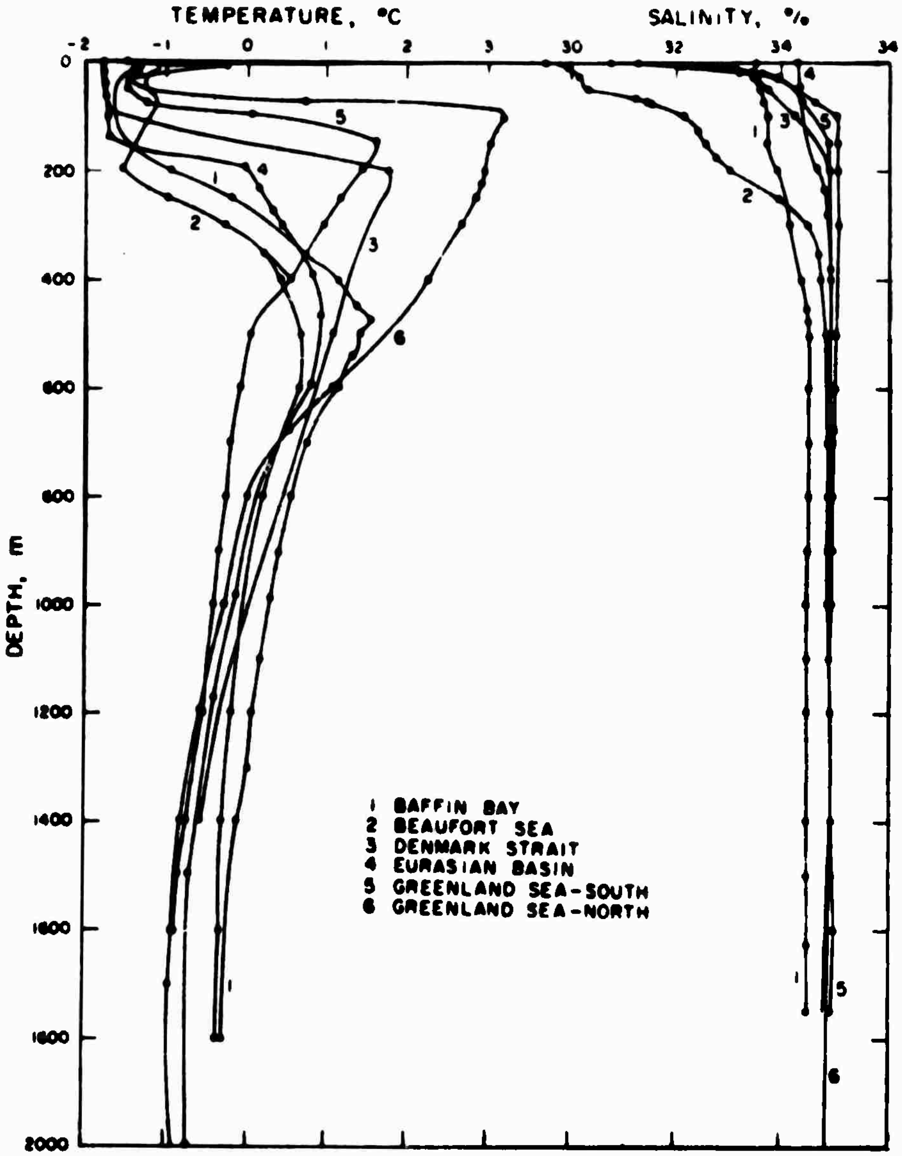

Figure IV-4 VERTICAL DISTRIBUTIONS OF TEMPERATURE AND SALINITY AT SIX LOCALITIES the Greenland Sea and Arotic Ocean nearly uniform values in the range of 34.9 to 35.1°/oo.

(3) Beneath the Atlantic layer is bottom water with temperatures below 0°C and the same uniform salinities attained in the Atlantic layer (34.93 to 34.99°/oo). Deep temperatures vary slightly from basin to basin; in the Canada Basin they are about -0.30 to -o.40°C, in the Eurasian Basin -0.70 to -0.80°C, in the Greenland Sea -1.2°C, and in Baffin Bay -0.45°C.

The density of cold water is incluenced much more by salinity than by temperature - thus the vertical distribution of density closely parallels that of salinity. On the basis of density, arctic waters show a two-layer system, with a thin, less dense surface layer separated from the main body of water of quite uniform density by a strong pycnocline. The pycnocline restricts vertical motion and the vertical transfer of heat and salt, and hence the surface layer acts as a lid over the large masses of warmer water below.

There is little spatial variation of water surface temperature throughout the Arctic. Only those areas that are normally ice-free all year exhibit temperatures significantly above freezing. These areas are influenced by currents carrying warmer water into the Arctic and remain ice-free for that reason. Seasonal temperature fluctuations occur in areas that are typically ice-free seasonally (July to September) including the coastal sectors of the peripheral seas of the Arctic Ocean and around eastern and northern Baffin Bay. Areas in which major currents carry arctic water the North Atlantic remain perennially ice-covered and have temperatures close to the freezing point at all times.

In [illegible] closed basins in higher latitudes (e.g. Arctic Ocean, Baffin Bay) there is a surface outflow of relatively fresh water with some saline water entrained with it and an inflow of more saline water at depth. In such basins there is an increase in surface salinities from the head of the basin, close to the source of fresh water, toward the mouth. This can be attributed to progressive mixing between the surface layer and the underlying more saline water.

The general conditions for basins in high latitudes are present in the Arctic Ocean and Baffin Bay. The surface salinity in summer is lowest along the periphery of the Arctic Ocean from the Mackenzie River Delta to the Kara Sea, because major rivers discharge throughout that area. Salinities are never much below about 27°/oo, except very close to a river mouth, because the fresh water is efficiently mixed with more saline water on the shallow continental shelf by the action of winds, tides, and currents. Salinities in general increase progressively in the direction of flow of the surface water and attain maximum values in those areas where currents are directly introducing North Atlantic surface water into the Arctic Basin. The North Atlantic surface water is heavier and sinks within the central basins to subsurface levels. Though no large rivers discharge into Baffin Bay, the same general picture of the surface salinity holds. The flux of Arctic Ocean water through the sounds has lower salinity and hence is less dense and contributes primarily to the surface layer. Intensive melting of ice in summer contributes significant amounts of fresh water locally, and hence relatively low salinities (31°/oo) may be observed in central Baffin Bay and along Baffin Island associated with the central pack and the outflowing Canadian Current.

Surface salinities in winter can be higher by 0.5 to 1°/oo in the Arctic Ocean and by 1 to 2°/oo in the peripheral seas, as at this season melt and runoff are not contributing fresh water but mixing with more saline waters is continuous. However, the general pattern of distribution remains similar. In winter, (Octo-

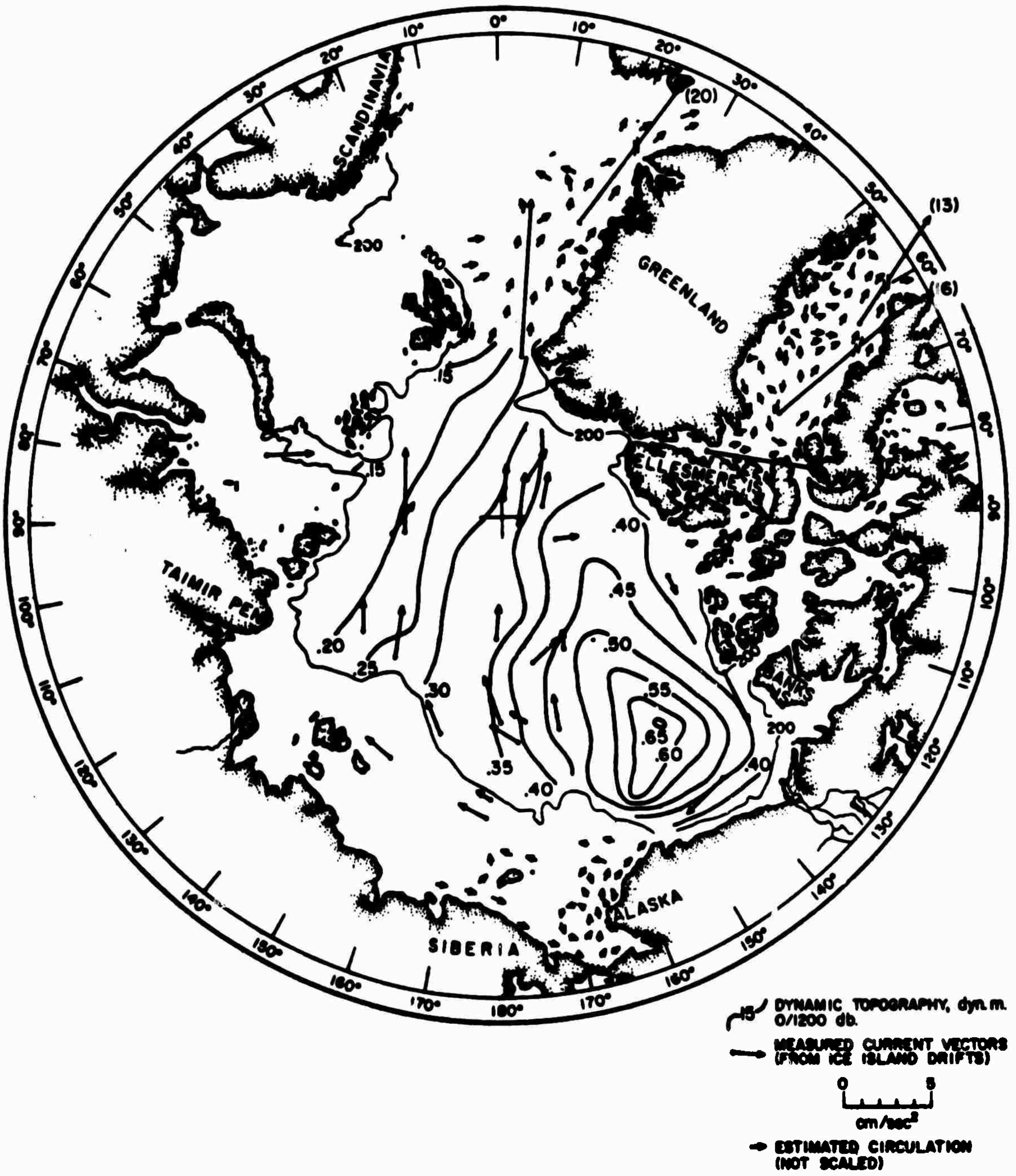

Figure IV-5 COMPOSITE SURFACE CIRCULATION (FROM DYNAMIC TOPOGRAPHY, STATION DRIFTS, AND ESTIMATED FROM TEMPERATURE AND SALINITY DISTRIBUTION. STATION DRIFT VECTORS ARE SCALED TO SPEED SCALE, EXCEPT THREE WHICH ARE LABELLED.)

ber to April), a process of considerable importance in modifying the surface waters takes place. When ice grows from seawater, the salt is largely excluded from the ice and results in a local increase in salinity and hence density of the remaining water. However, the ice cover is not continuous and leads are continually opening and closing. Heat loss to the atmosphere from the open water of leads occurs perhaps 100 times faster than through ice; thus freezing is more rapid and considerable salt will be concentrated in the waters of leads, where also temperatures slightly below the freezing point may be observed.

Probably some sort of cellular convection is induced, with downward-moving columns under the leads carrying excess salt and supercooled water. This effect may reach as deep as 325 ft, the excess salt contributing to the maintenance of the strong salinity (density) gradient present in the lower part of the surface layer, while upward motion elsewhere slowly transports the heat of the Atlantic layer toward the surface.

B.4.1.3 Currents

The surface circulation has been largely derived from the observed drift of various manned ice islands, floe stations, and ships (Fig. IV-5). Current measurements show that on the average the drifting stations and the surface water tend to move in similar directions at similar speeds, though there may be considerable variations between them over short periods. The mean pattern is depicted in Figure IV-5 in which the velocity vectors of the various drifting stations are plotted together with the mean dynamic topography calculated from over 300 oceanographic stations.

The circulation of the arctic water is in part created by density differences and in part is wind-induced, which is confirmed by theoretical studies. The net effect of tides is unknown, although there may be some asymmetry in their action which would modify the circulations. The surface waters from the whole Eurasian side of the Arctic Ocean tend to move toward the North Pole. This flow is on the average slow, perhaps 0.05 to 0.1 knot; but after passing the region of the pole, the flow becomes more concentrated and then exits from the basin as the East Greenland Current. In the Beaufort Sea the surface waters have a clockwise movement, apparently a result of the general wind pattern, such that they tend to flow to the southwest along the shelf off the Canadian Arctic Archipelago and to the north in the area north of Bering Strait.

Around the Greenland Sea there is a large cyclonic circulation, with average speeds in the range of 0.2 to 0.4 knot. Inflow of North Atlantic water, both at the surface and at deep levels, occurs along the east side of the sea as the West Spitsbergen Current. The East Greenland Current is the major flow south on the west side; surface water from the Arctic Ocean contributes to the upper layers, while the deeper waters are largely from the West Spitsbergen Current, completing the cyclonic gyre. The current follows closely the continental slope, and over the wide continental shelf of the northern area (77 to 80°N) the currents tend to be weak and variable. The East Greenland Current seems to accelerate toward the south, attaining speeds of 0.3 to 0.8 knot near Denmark Strait.

The same general pattern of circulation is found in Baffin Bay. A cyclonic circulation dominates the bay; inflow of North Atlantic water occurs off western Greenland through Davis Strait, and inflow from the Arctic Ocean through Smith, Jones, and Lancaster sounds. The Canadian Current runs south along Baffin Island, and as it accumulates water from the various inflows it generally shows higher speeds toward the south.

Direct current measurements from Fletcher's Ice Island, T-3, in 1965 disclosed an unexpected phenomenon - a core of high-speed current (approximately 1 knot) located in the major pycnocline separating the Arctic water from that of the Atlantic layer at depths of 250 to 650 ft. This swift motion occurs only occasionally and has been observed to grow and then decay within a few days. Presumably this motion is associated with certain atmospheric phenomena and is at present the subject of more intense study.

The circulation of the Atlantic layer has been deduced from the distributions of temperature and salinity. Recent direct current measurements of this layer from U.S.S.R. and U.S. drifting stations confirm the general pattern of motion (Fig. IV-6). On entering the Eurasia Basin from the Greenland Sea much of the water flows east along the edge of the Eurasian continental slope. The water enters the Amerasia Basin on a broad front across the Lomonosov Ridge. There appears to be a general cyclonic circulation in the Eurasia Basin and a smaller anticyclonic gyre in the Beaufort Sea. Variations between the mean flow pattern and the direct measurements are apparent, but observations are as yet so sparse that it is not known whether the deduced mean pattern is in error or whether periodic or aperiodic effects are significant in the motion.

Direct current measurements in the Atlantic layer of the East Greenland Current show there is little variation in velocity with depth. Thus the deep motion is similar to that of the surface water. Only a few current measurements have been made in the Atlantic layer of Baffin Bay and the data are still being analyzed. Deductions from the mass and temperature fields indicate a circulation similar to that of the Greenland Sea (without the loss of water into the Arctic Ocean). The speeds presumably would be of the same order as those of the surface currents. Likewise, practically no current measurements in the bottom

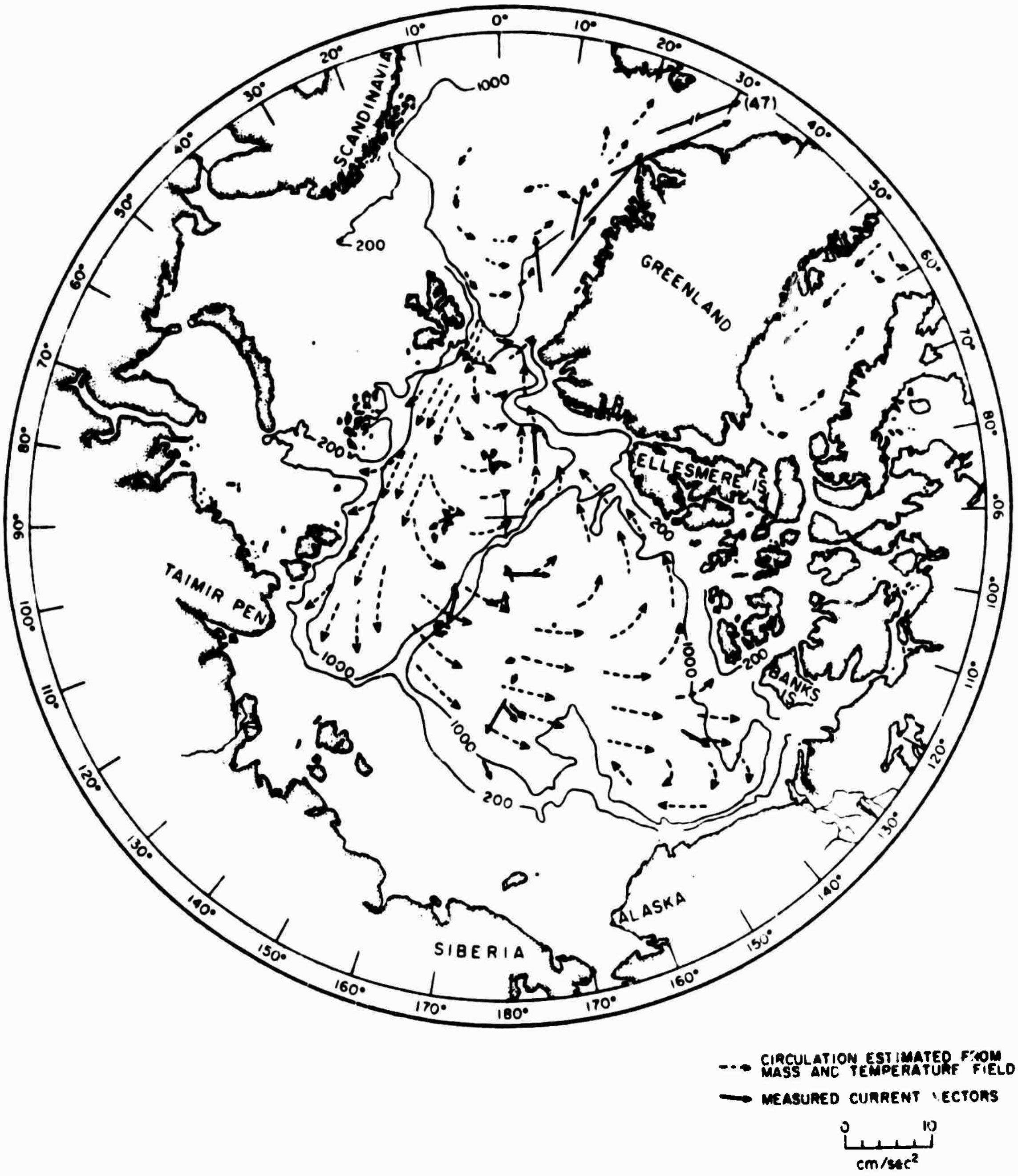

Figure IV-6 COMPOSITE ATLANTIC LAYER CIRCULATION (FROM DIRECT CURRENT MEASUREMENTS AND ESTIMATED FROM TEMPERATURE AND SALINITY DISTRIBUTION) currents in over 10,000 ft of water from T-3 gave results similar in both speed and direction to that of the Atlantic layer above. Thus, though the circulation of the bottom water is essentially unknown, it appears that it may be similar to that of the Atlantic layer. The same situation may exist in the Greenland Sea and Baffin Bay as well.

B.4.1.4 Advection Boundaries

The major advection boundaries are the Bering Strait, the Greenland-Spitsbergen strait, and the straits through the Canadian Arctic Archipelago. Oceanographically, the best known of these is the Bering Strait, through which, in summer, surface Bering Sea water flows north into the Arctic Ocean. The volume transport is about 2.7×10^7 ft^3 /sec. In the east channel of the strait, speeds normally range between 1 and 2 knots, although speeds over 3 knots have been measured. In the west channel, speeds are lower by a factor of 2 to 3. The situation in winter is still essentially unknown, though it has been suggested that the northward flow may be only one-fourth that of summer or may even reverse on occasion. Year-round monitoring of the flow in the strait has recently been started with a nuclear-powered current meter and telemetry system in the eastern channel near Fairway Rock.

The general flow through the passages of the Canadian archipelago is from the Arctic Ocean toward Baffin Bay. Direct current measurements are essentially nonexistent. Documentation of the drift of ice island WH-5 through Nares Strait confirms the general flow out of the Arctic Ocean through this strait. However, there was evidence of large pulsations in the southerly flow; the indicated periodicity of a few days to weeks suggest that major atmospheric disturbances may be important in significantly altering the flows through these channels.

The strait between Greenland and Spitsbergen provides the primary connection between the waters of the North Atlantic and the Arctic Ocean; water flows into the Arctic Ocean on the eastern side of the strait and out of the Arctic Ocean on the western side as the East Greenland Current. A new concept of the circulation in the Greenland Sea and of the East Greenland Current has resulted from an analysis of current measurements made from ARLIS II during its drift along the East Greenland during the winter of 1965. The measurements showed the volume transport of the current to be 1 x 10^9 ft^3 /sec, an order of magnitude larger than previously estimated. This large transport is apparently a major circulation internal to the Greenland-Norwegian seas; the outflow and inflow from the Arctic Ocean represent only minor contributions and subtractions from a large cyclonic circulation. The Arctic water portion of the current to a large extent controls the ice distribution, and so its presence is manifested out of all proportion to its small contribution to the total transport.

B.4.1.5 Peripheral Seas

Of the seven seas which constitute a portion of the Arctic Ocean six are worthy of brief mention. The seventh, Beaufort Sea, is physically and oceanographically a part of the Canada Basin. The suggestion has even been made that the term be dropped but custom and habit will probably secure its place in common usage.

The Barents Sea occupies nearly a half-million square miles on the continental shelf of Eurasia. It has free contact with the Norwegian Sea on the west and with the Arctic Ocean on the north; and it is deeper than the other peripheral seas, much of it more than 650 ft deep. The bottom is more like a continental borderland than a shelf, as it has both exceptionally shallow and deep areas scattered through it. The flat shelf areas are east and southeast of Svalbard and in the southeastern part of the sea. An east-west ridge at a depth of 650 ft connects the shore areas around Zemlya Frantsa Iosifa with Svalbard, and a north-south ridge at 1,000-ft depth separates the western Björnöya basin, with depths of over 1,300 ft, from the eastern basin. The eastern depression extends southwest between Zemlya Frantsa Iosifa and Novaya Zemlya with general depths over 1,000 ft and occasional depths over 1,300 ft. In the extreme north there is a third depression between Svalbard and Zemlya Frantsa Iosifa.

The White Sea is a large extension of the Barents Sea that projects more than 250 mi into the European mainland. The approach from the Barents Sea is appropriately called Voronka (funnel), and is about 60 mi wide. The entrance narrows to the southwest until between the Polvostrov Kol'skiy and the mainland it becomes a strait, called Gorlo (throat), 25 mi wide, where ice obstructs navigation even when the western, inland, and wider part of the White Sea is reasonably ice-free. The White Sea is for the most part less than 330 ft deep, but the floor of Kandalakshskaya Giba in the northwest descends to more than 1,000 ft below sea level.

The surface characteristics of both climate and sea ice in the Barents Sea result primarily from an influx of warm water from the Norwegian Sea. This warm current flows north along the coast of Norway; a southern branch enters the Barents Sea along the north coast of Norway and Kol'skiy Polvostrov as the North Cape Current; the northern branch flows north of Björnöya and then turns northwest, passing along the south and west coasts of Svalbard as the West Spitsbergen Current. Off Varanger Halvöya the North Cape Current flows at 0.3 knot but the velocity decreases to the east. The current splits at Varanger Fjord - one part flows in a belt 50 to 60 mi wide to the entrance of the White Sea, the other curves northeastward across the Barents Sea and passes north of Novaya Zemlya into the Kara Sea. This branch is weak in the north and south of its course, but between 36°E and 44°E it runs at approximately 0.75 knot.

The major inflow of cold water into the Barents Sea is between Novaya Zemlya and Zemlya Frantsa Iosifa. This current also branches into two parts - one part flows southwest of the archipelago and the other west as the Bear Island Current. In the southeast the general movement of water is toward the Kara Sea except for the Litke Current which has the reverse direction. It moves west through the northern half of Proliv Karskiye Vorota and then northwest along Novaya Zemlya joining the general northerly movement there. In the White Sea there is a weak outward current in spring and summer and an equally weak counterclockwise eddy within the basin.

The Kara Sea is exceptionally shallow in the east but relatively deep for its position on the continental shelf in the west. Off the east coast of Novaya Zemlya there is a 40-mi wide basin that in some places is 1,750 ft deep. A ridge in the north of the Kara Sea separates the basin from the Arctic Ocean. Between Polvostrov Yamal and Novaya Zemlya there is another trough of deeper water. There are two more basins in the north of the sea which have maximum depths of about 1,700 ft adn are separated by a ridge. Ostrova Uyedineniya, Vize, and Ushakova form the highest elevations of this ridge. In the southeast of the Kara Sea depths average only 165 ft 50 to 120 mi from the shore. The water adjacent to the Ob' and the Yenisey rivers is also exceptionally shallow.

The only important current in the Kara Sea forms a closed counterclockwise circulation in the western portion. The gyre begins in the east with Ob' and Yenisey waters which spread as they leave the estuaries. One branch flows to Novaya Zemlya where it turns southwest to Proliv Karskiye Vorota. Within the main circulation are two small weak counterclockwise eddies. Water also enters the Kara Sea around the north of Novaya Zemlya from the Barents Sea and eventually mixes with the Ob-Yenisey waters.

Tides in the Kara Sea are semidiurnal and relatively weak. They come from the Barents Sea along eastern Novaya Zemlya and from teh Arctic Ocean along western Severnaya Zemlya. They merge at Ostrov Uyedineniya and progress southwest. The average amplitude is 1.5 to 3 ft but winds commonly increase the tidal range by 3 ft or more.

Both the Laptev and the East Siberian seas are shallow basins with gentle shores. The edge of the continental shelf is up to 400 mi offshore Only in the northeastern Laptev Sea, off Severnaya Zemlya, are depths greater than 300 ft. The western sector of the East Siberian Sea, south of the Novosibirskiye Ostrova and east to the Kolyma River, is exceptionally shallow with many shoals. Between the Indigirka and the Kolyma rivers an almost continuous shorebank, defined by the three-fathom curve, extends about 23 mi out from the shore. From the Kolyma east to Mys Shmidta, the coastal water is deeper. There are only a few islands, and these, with the exception of Medvezh'i Ostrova, are close to the shore. The sea deepens slowly to the northeast; maximum depths are 150 to 180 ft.

The general flow of water in both the East Siberian and Laptev seas is counterclockwise. There is a weak easterly coastal current which is modified by water from the large rivers which forces it offshore in a northeasterly direction at 1 knot; counterclockwise eddies develop when it is caught in coastal indentations. The major current entering the Laptev Sea comes through Proliv Vil'kitskogo between Mys Chelyuskin and Severnaya Zemlya. It is joined by a cold current flowing southeastward along Severnya Zemlya at 0.2 knot, and the combined waters move along the Taimyr coast into the shallow part of the Laptev Sea. At the Lena Delta the current splits; and one part, flowing along the west side of the Ostrova Novosibiriskiye at 0.5 to 1 knot, sets to the north of the archipelago and joins the main arctic drift. The other part flows through Proliv Dmitriya Lapteva and other straits into the East Siberian Sea.

The waters that pass through the straits separating Novosibirskiye Ostrova and the mainland spread out on reaching the East Siberian Sea. The main branch near the coast flows at approximately 0.3 knot. A branch of this current is believed to pass north and west of Ostrov Vrangelya. In summer a current appears to reach through Bering Strait and flows northwest to the middle of Proliv Longa; its direction may be reversed in the winter.

North of the coastal currents in both the Laptev and the East Siberian seas, the water flows in large counterclockwise eddies. Farther north still is a west-northwest current which runs northwest at the Ostrova De-Longa and passes north of the Ostrova Novosibirskiye into the Laptev Sea. It continues northwest across the northern margin of the sea and flows north of Severnaya Zemlya.

Tidal progression is southward from the Arctic Ocean in both seas. Tides are semidiurnal and their range is 1 ft, although it may be raised to 10 to 12 ft with an onshore wind. Tidal currents flow from 0.5 to 0.8 knot in the Laptev Sea but are weaker in the East Siberian Sea.

The Chukchi Sea is nearly 600 mi across, north from Bering Strait toward the Pole. Its floor is probably the flattest plain in the world and over extensive areas slopes are less than 12 in/mi. The greater part of the sea within 120 mi of the coast is from 50 to 100 ft deep. The overall mean depth is about 165 ft. Low swells rise from the plain, the best known being the Herald Shoal that comes to within 50 ft of the surface at about 70°30'N midway between Wrangel Island and Alaska. The bed of the sea is probably corssed by a network of valleys formed subaerially when the floor was dry land during the Pleistocene. One of the large landforms of this type is the Hope Submarine Valley which leads west from Cape Thompson and has relief of about 30 ft. Poorly formed terraces have been recognized below sea level and the remains of a former split can be traced 25 mi north from Cape Prince of Wales.

The continental slope commences at about 650 ft and is crossed by several submarine valleys. Chukchi Cape, 125 mi across, and the smaller Northwind Seahigh are outliers of the continental shelf beyond the slope.

Water enters the Chukchi Sea through Bering Strait. This current appears as warm and low-salinity water that converges on Point Hope and there turns east toward Point Barrow. The waters on the Siberian coast (in the vicinity of Bering Strait) are cold and of high salinity. It was thought that this water passes south through Bering Strait, but it now appears that it is north-flowing and its properties derive from divergence and upwelling. A cold current passes east through Proliv Longa. There is consequently a tendency for a clockwise circulation in the Chukchi Sea.

B.4.2 Other Arctic Seas

B.4.2.1 Norwegian and Greenland Seas

The Norwegian and Greenland seas are bounded on the west by the Greenland continental shelf and on the east by the Norwegian, Barents, and Svalbard continental shelves. The depths of the edges of the shelves vary greatly. A large area of the shelves is deeper than 600 ft and frequently extends to 1,320 ft. Seaward of the shelves are the continental slopes with gradients of 1:15 to 1:40 leading down to the abyssal plains.

The Norwegian shelf is 50 to 140 mi wide in its southern and central parts but narrows to 10 mi off the Lofoten Islands. It is more irregular than normal continental shelves and is crossed by a number of troughlike gullies. These are associated with the major fiord systems of Norway, and their present form is mainly the result of erosion by the Quaternary ice sheets. The glaciation also produced morainic ridges which are found out to the shelf edge. The Svalbard shelf is 10 to 40 mi wide. It is traversed by deep glacial canyons originating in the fiords of Spitsbergen and also contains elongated depressions parallel to the coast. This shelf extends to the banks south of Björnöya, while eastward to the banks off northern Norway is the shelf area of the western Barents Sea. The latter shelf is deeper and less irregular than the shelves off Norway and Svalbard.

The Greenland continental shelf north of the Denmark Strait is generally broad (60 to 170 mi) but narrows to about 15 mi in the extreme north. It is widest at the Belgica Bank. Typically the bottom is irregular and rough, but local smooth areas are common and there are a number of shoal areas less than 330 ft in depth. Troughs are present in the shelf and are both parallel and normal to the shoreline. South from Denmark Strait the Greenland Shelf narrows from 170 mi in width to 25 mi at Kap Farvel. It is crossed by prolongations of the fiords of East Greenland. A rise of 165 to 330 ft - the terminal moraine of the Quaternary ice sheet - is found all the way along the edge of the shelf, and it closes these channels against the sea.

Around Iceland the continental shelf is up to 80 mi wide in the west and north and 50 mi in the east, but off the south coast it narrows to 10 mi. In the north and west it is crossed by gullies that start in the fiords and inlets of the Icelandic coastline; in the southeast pronounced gullies seem to have no such link.

The dominant feature of the central parts of the Norwegian and Greenland seas is the mid-oceanic mountain range forming the Iceland-Jan Mayen and Mohn ridges. This is the continuation of the Mid-Atlantic Ridge which rises above sea level in Iceland. The Mid-Atlantic Ridge is characterized in its most elevated region by a central rift zone and by the fact that earthquake epicenters tend to lie on or very near to its axis. These features continue in the Iceland and Jan Mayen ridges. A narrow, deep rift valley, 2 to 3 mi wide at the 10,800-ft isobath, extends along their axial lines and has well-developed structural benches on its walls. It ends at 78°30'N where it meets the Spitsbergen block. The base of the ridges extends about 35 mi from the rift, and their crests lie at depths less than 7,200 ft. On both flanks of the ridges are many seamounts. The mountains of the ridge systems rise above the sea as the island of Jan Mayen. A massive, almost unbroken feature, the south Jan Mayen Ridge, runs south from the island; its upper surface is levelled with least depths between 2,640 and 3,600 ft.

The Greenland Sea Basin was formerly considered to be separated from the Arctic Basin by the Nansen Sill, but recent Soviet expeditions have shown that a continuous sill does not exist because it is cut through the middle at about 1°E longitude by the deep Lena Trough, the minimum depth of which is 10,200 to 11,200 ft. The mid-oceanic ridge turns in a northwesterly direction from a region to the west of Prins Karls Forland and runs in what is known as the Spitsbergen Fracture Zone to the northeastern tip of Greenland. This results in a bottom topography with a complicated ridge and trench structure. To the east of this zone is the gently undulating Yermak Plateau which extends for 130 mi from the northwestern corner of Spitsbergen with crests at depths shallower than

3,000 ft.

The basin of the Norwegian Sea is separated from the abyssal plains of the northernmost Atlantic, which has depths as great as 10,000 ft, by the Scotland-Greenland Ridge. The eastern part, between Scotland and the Faroes, is known as the Wyville Thomson Ridge and most of it is less than 2,000 ft in depth; but between Faroe Bank and the Faroe Islands is a narrow channel through the ridge with a sill depth of about 2,700 ft and a least width, at the 650-ft isobath, of about 13 mi. The Faroes-Iceland Ridge lies mainly at a depth of 1,300 to 1,650 ft, but the central part is shallower than 1,300 ft and has several peaks. It is crossed by three saddlelike depressions with depths of 1,375 to 1,600 ft. The Iceland-Greenland Ridge lies at depths between 1,000 and 1,300 ft, but it is cut from northeast to southwest by a narrow channel which has a sill depth of 2,000 ft at about 26°30'W.

The deeper parts of the Norwegian Sea can be divided into three regions. The Iceland Basin lies between the Iceland-Jan Mayen and South Jan Mayen ridges and has depths of 6,500 to 8,200 ft. The Norwegian Abyssal Plain, to the east of the South Jan Mayen Ridge, is the most extensive of the three regions and has a considerable area deeper than 11,500 ft. A number of tall, isolated steep-flanked seamounts rise from this plain. To the east it is bounded by the Norwegian Plateau, a flat-topped feature generally shallower than 6,500 ft. An Extension of this plateau to the northwest separates the Norwegian Abyssal Plain from the Lofoten Abyssal Plain. This third region is relatively extensive, and it has a large area greater than 10,500 ft in depth. It is bounded on the north by the Mohn Ridge.

Beyond the Mohn Ridge is the basin of the Greenland Sea. It consists in the south of the Greenland Abyssal Plain, at a depth of 12,000 to 12,500 ft, with the deepest parts toward the southeast and east. This is separated from the smaller Boreas Abyssal Plain by the Greenland Fracture Zone, which has a crest shallower than 9,000 ft and many individual peaks shallower than 6,000 ft and is a precipitous feature with gradients up to 1:10 on its steeper southern side. To its north is the Hovgaard Fracture Zone parallel to and slightly smaller than the Greenland Fracture Zone. The Hovgaard Fracture Zone separates the Boreas Abyssal Plain from a small basin which is probably not a true abyssal plain. This has a smaller irregular relief smoothed by a blanket of sediment and lies immediately to the south of the Spitsbergen Fracture Zone.

The water masses of the area together with their temperature and salinity characteristics are listed in Table 4-I.

TABLE 4-I

Temperature and Salinity Characteristics of the Water Masses of the European and American Arctic and Subarctic Seas

Northeast Atlantic water	9.5°C; 35.35°/oo
Irminger-Atlantic water	4° to 6°C; 34.95 to 35.10°/oo
Arctic water	<0°C; <34.0°/oo
Labrador Sea water	3.4°C potential temperature; 34.89°/oo
Northeast Atlantic Deep water	3.0°C potential temperature; 34.95°/oo
Northwest Atlantic Bottom water	0.8° to 1.5°C; 34.91°/oo
Norwegian Sea Deep water	<0°C; 34.92°/oo
Arctic-Atlantic water	0° to 2°C; 34.8 to 35.0°/oo
North Icelandic Winter water	2° to 3°C; 34.85 to 34.9°/oo

The two dominant features are the warm currents which form the ends of the Gulf Stream system and the cold currents that derive from the Arctic Basin. Taking the warm currents first, the North Atlantic Drift divides at the Mid-Atlantic Ridge at about 51°N latitude. One arm moves northward parallel to the ridge toward Iceland where it becomes the Irminger Current. This latter carries the Irminger-Atlantic water and bifurcates to the west of Iceland, one branch proceeding northward and then turning east along the north coast of Iceland and the other turning first west and then south to flow along the East Greenland slope eventually to round Kap Farvel and flow northward along the West Greenland slope in the Davis Strait.

A second arm of the North Atlantic Drift carries warm northeast Atlantic water eastward toward the British Isles and passes through the Faroes-Shetlands Channel and to the west of the Faroe Islands into the Norwegian Sea where it is known as the Atlantic or Norwegian Current. This current moves northward along the Norwegian coast and gradually becomes cooler and less saline. Off northwest Norway it divides into the North Cape Current, which flows eastward into the Barents Sea, and the West Spitsbergen Current, which continues northward past Spitsbergen to enter the Arctic Basin. Part of the West Spitsbergen Current turns westward off the northern part of Spitsbergen and flows toward the Greenland Shelf. It then turns south and proceeds below the East Greenland Current as a warmer more saline intermediate layer at a 650- to 1,300-ft depth.

The main cold current of the region is the East Greenland Current. This carries ice and Arctic water, with subzero temperatures and low salinity, from the Arctic Basin along the whole length of the East Greenland shelf to round Kap Farvel and enter Davis Strait. There are two branches of the East Greenland Current. The first, the Jan Mayen Current, flows eastward to the north of Jan

Mayen in the region of the Mohn Ridge. The second, the East Icelandic Current, flows southeastward past the northeast coast of Iceland sometimes reaching the north coast of the Faroe Islands or beyond. Lesser sources of Arctic water and ice are the Björnöya and East Spitsbergen currents. These originate in the northeastern part of the Barents Sea and move southwestward over the Svalbard Shelf, the former to reach Björnöya and the latter to round Sorkapp on Spitsbetgen and flow northward between the Spitsbergen coast and the West Spitsbergen Current. The distribution of the sea ice is determined by the East Greenland Current and its two branches and to a lesser extent by the East Spitsbergen Current.

Between the East Greenland Current and its branches, on the one hand, and the Atlantic Current, on the other, an effective mixing takes place and wide areas in the Norwegian and Greenland seas are covered by mixed water. Similar mixing - though on a smaller scale - between warm and cold currents takes place on the Svalbard and Barents shelves and along the East Greenland slopes south of Denmark Strait. Where the warm and cold currents meet, extremely sharp temperature gradients can occur in the near-surface layers. Coastal water of varied composition and temperature occurs along the coasts of Iceland and Norway. These coastal waters mix along the edges of the Norwegian and Icelandic shelves with the warm currents that flow outside and beneath them.

The speeds of the various currents are not well-established. There is evidence to suggest that there are frequent changes becuase of the wind, but the direct measurement of the Eulerian form of the motion over long periods of time by means of moored recording current meters has only just begun. Some estimates of the Lagrangian form of motion have been obtained from the sets of ships and the drift of floating objects. These put the speeds of the Atlantic Current and East Greenland Current at 0.5 to 1 knot. Locally the East Greenland

Current can reach very high speeds; for example, it is 3 knots just south of the Denmark Strait. The basins of the Norwegian and Greenland seas contain a very nearly uniform deep water with a salinity of about 34.92°/oo and a temperature of about -1°C. The mixed water in the upper layers, primarily in the Greenland Sea and to a smaller extent in the Norwegian Sea, is cooled in winter; but before it can freeze it reaches a higher density than that of water below it and so sinks to form the deep water. In the region of the Iceland-Jan Mayen and Mohn ridges Arctic Intermediate water appears above the Norwegian Sea Deep water and below the Arctic water of the East Greenland Current System. In the Icelandic coastal area, vertical mixing of Atlantic Water and Arctic water in winter results in a homogeneous water in the uppermost 650 to 1,000 ft. This water has a temperature of 2 to 3°C and a salinity of 34.85 to 34.90°/oo and is called North Icelandic Winter water.

The Norwegian Sea Deep water flows into the North Atlantic Basin through the channel between Faroe Bank and the Faroe Islands. It flows at the bottom of this channel at a speed in excess of 2 knots, and as it does so the northeast Atlantic water which lies above it is entrained into the flow. The resultant mixing produces the northeast Atlantic Deep water. This water mass has minor constituents because at times the Arctic-Atlantic water flows through the same channel. Also, the Norwegian Sea Deep water, Arctic-Atlantic water, and North Icelandic Winter water - particularly near Iceland - overflow the Faroes-Iceland Ridge and proceed down its southern flanks, entraining overlying northeast Atlantic water as they do so, to eventually join the outflow from the Faroe Bank Channel as it flows westward at the foot of the ridge.

The northeast Atlantic Deep water turns south when it meets the Reykjanes Ridge; at about latitude 53°N it breaks through this ridge and flows northward along its western flanks to fill most of the Irminger Sea at a depth of 5,700 to 8,200 ft. Above it, in the Irminger Sea, is Labrador Sea water with its core at 1,700 to 4,300 ft. This water is formed as the result of the vertical mixing from the surface down to 4,600 ft of low salinity water in the Labrador Sea in winter. Below the northeast Atlantic Deep water in the Irminger Sea is the northwest Atlantic Bottom water which originates with the overflow of water from the Norwegian Sea across the Iceland-Greenland Ridge in the region of the narrow deep channel in the Denmark Strait. The overflowing water is at times Norwegian Sea Deep water and at times Arctic-Atlantic water. In both cases the overflow proceeds at high speed down the East Greenland continental slope and entrains first overlying Irminger Atlantic water and later Labrador Sea water and northeast Atlantic Deep water to produce a water mass of high density which fills the bottom parts of the basins of the Irminger and Labrador seas. Thus, in addition to a counterclockwise horizontal circulation in the upper part of the water column with the northeast Atlantic water entering this area and the East Greenland Current leaving it, there is also a circulation in the vertical plane with the inflow of the northeast Atlantic water being compensated for by deeper outflows over the Scotland-Greenland Ridge of Norwegian Sea Deep water and Arctic-Atlantic water.

B.4.2.2 Baffin Bay

The central parts of Baffin Bay and Davis Strait are occupied by the Baffin Basin that is separated from the deep waters of the Labrador Sea by a broad sill between Cumberland Peninsula and Greenland north of Gadhaab. The continental shelf on both sides of Baffin Bay is crossed by U-shaped troughs that head into fiords and are believed to have been glacially formed. The valleys end abruptly at the continental slope.

The narrow channel that separates Canada from northwest Greenland consists of two basins - Hall and Kane - joined by Robeson and Kennedy channels and Smith Sound. In the north, at the narrowest part, the two shores are separated by only 10 mi. The surface of the floor is hummocky. The shallowest part is in the north of Kane Basin where there is a submerged watershed at 300 to 600 ft. In both directions the floor sinks away to 1,300 to 1,600 ft near the Lincoln Sea and about 2,100 ft at the south end of Smith Sound.

The bathymetry of Jones Sound is not well-known. The central part may have depths of about 2,300 ft, and this is joined by entrenched valleys that cross the shallower sea floor on either side. Lancaster Sound has a wide flat floor that drops eastward to 2,600 ft where it enters Baffin Bay. As in Jones Sound, submerged valleys enter the main trough from the side fiords and channels.

The surface circulation in Baffin Bay is counterclockwise. The West Greenland Current flowing north along the coast is formed by the combination of the Irminger and East Greenland currents near Kap Farvel and is relatively warm for the latitude. Its velocity probably decreases from about 1 knot in the south to less than half this amount in northern Baffin Bay where complex gyrals develop as the current is deflected to flow south off the Canadian coast as the cold Baffin Current. This current also receives water from Kane Basin, Jones Sound, and Lancaster Sound, although the exact quantities are not known and are probably variable. In both sounds it appears as though there is a westerly setting furface current on the north side and probably for the whole width of Lancaster Sound below 650 ft. Icebergs in the Baffin Current may move on the average of 0.3 knot.

Three water masses have been recognized in central Baffin Bay. In the upper

650 ft the temperature is about -1.5°C and the salinity 33.8°/oo; the characteristics of this mass originate from a combination of the West Greenland Current, the arctic inflow through the channels, and local cooling. Beneath the upper layer and down to 3,300 ft is a warmer mass with a temperature above -0.5°C and salinity of 34.2 to 34.5°/oo. In the deepest parts of the basin there is cold water with a temperature below -0.5°C and salinity of 34.45°/oo. The upper two layers are also present in Lancaster and Jones sounds.

B.4.2.3 Labrador Sea & Gulf of St. Lawrence

Off northern Labrador the mountainous coast is matched by a rapid deepening of the of the sea floor, and the 300-ft isobath is only a few miles offshore. Deep, glaciated trenches lead from the fiords across the shelf. Many of the fiords have sills at their mouth at depths of 65 to 200 ft, and the waters behind reach depths in excess of 650 ft. South of Nain the continental shelf is generally wider and inshore depths are correspondingly less. All of Newfoundland is surrounded by a wide continental shelf. Off the Strait of Belle Isle and northeastern Newfoundland the shelf is 120 mi wide and broadens to 160 mi at the Grand Banks.

The Labrador coast is washed by the Labrador Current. This great stream of cold water pours south in a belt approximately 150 mi wide. The inshore part is uniformly cold with surface temperatures not above 5°C in late summer as far south as Hamilton Inlet; and off the Atlantic coast of Newfoundland the surface temperature is lower than 10°C.

The Gulf of St. Lawrence is a shallow continental sea that is penetrated by two deep channels. The larger, the Laurentian Channel, extends from the continental shelf south of Newfoundland almost to the Saguenay River. For much of its distance it is about 1,400 ft deep. Its origin is controversial. A shorter and shallower channel separates from the main trough and then divides again to pass north of Anticostia Island and into the northeast corner of the Gulf; its average depth is 850 ft.

In addition to the St. Lawrence water, water masses enter the Gulf through Strait of Belle Isle and round the southwest corner of Newfoundland. The outlet is mainly on the south side of Cabot Strait. The northern current carries cold water into the Gulf, and the whole of the Quebec north shore has sea surface temperatures from 8 to 11°C in late summer. The water entering on the north side of Cabot Strait is relatively warm and temperatures of 15 to 16°C are only a degree or two colder than the shallow waters around Prince Edward Island.

B.4.2.4 Hudson Strait and Bay

Hudson Strait is deeper than most continental seas. At the southern entrance maximum water depths are about 2,250 ft. In the central section of the Strait, the sea floor is commonly below 900 ft and reaches 1,375 ft. Comparable depths occur at the west end. Coastal waters are shallow in the bight between Big Island and Foxe Peninsula.

The overall movement of water in Hudson Strait is from west to east and the mean outflow for the whole Strait is about 0.18 knot. There is, however, an inflow of the Baffin Island Current along the north side of the Strait which brings small icebergs into the Strait. Most of this water turns and joins the outflow on the south side but part penetrates into southeastern Foxe Basin.

The division between Hudson Strait and Ungava Bay is the 1,300-ft isobath. Water depths around Akpatok Island approach 300 ft and there is probably a trench running south, east, and northeast around the island with depths of 650 to 1,000 ft.

Tidal ranges in general in Hudson Strait are high (20 to 40 ft). On the west side of Ungava Bay they are among the largest in the world and at Leaf Basin can approach 60 ft.

Foxe Basin is exceptionally shallow. A large part of the southeast and northeast sectors are less than 100 ft deep. In the absence of soundings, it must be considered that all the western Baffin Island coast south of Air Force Island is too shallow for ships. Small boats have worked on these coasts but have experienced great difficulty with the wide tidal flats. Contrasting strongly with these conditions, depths of 600 to 1,500 ft occur in Foxe Channel where there is a broad submarine trough adjacent to Southampton Island. Surface currents in Foxe Basin are largely tidal. There is a general cyclonic movement which carries water that enters the Basin through Fury and Hecla Strait south at 0.3 to 1.1 knots. There is a smaller inflow of water along the west side of Foxe Peninsula.

Prince Regent Inlet, Gulf of Boothia, and Admiralty Sound have similar submarine topography. All are extremely shallow at the southern end. Depths increase northward in what are almost certainly submerged glacial troughs, becoming greater than 2,000 ft. Sills separate the sounds from Parry Channel.

Hydrographic surveys of Hudson Bay are incomplete but it is known that depths in excess of 900 ft occur and that the average depth is about 300 ft. Limestone rocks that outcrop in the northern and southern coastal areas are found beneath the Bay except on the east side. Several of the main river valleys entering Hudson Bay can be traced subaerially for considerable distances and are part of an original drainage system that flowed northeast into Hudson Strait.

James Bay, with an area of about 20,000 sq mi, is even shallower than Hudson Bay; the deepest part is little more than 150 ft and occurs in an elongated sea-floor depression that trends north-south and a little east of the center of the Bay. The west coast is very shallow with depths of 50 to 150 ft. The largest island in James Bay, Akimiski Island, is separated from the west coast by a shallow channel, passable only to small boats at high tide.

The James Bay trough extends northeast in a depression on the sea floor at about 150 ft; the depression then deepens to below 400 ft parallel to the arc-shaped central east coast of Hudson Bay. Westward, the bottom rises forming a plateau area less than 200 ft deep on which are located the Belcher Islands. Similar shallower zones surround the Ottawa Islands, Mansel and Coats islands, the south side of Southampton Island and the Hudson Bay Lowland in northern Ontario and Manitoba. Between Mansel and Coats islands there is a narrow trench over 600 ft deep that connects with a channel in the west end of Hudson Strait.

An extensive shallow water zone borders the west coast of Hudson Bay south of Chesterfield Inlet. North of 61°N the coast is rocky with many islets and reefs; farther south an 8-ft tidal zone of mud and boulders widens to several miles.

The circulation in Hudson Bay is counterclockwise. Part of the circulation is closed but there is inflow through Roes Welcome Sound and Fisher Strait (between Coats and Southampton islands) and from rivers. Outflow is between Coats Island, Mansel Island, and the northwest corner of Quebec. Preliminary data suggest that the water mass has arctic characteristics with bottom water temperatures of -20°C.

B.4.2.5 Canadian Arctic Archipelago

The continental margin northwest of the Canadian Arctic Archipelago consists of a shelf that slopes seaward and drops in a series of shallow steps to depths of 1,300 to 1,500 ft, 40 to 60 mi offshore. The channels between the islands are interpreted as drowned river valleys. Clearly, however, they have been much modified by glacial action and they characteristically have steep trough sides and a horizontal, although somewhat irregular, floor that is separated by sills into basins. The general depth in the channels is about 1,300 ft.

The virtual absence of ships from the ice-filled waters of the western Queen Elizabeth Archipelago has restricted the collection of oceanographic data. In the last decade, however, there has been successful sounding and sampling through the sea ice. The movement of ice islands in the channels and the distribution of driftwood suggests that there is a general southeasterly and southerly movement of surface water from the Arctic Basin through the Archipelago. Stronger local currents result from tides although the low tidal range must limit them except in narrow channels. The highest tides are found in the southeastern part of the area where at Resolute the maximum tidal range is about 6 ft. The range decreases westward through Parry Channel until along south Melville Island it is about 4 ft and on Prince Patrick Island 1.2 ft. Similar low ranges occur in the more northerly islands of the Sverdrup Basin.

Little is known of the oceanographic characteristics of the southwestern waters of the Archipelago. The submarine topography is interpreted most easily as a partly drowned land with today's channels representing major river valleys that were modified by Pleistocene ice. The main river in former times flowed from the east end of Queen Maud Gulf westward through Dease Strait, Coronation Gulf, and Dolphin and Union Strait to Amundsen Gulf. The lowland (now drowned) that it occupied deepens unevenly westward with local shallow sections notably at the east end of Dophin and Union Strait where one section is probably no more than 60 ft deep. In Amundsen Gulf ice erosion has resulted in depths of between 200 and 1,500 ft. Farther east in Coronation Gulf depths are about 600 ft, and on the north side of Queen Maud Gulf they vary from 180 to 350 ft. The shallowest area is in the east at Simpson Strait where shoals are numerous and the deepest channel is possibly no deeper than 35 ft.

Few reliable data are available on depths in Victoria Strait or M'Clintock Channel, both of which have been almost continuously blocked by ice since their discovery by Europeans. Channels suitable for northern movement of ships toward Parry Channel are restricted to Prince of Wales Strait in the west and Franklin Strait and Peel Sound (or Bellot Strait) in the east. Prince of Wales Sound has a general depth of 120 to 180 ft and ships of any draft can pass through it. In the eastern channels depths apparently increase from the southern end of Franklin Strait, where they are of the order of 300 ft to about 1,400 ft opposite Bellot Strait. This depth is maintained northward, but at Prescott Island the floor of Peel Sound rises to form a sill at 800 ft.

There are few observations about currents in this area, although it is believed that the majority are the result of tides, changing barometric pressure, and winds. The tidal range everywhere is low - of the order of 2 to 3 ft for spring tides to 1 ft or less at neap tides. The spring tides show a semidiurnal form but at neap tides, at least locally, a diurnal solar tide appears. All larger fluctuations of sea level are a direct result of pressure and wind changes;

and these, particularly the neaps, may completely override the normal tide.

The continental shelf between Point Barrow and Cape Prince Alfred varies in width from less than 50 mi in the west to nearly 100 mi in the center and northeast. Off Alaska the profile is characteristic of the continental oceanic margin in all parts of the world, consisting of the flat shelf and a steep upper continental slope that decreases as it reaches the Canada Basin at about 11,400 ft. East of the border the slope diverges from the mainland in a curving arc toward northwestern Banks Island, and the upper slope is gentler in this area than farther west. The continental shelf is crossed by several deep valleys. In the west is the Barrow Sea Valley which originates off Cape Franklin and deepens northeastward to become 20 mi wide and U-shaped north of Point Barrow. A second valley is 55 mi to the east. Northwest of Mackenzie Bay is the Mackenzie Sea Valley, a broad flat-bottomed valley apparently leading from a Pleistocene outlet of the Mackenzie River. One hundred forty miles to the northeast a smaller valley has been recognized. The largest of the shelf valleys is a broad asymmetrical feature, the Amundsen Trough, between Banks Island and the mainland. It has an average width of 24 mi and the floor is at about 1,400 ft. It is believed to mark the course of a major outlet glacier from the Pleistocene ice sheet.

Surface currents in the Beaufort Sea are light and probably irregular, depending largely on wind and pressure changes. The clockwise circulation of the Arctic Basin sets southwest and west at roughly 0.1 knot over the continental slope, but there is little evidence that it is found inshore. Along the Alaskan coast there is thought to be a reverse (eastward) current.

B.4.2.6 Bering and Okhotsk Seas

The submarine topography of the marginal seas of the north Pacific has unusual variety. The northern half of Bering Sea (from Unimak Island to Mys Navarin) is a continental sea with few areas deeper than 300 ft. The southern half, in contrast, is oceanic with depths in excess of 10,000 ft except close to the Aleutian arc where two submarine ridges project north from the islands. The more westerly reaches the Russian mainland and separates a small basin from the main Bering Sea; the easterly ridge curves back toward the Aleutians about 160 mi north of the arc.

The Sea of Okhotsk also shows striking differences between the northern margins, including the Kamchatka and Okhotsk coasts, which are underlain by typical continental shelves with depths of less than 600 ft, and a broad central basin more than 3,300 ft deep. A wide underwater lowland leads into the basin from Zaliv Shelekhova in the northeast. The basin deepens rapidly to a trough parallel to the Kuril arc south of 48°N, and the deepest point is more than 11,000 ft below sea level.

The currents in the Bering Sea are not well-known except for observations near Bering Strait. It is clear that a cold current flows southwestward along the Kamchatka coast and becomes strong off the Kuril Islands. Elsewhere there are several closed circulations and a net inflow to the sea between the Aleutian Islands. The general circulation in the Sea of Okhotsk is counterclockwise and runs strongest 20 to 40 mi offshore. A complex pattern of currents has been reported from the middle of the sea. The current moves north along western Kamchatka at 0.5 knot; in other sectors velocities up to 1 knot are recorded. The Sea of Okhotsk has one unusual feature which affects both the ice conditions and the climate. This consists of patches of substantially colder water (1 to 6°C) that are located at the entrance of Guba Shelekhova and Ostrova Iony and between the Ostrov Shantar and northern Sakhalin. These cold spots mark upwellings of colder deeper water the cause of which is still in doubt. They are associated with continuous dense fogs in the summer and with concentrations of ice in the winter. The exchange of water with the Pacific Ocean is not great due to the shallow depths (usually less than 1,600 ft) of the channels between the Kuril Islands. Pacific waters normally enter through the northeastern channels and Okhotsk water leaves by the southwestern channels.

B.4.3 Acoustics

The propagation of underwater sound in the Arctic Ocean differs in several ways from that in nonpolar oceans. In the Atlantic and Pacific oceans, the SOFAR channel lies at depths of 3,250 to 4,500 ft, but in the Arctic it is at the surface. Low-frequency sound is propagated to great distances in the Arctic SOFAR channel. Sound rays are alternately refracted upward in the water and reflected downward from the base of the ice. At great ranges signals consist predominantly of low frequencies, between 8 and 100 kHz. The roughness of the lower surface of the ice strongly attenuates the high frequencies but has a negligible effect on the low frequencies. Signals generated by small explosions have been recorded clearly at distances from 40 mi up to 1,500 mi. Beyond a range of 300 mi, sounds above 100 Hz are very weak. The Arctic SOFAR signal is dispersed so that an impulsive signal increases in duration as the range is increased. At a range of 300 mi the duration of a signal from a small explosion is about five seconds. At shorter ranges and over smooth bottoms such as abyssal plains, bottom-reflected arrivals can be of importance. They are late arrivals and increase still further the duration of the signal.

In the shallow water of the shelves, propagation characteristics depend strongly on bottom parameters. In general, long-range transmission is much more strongly attenuated along shallow-water paths than it is along deep-water paths. Dispersion is even more pronounced in shallow-water transmission.

B.4.3.1 Transmission Loss

The stable positive sound velocity profile in the Arctic Ocean results in a half-sound channel bounded at the interface of the water and ice. The divergence loss (N_d) in this situation can be approximated by:

$$N_d = 20 \log \frac{r_o}{4} + 10 \log \frac{R}{r_o/4} = 10 \log r_o + 10 \log R - 6$$

where r_o is the skip distance of the deepest traveling ray and R is the range. This equation represents first spherical spreading to one-quarter of the distance traveled by the deepest limiting ray from the source to where it strikes the surface and then cylindrical spreading thereafter.

In the Arctic the only other significant loss at low acoustic frequencies, where absorption can be neglected, is the loss suffered at each ray reflection at the ice-water interface. For long-range propagation where many reflections occur, this reflection loss can most simply be handled as a loss per unit distance. This loss will be a function of frequency, since the degree of roughness of the ice reflecting surface - and hence its scattering power - is a matter of the wavelength of the impinging sound. Therefore the total transmission loss (N_w) is represented by total loss = divergence loss + reflection loss, or

$$N_w = N_d + N_r = 10 \log r_o + 10 \log R - 6 + N_r' R$$

where N'_r is the reflection loss per unit distance.

Discrete frequency, smoothed curves of transmission loss in the range from 20 to 3200 Hz are given in Figure IV-7. Also shwon are the standard deviations of actual measurements from the smoothed curves. The deviations result from variations in the yield of explosive signals, reflection effects at the source and the receiver, variations of bottom depth, and measurement errors.

Since at the lower frequencies the wavelengths are comparable to the source and receiver depths of the loss measurements made in the Arctic, the Lloyd mirror effect must be considered. Figure IV-8 is a representation of this effect on transmission loss derived from the geometry of the arriving waves. Note the 6-db amplification of the signal at 40 Hz with a receiver at 200 ft. This amplification quickly becomes a loss as the receiver is moved toward the surface. The same holds true for the source depth of the sound. The curves in Figure IV-8 are idealized and represent observed results from the first minimum near the surface to the first maximum (or 6-db enhancement point) for any given frequency while below that there is considerable variation.

Figure IV-9 exemplifies the need to know the depth of the source to measure transmission loss. The two traces are for identical charge size and receiver depth with varied source depth. Note the greatly increased bottom bounce arrivals from the shallow shot. These arrivals comprise more than half of the total arriving energy, the rest being in the water-traveling rays. For the deep shot, the bottom bounce energy is a very small fraction of the total energy. The reason for this is that, for the shallow shot, reflections close to the source enhance the ray emanations which depart at the greater angles and which propagate as bottom-bounce rays. For the deep shot the small-angle, shallower-

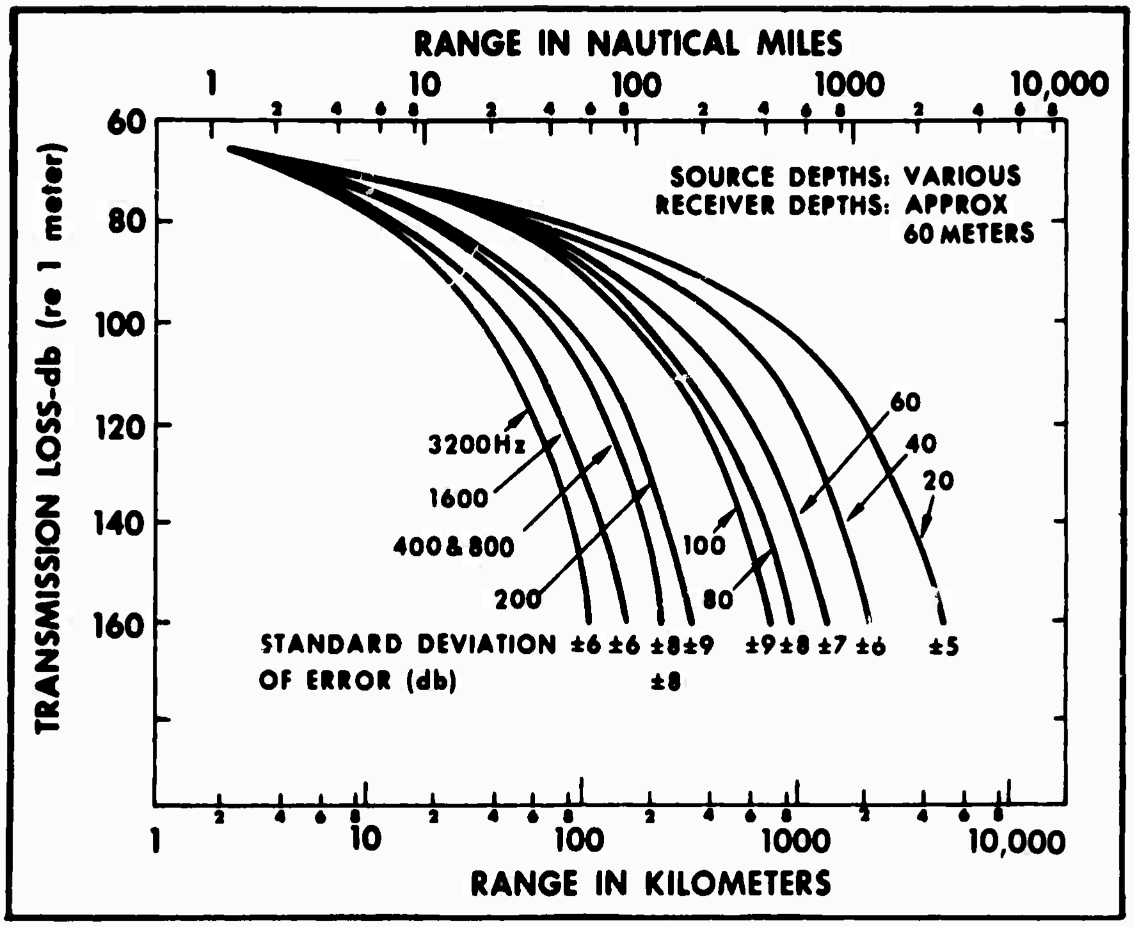

Figure IV-7 BEST-FIT TO TRANSMISSION LOSS MEASUREMENTS IN ARCTIC OCEAN

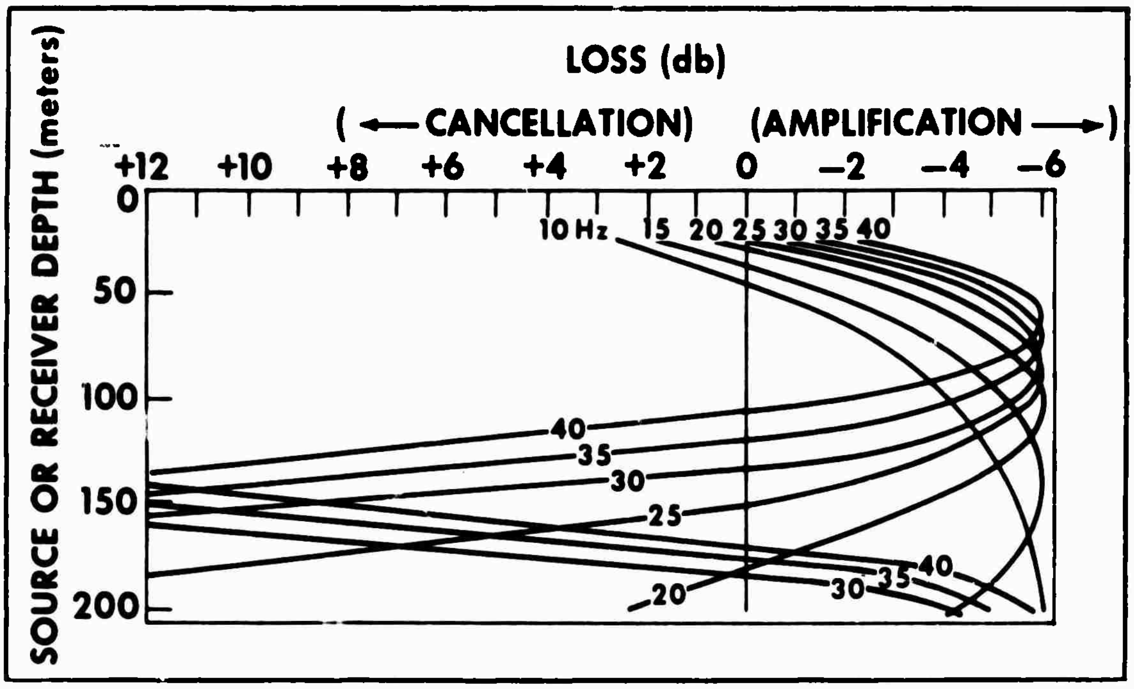

Figure IV-8 EFFECT OF SURFACE REFLECTION NEAR THE SOURCE OR THE RECEIVER ON SIGNAL

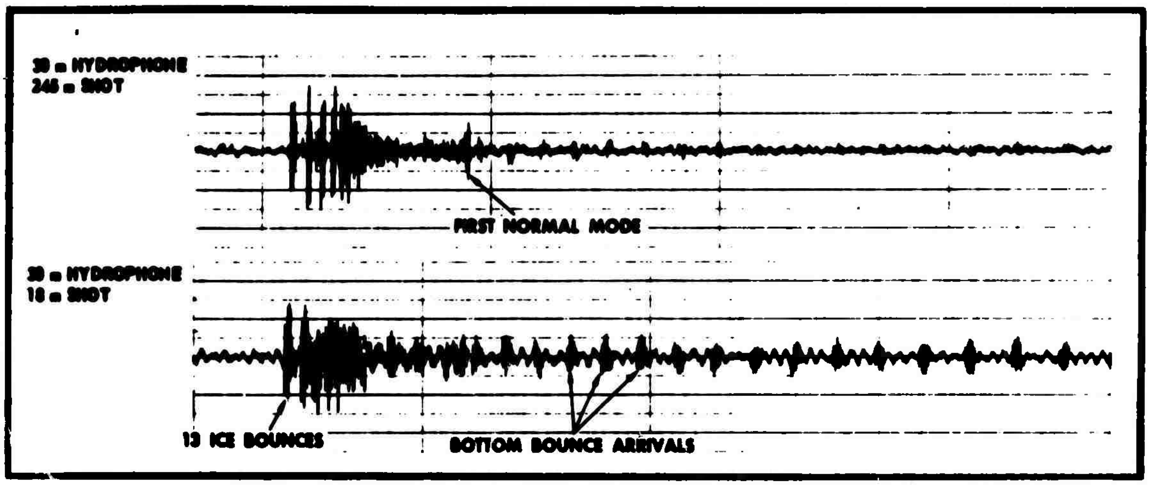

Figure IV-9 EFFECTS OF SOURCE DEPTH ON ARRIVING SIGNAL ENERGY traveling rays that do not strike the bottom are enhanced and the deeper-traveling rays are not.

Bottom depth along the path plays an important role in determining the transmission loss. Figure IV-10 shows the typical explosive signal arrival over a deepwater path. The first energy to arrive (group A) is from deep-traveling rays that arrive as individual energy packets. The next group (B) is comprised of the shallow-traveling rays that arrive in phase addition, giving what is sometimes called the first normal mode arrival. The last group (C) are the bottom-bounce rays that arrive at ever-increasing intervals. If there is water shallower than about 1,800 ft along the transmission path, ray theory predicts that group A arrivals will not be present. This represents a considerable reduction in total signal strength. If the bottom is deep but the source depth is below about 1,800 ft there will be an A and a C group but no B, since those rays propagate shallower than 1,800 ft. For a source below 1,800 ft with water shallower than 1,800 ft anywhere along the path, both groups A and B will be absent, leaving only C. These conditions greatly affect the energy of the arriving signal and explain some of the deviations mentioned above.

B.4.3.2 Ambient Noise

A one-year record of ambient noise levels is given in Figure IV-11. Although there are undoubtedly contributions from biological sources, the background noise in deep arctic water is attributed largely to ice activity. At the low end of the frequency spectrum the background is, in all probability, due primarily to gross ice movement (pressure ridging) that originates from nearby out to extreme distances. In the mid-frequency range, local ice-fissure generation and distant gross ice movement probably contribute about equally to the

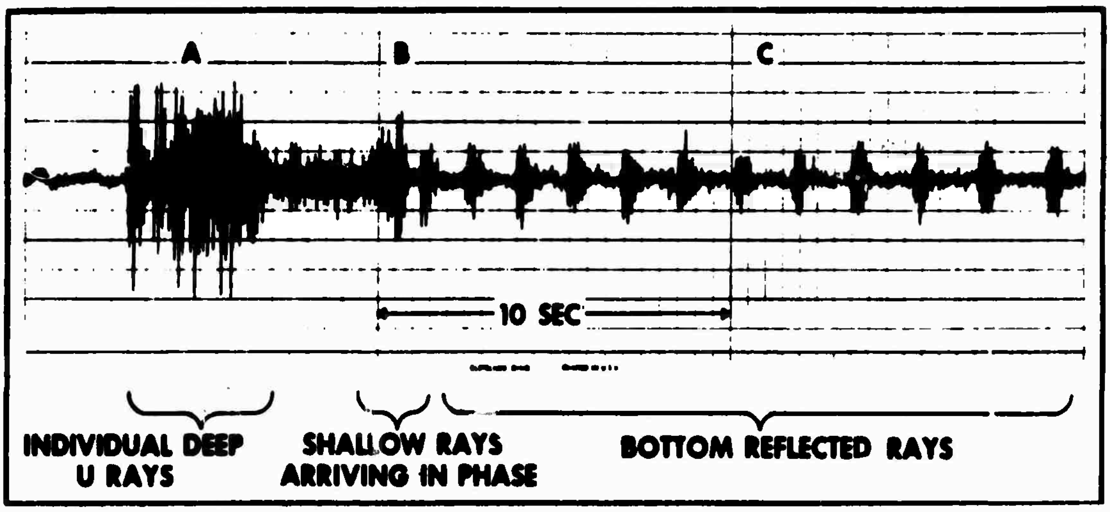

Figure IV-10 TYPICAL EXPLOSIVE SOURCE SIGNAL ARRIVING FROM LONG DISTANCE OVER A DEEP-WATER PATH

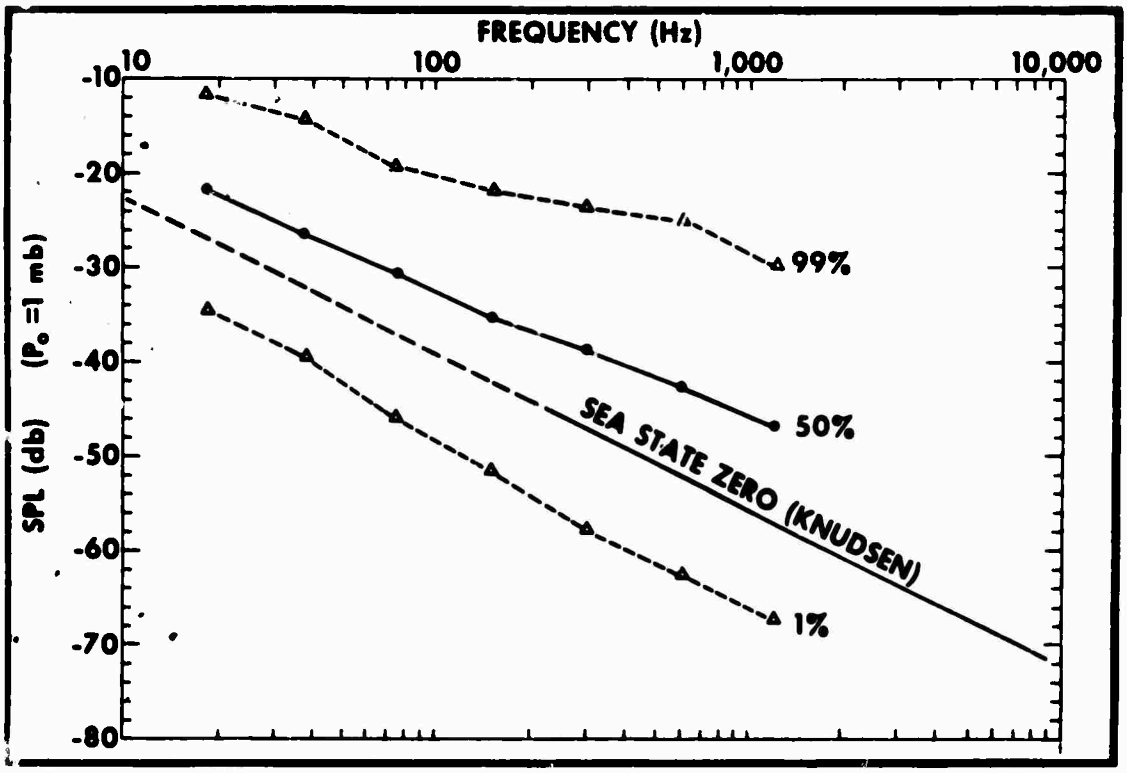

Figure IV-11 MEDIAN, 99, AND 1 PERCENTILES OF SOUND-PRESSURE LEVEL OF AMBIENT NOISE FOR PERIOD APRIL 1965 THROUGH FEBRUARY 1966 noise. The high frequencies are dominated by local ice cracks, which are caused by thermal changes during periods of low wind, and by wind turbulence at the air-ice interface during periods of high wind.

The presumption that low-frequency noise is more dependent on remote ice activity than high-frequency noise has been shown to be true. However, there is a systematic decrease in the degree of correlation at lower frequencies, suggesting that the lower frequencies are more affected by distant ice movement. There is enough correlation, nevertheless, to indicate that both the low and the high ends of the range are influenced by nearby ice activity.

B.4.4 Marine Biology

Nansen believed that the phytoplankton was sparse in the Arctic Ocean, but Soviet investigators found a considerable development. Others reported an intense bloom of diatoms in the lower layers of sea ice. Beyond these, only lists of species taken in net samples are available. Soviet samples from North Pole II in the Arctic Basin were collected. During the International Geophysical Year, chlorophyll *a* was used to determine the standing crop of phytoplankton, and carbon-14 uptake was used to estimate the amount of primary production. The annual productivity was found to be very low in comparison with that of other ocean areas. A net retained less than 10% of the chlorophyll *a* retained by the membrane filter, since many of the photosynthetic organisms in the arctic seas, perhaps the majority of the biomass, are minute, difficult to preserve, and often impossible to identify from preserved material. A relationship between cell counts, chlorophyll *a* concentration, and carbon-14 uptake has not been demonstrated.

The oceanographic variables affecting the magnitude of primary production are known, but the relative importance of each in the arctic situation must be assessed further. Many proposed mechanisms for the initiation of the annual phytoplankton bloom and the possible importance of production within the ice are based on ignorance of the arctic environment and the dynamics of primary production in the ocean. The surface water in the Arctic Ocean is not rich in dissolved inorganic nutrients and is not responsible for high nutrient concentrations encountered near the periphery of the permanent pack ice. The upper and lower limits of the Pacific water mass in the Canada Basin are regions of significant density change, and the layers near these limits are rich in detritus and nutrients. These limits form two scattering layers in the Beaufort Sea - one occurs at a depth of 150 ft coincident with the pycnocline between the surface Arctic and intermediate (Pacific) water masses. This layer (Pycnocline scattering layer) is deflected at 100 kHz frequency and correlated with an accumulation of pteropods, *Spiratella helicina*. The second layer is a more typical deep scattering layer which appears seasonally in the northern part of the Beaufort gyral; it is detected at 12 kHz and 100 kHz. Although no fishes have been captured within the layer due to the problem of trawling through pack ice, the appearance of the layer on the depth sounder implies a nektonic organism. The Arctic cod, *Arctogadus glacialis*, is probably the responsible species.

A small oxygen-minimum layer is present at the interface of the intermediate (Pacific) water and the Atlantic water at 660 ft. There is no evidence to show any special biological cause for this depletion, therefore physical manifestation of circulation and vertical eddy diffusion are probably responsible.

Published studies on the zooplankton under the permanent pack ice are even more fragmentary than the information available for the phytoplankton. Faunal lists are available from several expeditions and recent work on floating ice platforms; but the life histories, physiology, growth rates, reproduction, relative abundance, population dynamics, vertical distributions, and behavior of the animals are almost totally unstudied. The consequences of low primary production, very low temperatures, long light and dark seasons, and the "inverted bottom" formed by the pack ice are of enormous zoological interest and should be investigated extensively.

The nekton, including seals, are quite difficult to capture from the permanent pack ice. Most sampling devices for nekton and macroplankton depend upon a horizontal component of force provided by a moving ship, and the drift of the ice is generally too slow for such a purpose. Of the fishes only the polar cod (Boreogadus) has been taken from the center of the Arctic Ocean, although bottom photographs have shown at least one other species of fish.

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B.5 CLIMATE

B.5.1 General

Four aspects of the northern climate may be said to characterize the Arctic. The first is a distinctive regime of daylight and darkness together with low solar elevation which give rise to a prolonged period of radiational loss from the earth's surface. Second are the surface weather systems associated with the large-scale, cold-cored circumpolar vortex present in the free atmosphere over the area. This is a function of the differential heating between the equator and pole and of the earth's rotation. These upper air flow patterns are apparent at 5,000 ft (850 mb) but reach their greatest intensity near 33,000 ft (200 mb). The surface cyclones and anticyclones responsible for much of the weather are embedded in and steered by this flow, so that changes in the shape and position of the vortex can radically affect regional conditions.

The third aspect is a surface cover of snow or ice for at least an appreciable part of the year. In cold regions the nature of the surface is largely a question of whether or not there is a snow or ice cover. The difference in the percentage of absorbed solar radiation is as great as 60% (up to 20% absorption with a fresh snow surface compared with greater than 80% in most snow-free regions). The immediate effect of changes in this cover on the heat exchange at the surface is so great that the region can be said to have two seasons only, with a swift transition of a week or two each at the spring thaw and the fall freeze-up. Over the permanent ice caps and the polar pack ice during the brief summer, the ice maintains the air temperature at about 32°F, while temperatures in snow- or ice-free areas can reach 70 to 85°F. Toward mid-latitudes in the periglacial zone, as the length of the snow-free period increases, variations from winter to winter in the timing and continuity of the snow cover become more important, and the number of freeze-thaw cycles in the year increases rapidly. The continental subarctic can experience minimum temperatures as low as the central Arctic Basin; but with the approach of total winter darkness at the pole, the persistence of low temperatures increases.

The presence of a snow or ice cover plays an important role in the modification of air masses crossing the region. Cyclones depend on temperature contrast, and generally speaking their intensity and frequency are closely allied to the sharp gradients at the margins of the snow or ice cover and open water. Over the central Arctic cyclones oftentimes stagnate and fill. The distribution of precipitation is largely related to the passage of cyclones over open water; and, as large areas of open water are conspicuously lacking, the occasional open leads in winter become important heat and moisture sources. However, the cooling of the air masses over the snow and ice reduces the water-holding capacity of the air to such an extent that the greater part of the Arctic is virtually a desert. The coldness of the air causes precipitation to characteristically take the form of small, dry, hard snow particles which are easily redistributed by the wind to produce blowing-snow storms. Middle and high clouds are also associated with cyclonic activity; but, owing to the dryness of the air, the clouds tend to be less dense over the central Arctic. Low cloud and fog, characteristic of coastal areas and the broken pack ice in summer, are essentially linked with the distribution of snow or ice and open water. The combination of overcast sky and an unbroken snow or ice surface produces the optical phenomenon of whiteout.

It is characteristic of snow and ice masses that once a certain critical growth is reached, they tend to be self-perpetuating. However, where there is a snow- or ice-free period, the relationship between the earth's surface and the atmosphere is finely balanced and any induced changes resulting in a crossing of the freeze-thaw threshold can significantly alter the complex association. Animal and human destruction of the natural cover and additions to the heat and moisture sources can initiate irreversible chain reactions.

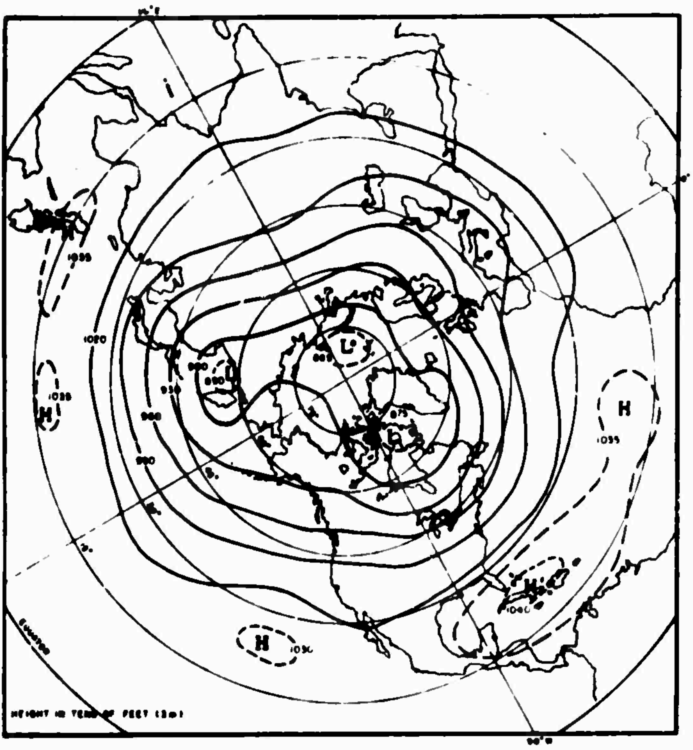
The fourth distinctive aspect of the arctic climate is a strong temperature inversion above the snow or ice surface resulting from strong radiational cooling. In winter, the average depth of this inversion layer is 4,000 to 5,000 ft (850 mb). Its development is most intense under calm, clear anticyclonic conditions, when outgoing radiation is unhindered and extremely low temperatures can occur. Inherently a stable feature under conditions of negative radiation balance at the surface in winter, it is only temporarily or partially cleared by the strong winds, cloud cover, or precipitation associated with cyclonic activity; and it rapidly reforms. In summer this type of inversion is confined to the pack ice and ice cap surfaces. It is less intense at this season and is frequently lifted off the surface by the presence of a shallow mixing layer over the ice itself. The configuration of the surface plays an important role in the local intensity and persistence of the inversion. Where the dense cold air is trapped in hollows and valleys, as for example in the Yukon and northeastern Siberia, some of the lowest temperatures have been recorded. Over the convex surfaces of the ice caps, the dense cold air appears to drain off in surges under gravitational flow (katabatic winds).

B.5.2 Atmospheric Circulation

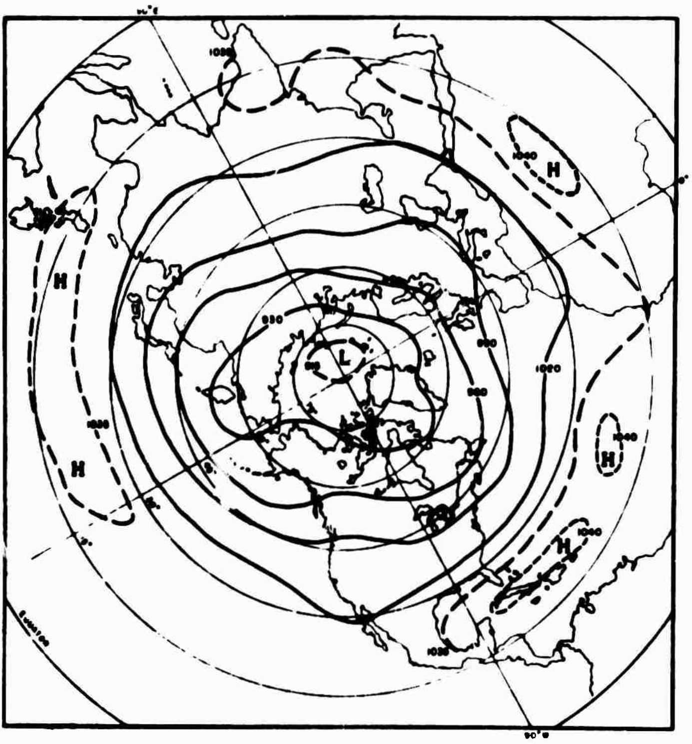
Considering both the descriptive and predictive aspects of climate at high latitudes, the dominant factor is the presence of the cold-dored westerly vortex in the middle and upper atmosphere. Its north-south temperature gradiant is generally steepest in mid-latitudes, where it increases with height to a maximum near 33,000 ft and is associated with a zone of strong westerly flow. The westerly jet streams at the periphery of the vortex play a vital role in the development and steering of the surface cyclones and anticyclones, so that variations in weather (and individual meteorological elements) from day to day, season to season, and year to year, and from one region to another, are closely linked with deformations of this upper vortex. Those areas below the periphery of the vortex are generally characterized by a succession of eastward-moving frontal cyclones, whereas regions under the sluggish cold core of the vortex experience persistent cold lows - cold, non-frontal systems resulting from the occlusion of the frontal storms. The majority of the storms are occluded before they reach the Arctic. Aloft, the cold low centers of the circumpolar vortex appear to be related to the upward growth of these occluded systems in which adiabatic cooling occurs. The movements of these cold lows are slow, erratic, and difficult to predict.

Anticyclones are typically large irregular areas of high pressure without distinct centers and characterized by subsidence and stability. In the Arctic, they are frequently shallow surface systems; but when they are associated above with the northward ridging of the circumpolar vortex and are capped by warmer air of subtropical origin, they become deep and persistent. From time to time smaller migratory anticyclones move out of the Arctic directed by the upper flow toward the southeast, but more characteristically these systems are very slow-moving and tend to oscillate around preferred positions or to change shape rather than to follow clearly defined paths.

Figure V-1 shows the mean situation for January, April, July, and October



a. 12-YEAR MEAN, JANUARY 1947 TO 1958



b. 12-YEAR MEAN, APRIL 1947 TO 1958

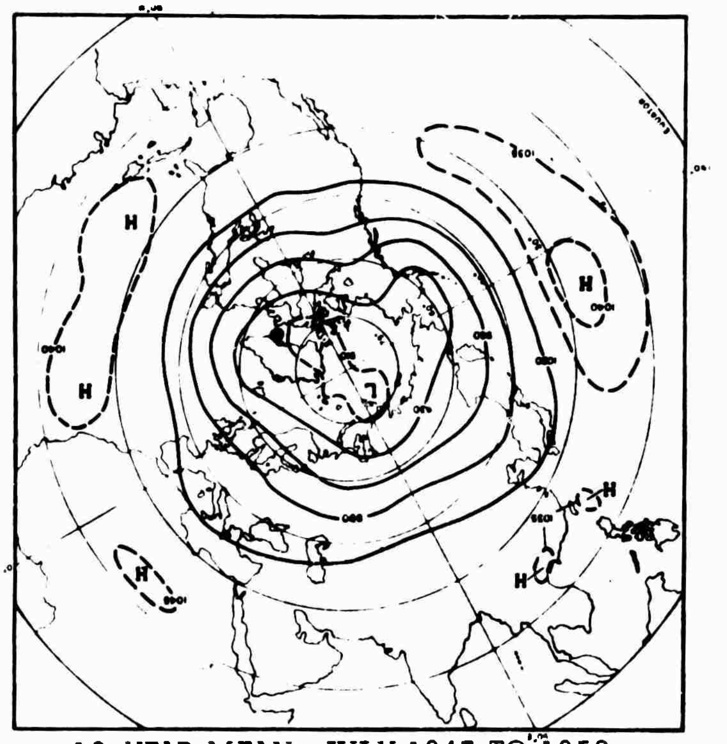
FIGURE V-1 MEAN HEIGHT OF THE 700-mb SURFACE (O'CONNOR, 1961[21]) at approximately 10,000 ft above sea level. The vortex shows a marked three- to four-wave pattern which appears to reflect the influence of the earth's major topography and the differential heating of the continents and oceans.

In winter (Fig. V-1a) the vortex shows three centers - one each over the Canadian Arctic Archipelago, Kamchatka, and Novaya Zemlya. Two major troughs occur over eastern North America and off the east coast of Asia, and there is a third weaker trough over eastern Europe. The main ridges lie over the eastern Pacific and Rocky Mountains and over the eastern Atlantic and Scandinavia, with the third weaker ridge over the Urals and western Siberia. The well-developed ridge over the longitudinal barrier of the Rocky Mountains and the deep trough downstream over eastern North America are quasi-permanent features of the general circulation throughout the year. In winter, however, they are most firmly established as the equator-to-pole temperature gradient is greatest, the vortex most intense, and the westerly flow strongest. As the gradient builds up during the fall, the vortex expands and the belt of strongest westerlies shifts southward.

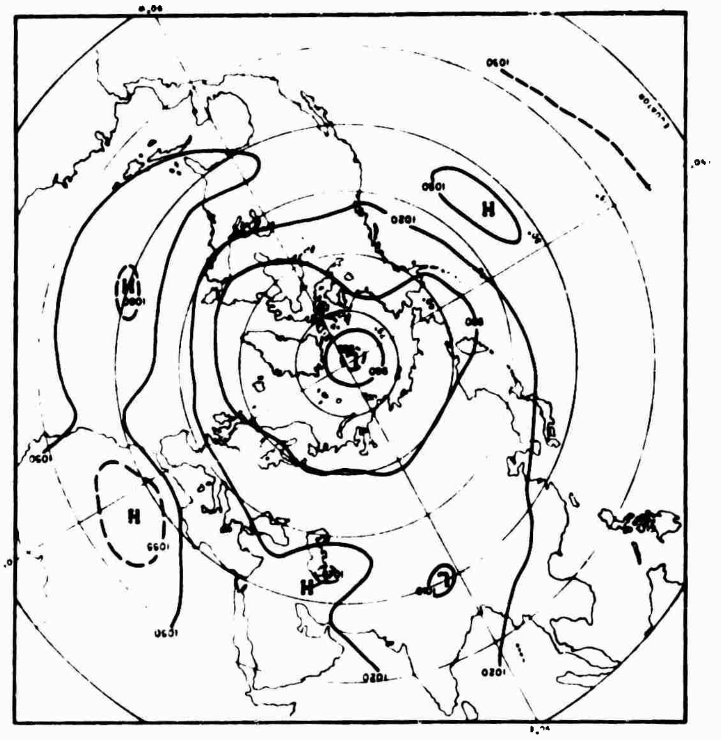
Owing to the strong southerly and northerly components of the main westerly current, cold dry arctic air is carried far to the south especially over central and eastern North America and eastern Asia, while warm moist air over the oceans of lower latitudes is drawn northward over the Atlantic and Pacific oceans toward the Arctic. In the troughs over eastern North America and Asia, temperature gradients between continental and oceanic air are strongest, and here are located the major cyclogenetic regions of the hemisphere at this season. The surface cyclonic disturbances, steered in the direction of the upper air flow, move northeast toward polar latitudes and frequently spiral in toward the upper vortex centers. As they approach high latitudes, they generally lose much of their temperature contrast, and the warm air is undercut and lifted off the ground by the colder air (i.e. the cyclone becomes occluded). Over the north Pacific and Atlantic there are the two main graveyards where many storms stagnate and fill. Occasionally cyclones are regenerated at the icewater margin, and storms continue toward the central Arctic. Anticyclogenesis occurs just east of the major ridges, the Yukon showing a particularly high frequency; while occasionally systems move out to the southeast, they can remain stationary for many days at a time.

Through the spring and in summer (Figs. V-1b and c), as the north-south temperature gradient diminishes, the vortex weakens and contracts to a single mean center near the pole. The mean location of the belt of strongest westerly flow and the corresponding tracks of surface cyclones now lie over higher latitudes. With the lessening of the energy in the vortex and its contraction to the north of the Tibetan plateau, the wave pattern is weaker and the circulation tends to be more zonal (i.e. from west to east). With the heating of the snow-free land surfaces at this season, cyclogenesis is frequent inland over the subarctic and at the ice-water-land margin of the Arctic Ocean. The mean surface pressure chart for July shows a weak amorphous circulation pattern, hiding the considerable small-scale fluctuations that can occur from day to day. Small-scale variations include the successions of weak, rapidly moving cyclones and the oscillating cold lows that persist below upper-level vortices over the central Arctic. In summer, cyclones predominate over the Arctic and subarctic, except within the zone from 70 to 75°N where anticyclones are more frequent.

With the exception of the Pacific trough, the autumn patterns both aloft (Fig. V-1d) and at the surface already closely resemble those of midwinter.



c. 12-YEAR MEAN, JULY 1947 TO 1958

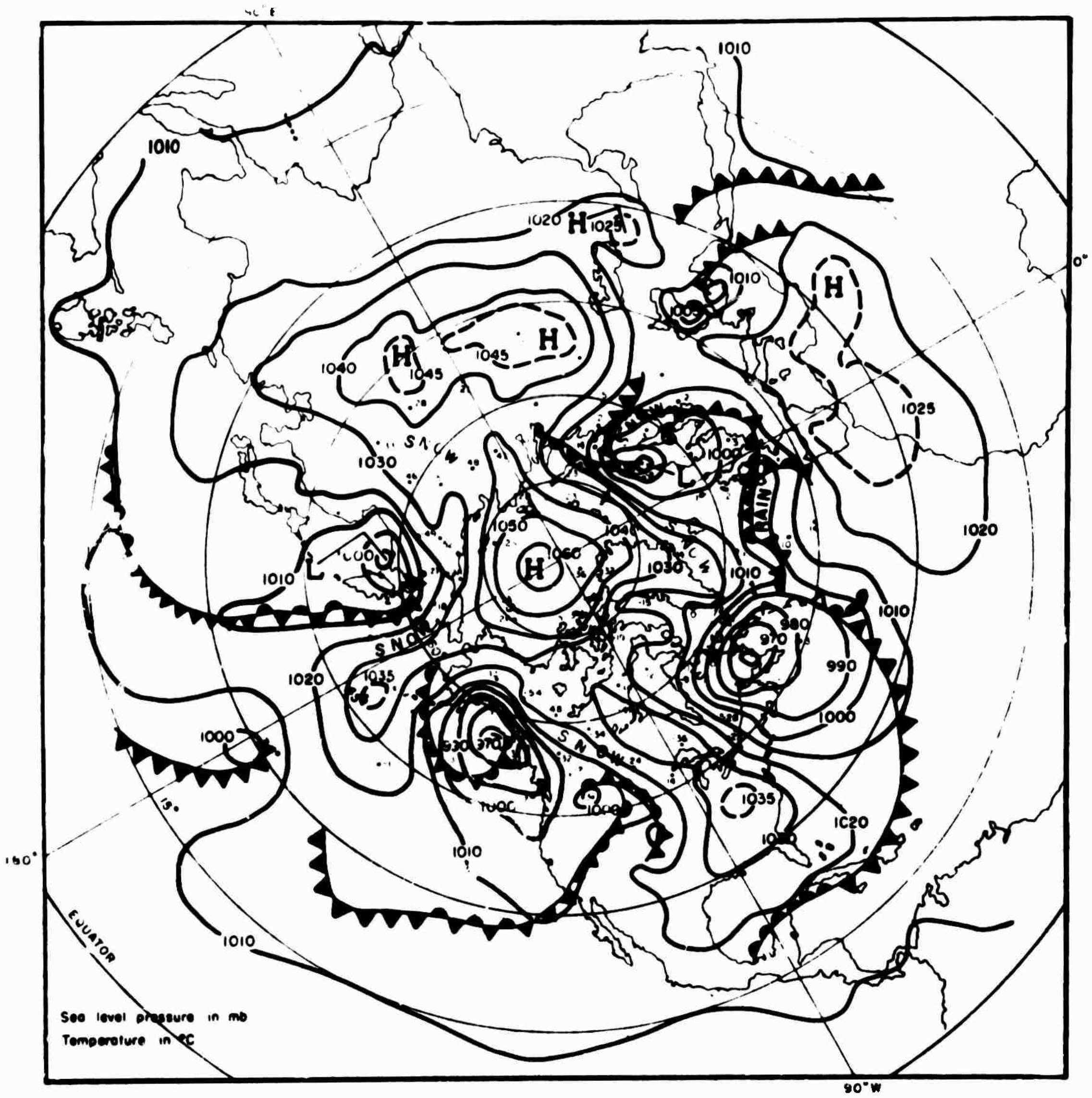


d. 12-YEAR MEAN, OCTOBER 1947 TO 1958

FIGURE V-1 (CONT'D) MEAN HEIGHT OF THE 700-mb SURFACE (O'CONNOR, 1961[21])

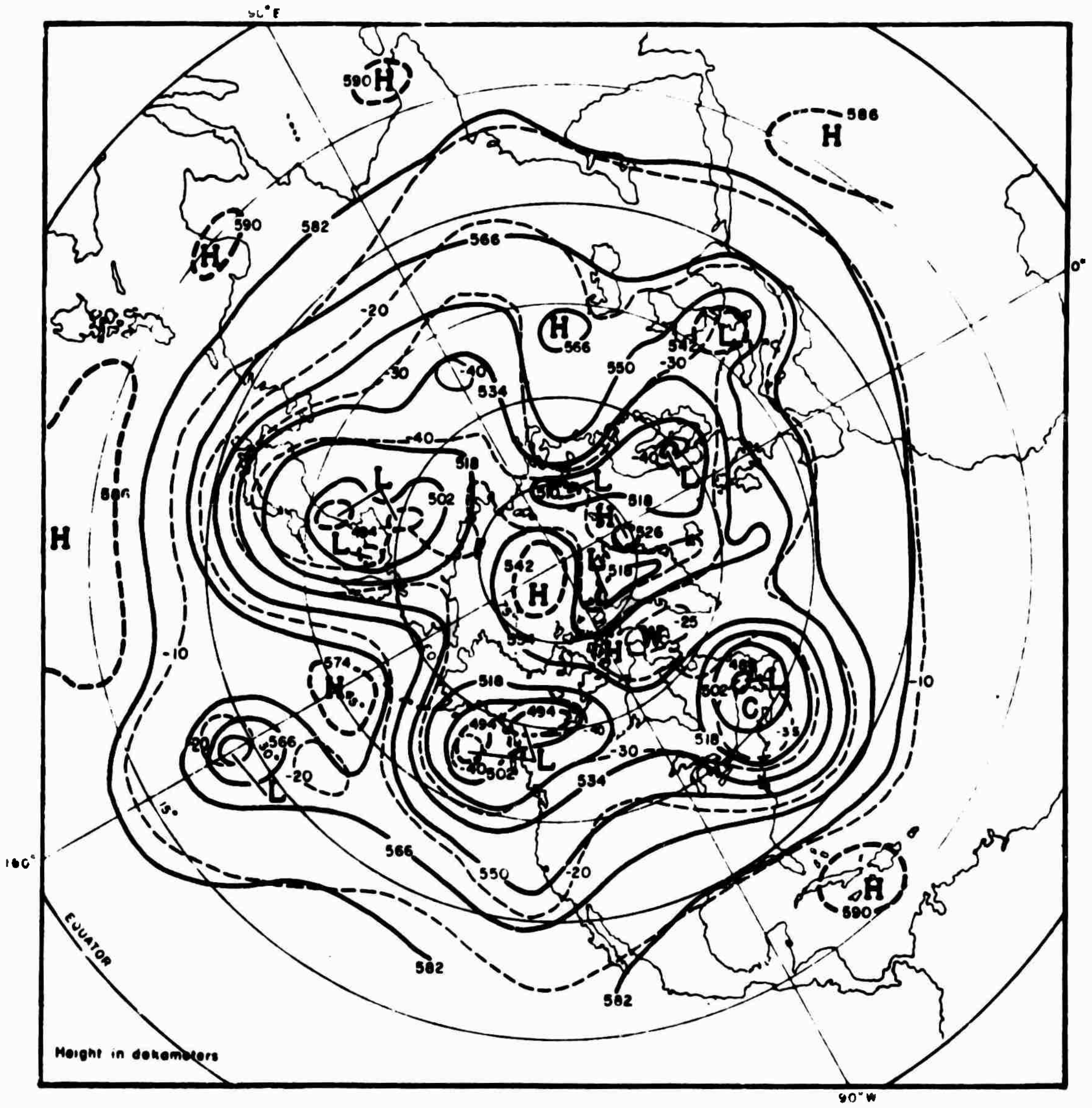
The change from a well-developed wave pattern with strong northerly and southerly components (a meridional situation) of easterly flow to a more zonal (east-west) flow pattern, already recognized in the seasonal means, is one of the most significant features of the northern circumpolar vortex with respect to arctic and subarctic climates. It is especially remarkable in the colder season, when the patterns are more strongly developed, that the vortex appears to undergo a cycle of deformation of a period of roughly a month from more zonal flow to strongly meridional conditions, when the long-wave pattern intensifies. The major cold troughs over the continents extend rapidly toward the equator, while the warm ridges over the Pacific and Atlantic reach northward, frequently as far as the inner Arctic Basin. One sector may experience a more zonal flow while meridional conditions exist at other longitudes, but the sequence and mechanism of these changes has still not been fully explored. During zonal flow, however, a slow progression of the long waves toward the east is characteristic; as the zonal flow breaks down, the situation frequently becomes quasi-stationary and then shows an apparent slow retrogression toward the west.

The effect on arctic and subarctic climate of these major changes in general circulation is best described with the aid of synoptic examples. Figures V-2a and b illustrate a well-developed meridional circulation on January 6, 1959. At approximately 18,000 ft (500-mb surface), the pattern is dominated by a warm ridge extending northwestward from the Atlantic to the pole, with other major ridges over the Pacific and western Siberia and a weaker ridge over North America. The cold vortex has four main centers located in the troughs over subpolar latitudes. The strongest westerly flow is associated with the strongest temperature gradient.



a. Sea level, 1230 GMT, 6 January 1959.

Figure V-2 SELECTED DAILY SYNOPTIC CHARTS FOR SEA LEVEL AND THE 500-mb SURFACE



b. 500-mb level, 1200 GMT, 6 January 1959.

Figure V-2 (Cont'd) SELECTED DAILY SYNOPTIC CHARTS FOR SEA LEVEL AND THE 500-mb SURFACE

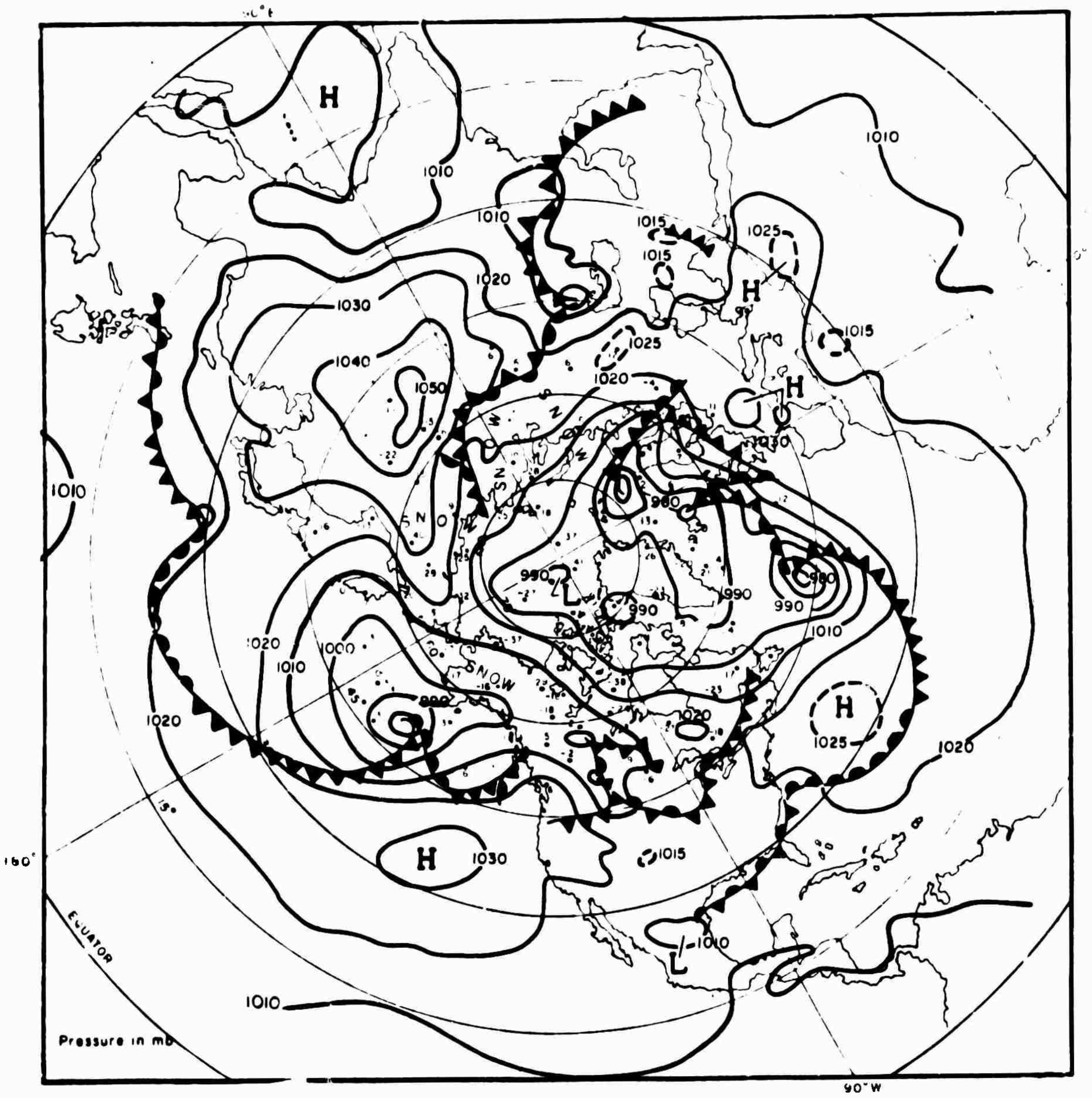
The sea-level pressure chart reflects these circumstances. Strong anticyclonic conditions exist over the Arctic Basin and south over North America and Siberia; it is essentially a cap of very cold, dense, dry arctic air. These polar outbreaks over the continental masses temporarily extend the arctic climate into middle latitudes. Frontal cyclones below the strong westerly currents are being steered around the troughs into graveyard areas - on this occasion over the Gulf of Alaska, Labrador (with a characteristic extension of the trough over Davis Strait and Baffin Bay), and Scandinavia. The upper ridge over the Atlantic-Davis Strait region successfully blocks the main line of entry of storms into the Arctic Basin, but over northeastern Siberia weak cyclonic disturbances are being guided around the upper vortex from the Sea of Okhotsk northward to the Arctic Ocean where they soon fill and disappear. Similarly, storms moving toward the Barents Sea around the European trough rapidly weaken and fill in the vicinity of the upper low center over Novaya Zemlya.

Figure V-2a also shows spot temperature reports and regions of falling snow. In general the anticyclonic areas with clear, calm conditions are the coldest with continental temperatures lower than those over the Arctic Basin. The lowest values (below -58°F) are recorded in the valleys of the Yukon, Alaska, and northeastern Siberia; and local fog (probably ice fog) is frequent where temperatures are below -40°F. Blowing snow is reported where pressure gradients are particularly steep - over the northern Canadian Arctic Archipelago for example, where strong winds and a temperature of -27°F combine to give a severe wind-chill effect.

Higher temperatures and storms are associated with the four cyclonic centers of activity and with open water where the strongest temperature and humidity contrasts and pressure gradients exist. Over northeastern Siberia and the Gulf of Alaska these gradients are emphasized by the strong relief barriers which cut off the interior from the full effect of the open ocean. The 32°F recorded on the western shore of Greenland near the open water of Davis Strait contrasts with the -58°F at the same latitude in western Canada. Over northern Eurasia in general, the temperature distribution is a mirror image of that over North America, with temperatures near 32°F at the Pacific coast falling rapidly to -58°F in northeastern Siberia and then varying between -22°F and -40°F over the central section with a rapid rise to -4°F over the White Sea and north Scandinavian coasts. Snow, cloud cover, blowing snow, high winds, and high wind-chill are generally associated with these strong cyclonic centers.

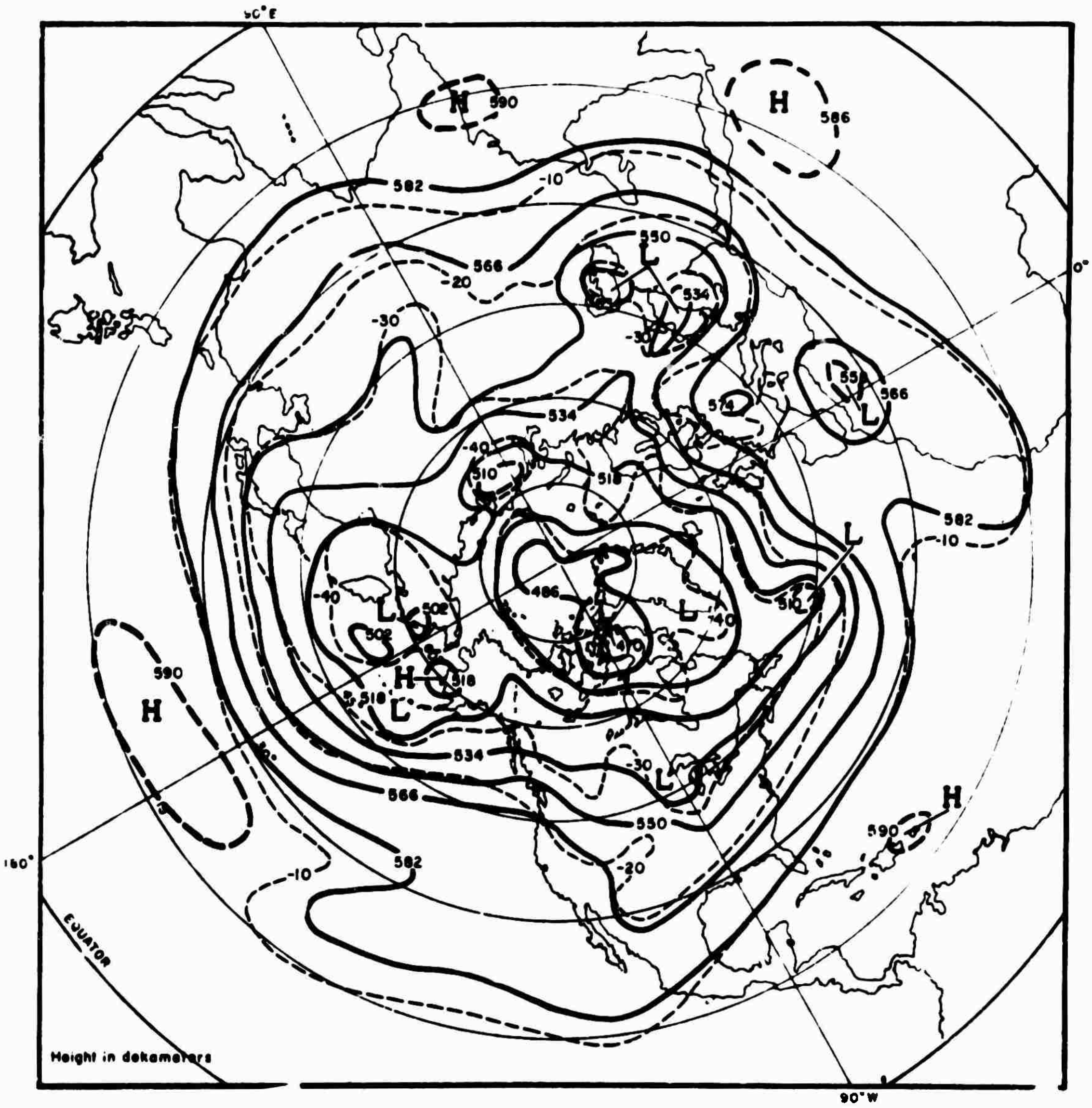
Figures V-2c and d illustrate a more zonal circulation on February 26, 1959. At the 500-mb level the vortex shows a single major cell and is more nearly circumpolar with a characteristic tendency toward asymmetry. The jet streams and associated storm tracks cross more northerly latitudes over northwestern Europe and west-central Siberia.

At sea level cyclonic circulation dominates the Arctic and subarctic except for the northeastern extension of the Siberian anticyclone linked by a narrow weak ridge to a small high-pressure center south of Hudson Bay. Trains of frontal cyclones move eastward with the westerlies, spiraling in toward the upper vortex center. In general, circulation patterns are weaker over northern latitudes below the central vortex with less striking regional contrasts in weather than on January 6. The major difference lies in the weak ridge development over the Pacific and Atlantic; instead there is a ridge of high pressure over central Europe and Scandinavia, and a series of Atlantic storms is being steered through the North Atlantic gateway and into the central Arctic under



c. Sea level, 1230 GMT, 26 February 1959.

Figure V-2 (Cont'd) SELECTED DAILY SYNOPTIC CHARTS FOR SEA LEVEL AND THE 500-mb SURFACE



d. 500-mb level, 1200 GMT, 26 February 1959

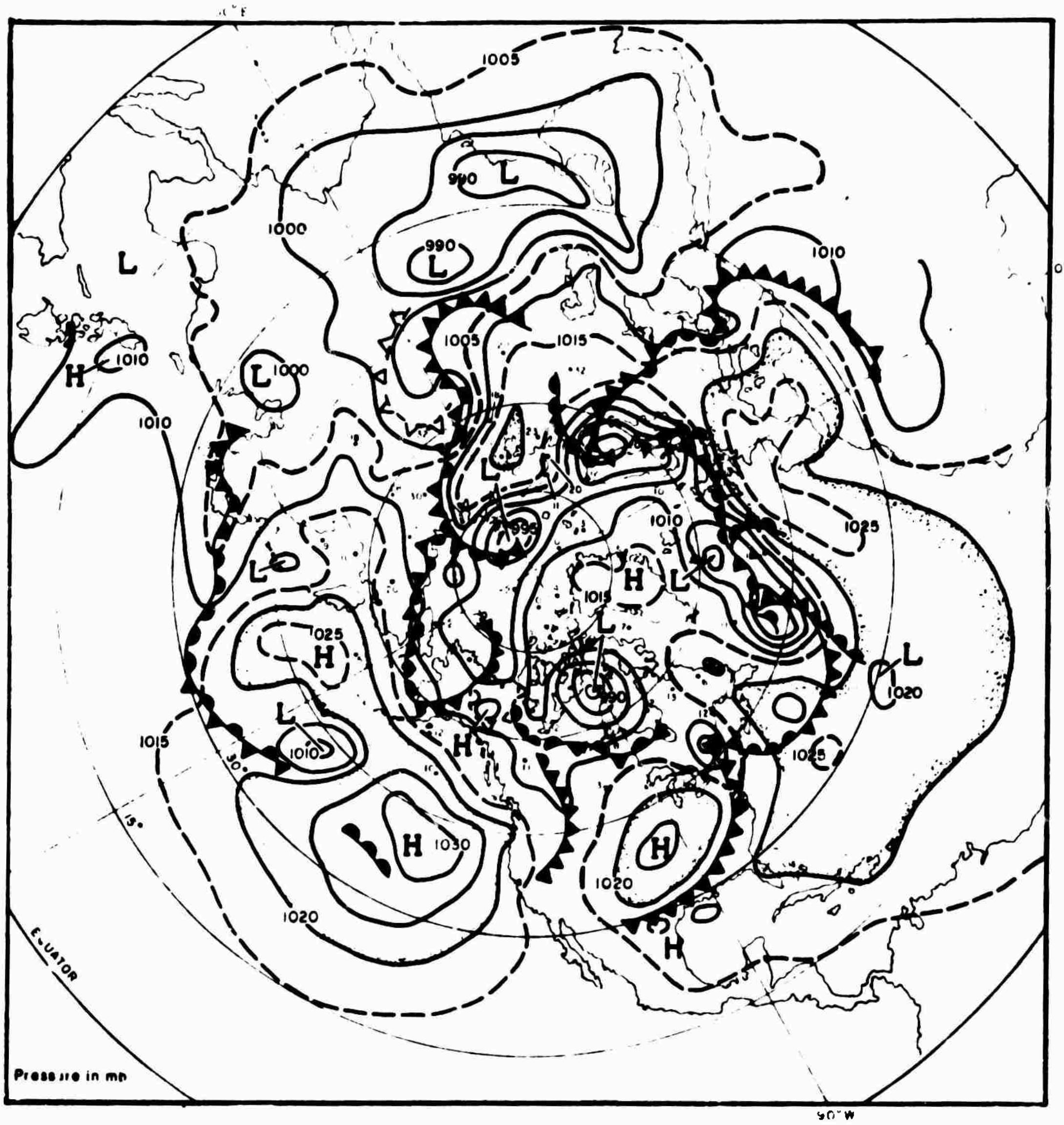
Figure V-2 (Cont'd) SELECTED DAILY SYNOPTIC CHARTS FOR SEA LEVEL AND THE 500-mb SURFACE

the vortex center.

The cold low center at the surface near the pole is giving neither cloud nor precipitation and winds are light, so that surface cooling is unhindered and temperatures as low as -35 and -42°F are reported on ice islands near the pole. However, the active storms entering the Barents Sea are associated with cloud, snow, and temperatures around the freezing point; and 48 hours later the temperature on one of the ice islands had risen to -2°F as a storm approached the pole. An overcast sky with continuous light snow and 22-knot winds were also reported. The Atlantic gulf of warmth shows up well on this chart. It is the major source of warmth and humidity in the central Arctic at this season and is reflected in almost every climatic distribution.

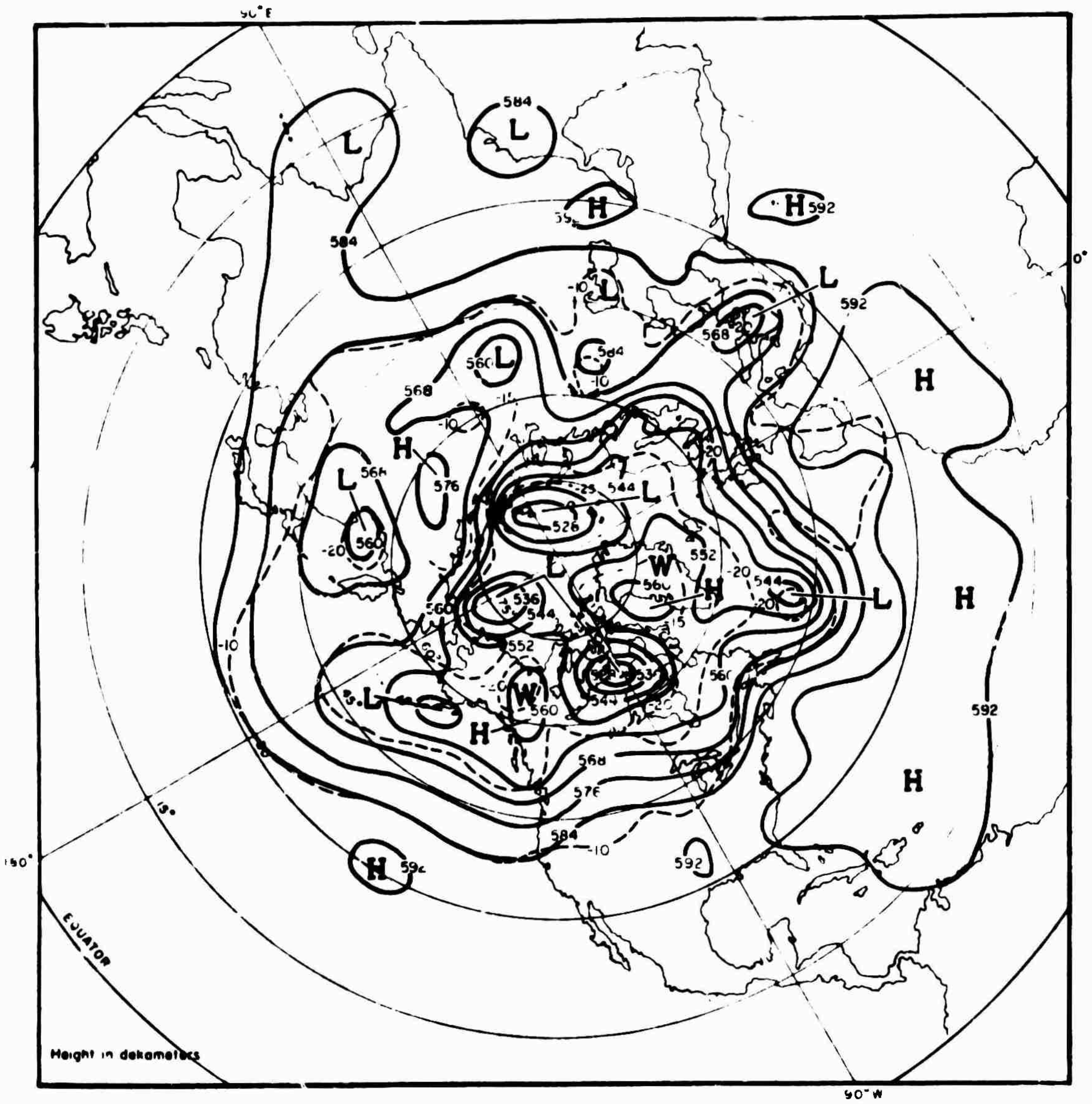
Over North America the eastern areas are now under the influence of northwesterly flow from the Arctic; and the cold, dry conditions of northern Quebec and Labrador contrast sharply with the abnormal warmth and heavy snowfall of January 6. On the other hand, western North America reports temperatures 36 to 54°F higher than on the previous occasion as Pacific storms cross the mountain barrier, reappear as weaker centers, and approach the plains. Over Eurasia surface storms are moving across central Siberia to the arctic coast while the weakening anticyclone is now confined to eastern Siberia. As a result of this southerly flow, the snow, and the clouds, temperatures here are also 36 to 54°F higher over much of the region compared with January 6, while the eastern coast of Asia is experiencing colder weather under the cold northerly flow around the anticyclone cell. Northwestern Europe, in the warm southwesterly airstream, reports temperatures that are, almost without exception, above freezing.

Figures V-2e and f, charts for July 2, 1959, illustrate a summer situation. Here the 500-mb map shows the weaker, contracted vortex of the summer season;



e. SEA LEVEL, 1230 GMT, 2 JULY 1959

Figure V-2 (Cont'd) SELECTED DAILY SYNOPTIC CHARTS FOR SEA LEVEL AND THE 500-mb SURFACE



f. 500-mb level, 1200 GMT, 2 July 1959

Figure V-2 (Cont'd) SELECTED DAILY SYNOPTIC CHARTS FOR SEA LEVEL AND THE 500-mb SURFACE

at the surface the broad-scale patterns are less well-defined and the cyclonic and anticyclonic systems comparatively weak and shifting. Over Siberia the strongest upper westerly flow and steepest temperature gradient are over the northern coastal region while on the North American side this belt lies farther south. A succession of small surface cyclones is moving eastward in this flow to fill-in (eventually) below the shallow vortex centers aloft. At the surface a very great temperature contrast exists between the Arctic Ocean or coastal stations, where temperatures are close to 32°F, and the inland stations. Even at sheltered locations near the coast, temperatures soon rise to over 50°F and some 250 mi inland reach values as high as 77 or even 86°F. The high frequency of frontal storms along this belt in summer may be associated with these strong temperature gradients.

Over North America the 500-mb chart shows weak ridge development over the Davis Strait-Greenland area and the northwest, almost cutting off the cold vortex center over Keewatin. At the surface weak high-pressure systems and a cold low center reflect this upper pattern, while weak frontal systems are beginning to move down over Alaska and the Yukon as the upper flow strengthens in the next 48 hours and the cold low gradually fills. The reported temperatures over this sector are notably lower than those over Eurasia. This is largely a result of the complex interpenetration of land, ice, and open water (with Hudson Bay extending almost to 50°N), together with the fact that a fair proportion of these stations have coastal locations. A further factor of influence in the more southerly regions is the difference in the local time of observatuon. At 12:30 p.m. at the Greenwich meridian it is late afternoon near 90°E but only early morning at 90°W.

The main Atlantic storm track at this season typically crosses southern Scandinavia, while higher pressure is predominant over the Barents Sea. On July 2 temperatures as high as 54°F were recorded on the north coast of Scandinavia, with 52°F on Novaya Zemlya. This results not only from the warm southerly airstream on this occasion but also from the generally northward drift of relatively warm Atlantic waters. In turn, the coolness of the east coast of Greenland is also due to a large extent to the southward drift of the Greenland current. Over the northeastern Pacific the configuration of the coastline is not so favorable to the northward penetration of the warm surface waters into the central Arctic, and temperatures on the north Alaskan coast remain near 32°F.

Drizzle and light rain are reported from place to place in the vicinity of fronts; over the Beaufort Sea and Zemlya Frantsa Iosifa there is continuous snowfall. Overcast skies and fog are widespread over northern coastal areas as a result primarily of the cooling effect on the air of the ice and cold northern waters. Here the local factors of station situation and wind direction play an important role.

Several basic considerations concerning the variability of northern climates are illustrated in the following three examples:

(1) Large-scale patterns tend to persist. The stalled large-scale circulation patterns on January 6, 1959 are characteristic of well-developed meridional situations which can be highly persistent. Over the western hemisphere the broad pattern remained essentially unchanged from early January, with the extension of the Atlantic ridge, to January 24, when there was a shift to the east and storms once again traveled northward over the Atlantic. Regionally these stalled situations can result in extended periods of abnormally cold or mild temperatures, drought or heavier-than-normal snowfall, high

winds, or other extreme conditions. Over the eastern hemisphere there was a change to more zonal flow over Eurasia in the second half of the month, and this pattern tended to persist well into February.

(2) The modification of the tracks and frequency of surface cyclones and anticyclones is importantly influenced not only by relief barriers but also by the dynamic barriers formed by persistent well-developed upper ridges a[illegible]h centers. Areas of most frequent blocking are Scandinavia, Alaska, the Ural region, and the Davis Strait-Greenland sector. Blocking is characteristic of the spring (March) when anticyclonic conditions are most frequent over the continental areas of the high Arctic. At this season the snow and ice cover is still intact, the air is stable, and visibility is good so that flying conditions are generally at their best.

(3) The degree of expansion or contraction of the vortex can have serious effects on arctic operations by causing abnormal displacement of the upper westerlies. In the summer this becomes a major factor over the Arctic and subarctic. In August 1955 the strongest westerly flow lay over the rim of the Arctic Basin and was associated with an unusually high frequency and intensity of cyclones. This had repercussions on the distribution of ice and open water and seriously jeopardized the success of the supply mission to the Distant Early Warning Line sites under construction between Point Barrow, Alaska and Shepherd Bay, Canada. It was not until the polar vortex began to break down about September 5 and storms took a more southerly track across the Bering Sea and Alaska that an offshore lead was created permitting supply operations to continue along the coast of north Alaska. Here the steady offshore flow of cold moist air slowed down the disintegration of the pack ice in August and caused refreezing of the leads by thermal action.

In brief, since the tracks of surface disturbances are so closely related to the upper westerlies, the climate of any season or year, decade or epoch, over the Arctic and subarctic depends to a large extent on the relative frequency of either zonal or meridional conditions, on the most favored location of blocking, and on the degree of contraction or expansion of the vortex. The southern limit of the Arctic fluctuates accordingly. On a regional scale, the local shape, orientation, and intensity of the vortex; the location of individual cells; and the changes in its morphology are basic considerations. There appears to be a tendency for a particular type of large-scale circulation pattern to recur persistently in the same season after only a brief interruption. The nature of the pattern may differ considerably from one season to the same season of another year. Thus the average or normal conditions conceal several well-defined modes of general circulation, each remarkably stable over the Arctic.

The distribution of mean sea level pressure for January (Fig. V-3a) summarizes the winter situation. The major semi-permanent features are the Icelandic and Aleutian lows and the Siberian and northwestern Canadian highs linked by a ridge across the Arctic Ocean. The mean isobars give some indication of the general direction of flow. Figures V-3b and c supplement the mean map by indicating the most frequent areas of formation and stagnation of cyclones and anticyclones and their major tracks.

Over North America most anticyclones form over Alaska and northwest Canada. Cyclogenesis occurs over Alberta where systems from the Pacific produce closed surface lows on the eastern (lee) slope of the Rockies. These cyclones generally move east-southeast to a center of maximum frequency over the Great Lakes region where tracks converge from the southwest and south. Most of these disturbances

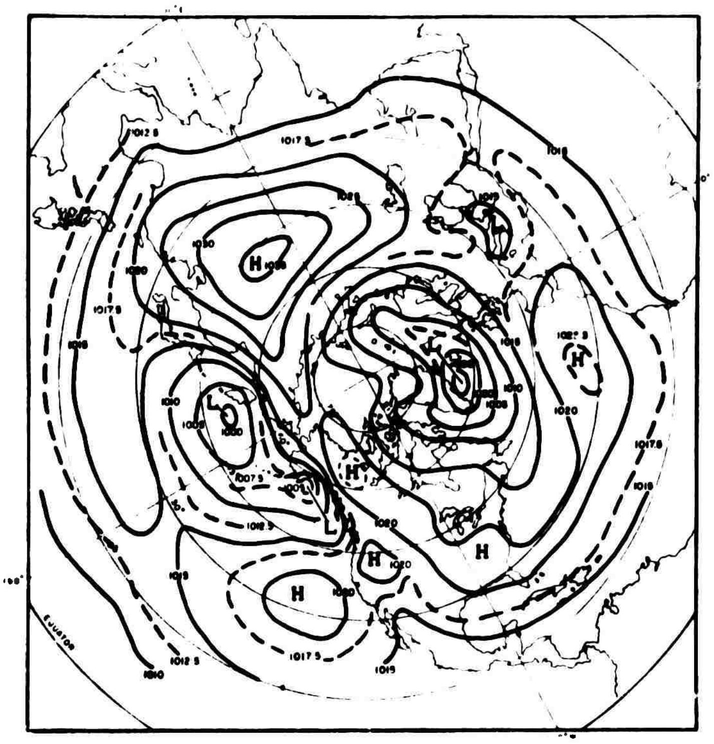


FIGURE V-3a SEA LEVEL PRESSURE (mb), JANUARY 12-YEAR MEAN (O'CONNOR, 1961[21])

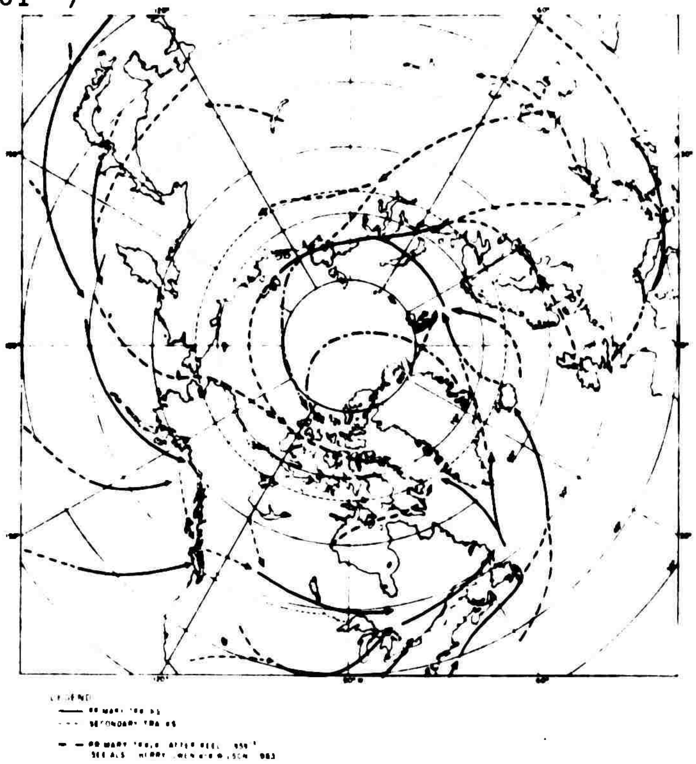


FIGURE V-3b PRINCIPAL TRACKS OF CYCLONES, JANUARY (AFTER KLEIN, 1957[15])

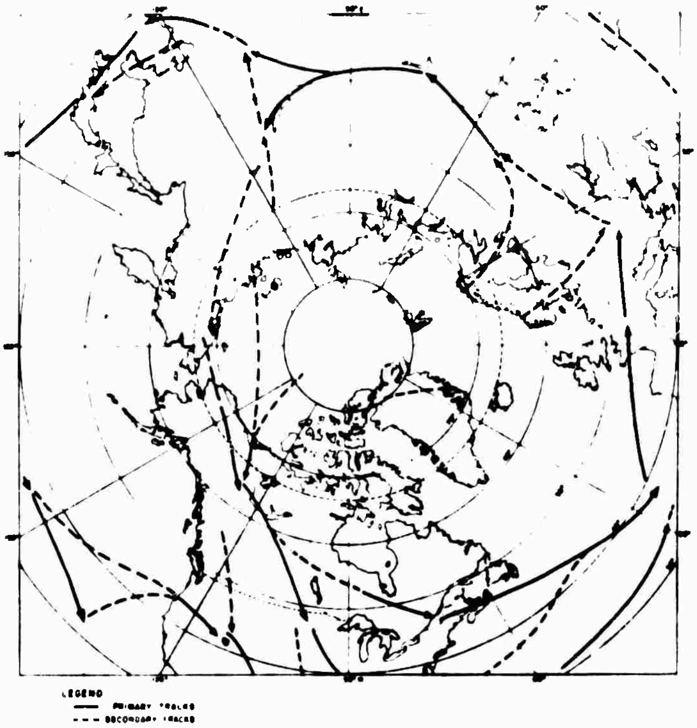


Figure V-3c PRINCIPAL TRACKS OF ANTICYCLONES, JANUARY (AFTER KLEIN, 1957[15]) are then steered northeast to Newfoundland. Here they are joined by cyclones from the east coast, an area which favors cyclogenesis because of thermal contrasts between the cold continental and warmer maritime air masses. From Newfoundland storms generally migrate either toward Iceland (their high frequency resulting in the semi-permanent Iceland low) or along the west or east coasts of Greenland. A few disturbances also enter Davis Strait by way of Hudson Bay or drift in erratically from western Canada below the cold upper vortex center; and an area of high cyclonic frequency is located west of Greenland.

Over Eurasia the westerly jet stream is split into northern and southern branches by the Himalayan massif. To the north of these mountains there is no cyclonic development comparable with that over continental North America, and the winter anticyclone over Siberia is persistent. Cyclones moving eastward over western Siberia are often old occluded storms and rapidly lose their intensity, although secondary cyclones occasionally form near Lake Baikal and deepen rapidly on moving toward the Pacific coast.

Major anticyclonic activity over western Siberia is related to the development and persistence of the strong upper ridge over the Urals. With the collapse of this blocking, the surface centers move to the southeast causing outbreaks of intensely cold air over China and Mongolia. These outbreaks are associated with strong cyclogeneses off the east Asian coast. The majority of these east coast storms are steered northeastward toward the semi-permanent Aleutian low to then stagnate or curve south. Occasionally storms enter the Arctic Ocean over northeastern Siberia or the Bering Strait; those that cross the Alaskan mountains appear to stagnate in less than 48 hours. From time to time new storms develop on the north side of these mountains after primary storms have stagnated to the south. In general, storms over the Beaufort Sea appear to be largely of the cold low type which is characterized by colder air near its center rather than around its periphery.

The daily charts of more recent years indicate that over the Arctic Basin many cyclones spiral in toward the pole and that cyclones outnumber anticyclones by 2 to 1. Storms breaking away from the Icelandic low usually move northeast to the Norwegian-Barents Sea area, and a primary track continues along latitude 75°N where polynyas or leads appear to exist in the Kara and Laptev seas. These frequent storms play an important role in the climate of the north European and Siberian coastlands. This primary track is joined by tracks from the Baltic, Black, and Caspian seas. Some storms stagnate near Novaya Zemlya, others regenerate and curve toward the pole. However, intense anticyclones occur to the north of the Beaufort and Chukchi seas, frequently in association with the extreme northward extension of the upper Pacific ridge. It is in March, at the time of the greatest frequency of blocking situations, that anticyclonic conditions are most frequent over the Arctic Basin.

The most striking pattern to emerge north of 60°N is the predominance of cyclonic activity over the Atlantic hemisphere and of anticyclonic activity over the Pacific hemisphere (the Aleutian low center lies to the south of 60°N). Regions of highest cyclonic frequency extend:

(1) From a maximum south of Greenland northeastward to the north of Norway and Novaya Zemlya, thence spiraling into the central Arctic. It has been suggested that the frequency over the Arctic Basin would be greater if the observation network were closer. These Atlantic storms are responsible for most of the weather and variation in the climatic elements over the Arctic Basin.

(2) Over Davis Strait and Baffin Bay. The role played by the 8,000- to

10,000-ft barrier of the Greenland ice sheet is an important element. There is evidence that the upper manifestations (700 mb) of the surface disturbances and fronts frequently cross the Greenland ice sheet with significant effect on the weather. These upper-level features can often be identified with renewed surface systems on the eastern side of Greenland. Again, when deep lows lie near southern Greenland or Iceland and there is easterly flow across the ice sheet, rapid development or intensification occurs to the west (in the lee) over Baffin Bay. The cyclonic activity of this region is the dominating factor in the weather of much of the Canadian Arctic Archipelago.

Regions of highest anticyclonic frequency extend:

(1) In an arc from eastern Siberia over the East Siberian and Beaufort seas to Alaska and northwest Canada, a reflection of blocking associated with the upper Pacific ridge. The maxima are over the Yukon, the Arctic Ocean, and northeastern Siberia. (The very high pressures frequently mapped over northeastern Siberia are due to an overcorrection in the reduction to sea level, though the anticyclonic frequency maximum may still be real.)

(2) Over Greenland. Again, the centers of very high pressure that frequently appear on sea-level charts are due to the reduction of reported surface pressures to sea level. More than a third of the ice sheet surface is above 850 mb (5,000 ft) and the crest over 700 mb (10,000 ft). The mean charts at 700 mb show no sign of such centers at any season, although there is always a ridge due to the topographic barrier. However, the frequency maximum is real. It is composed of both small, shallow systems lying between predominantly cyclonic regions to the east and west and large systems related to the extension of the main upper Atlantic ridge over Greenland during blocking situations. An analysis over this region of the 12 years from 1947 through 1958 shows a tendency

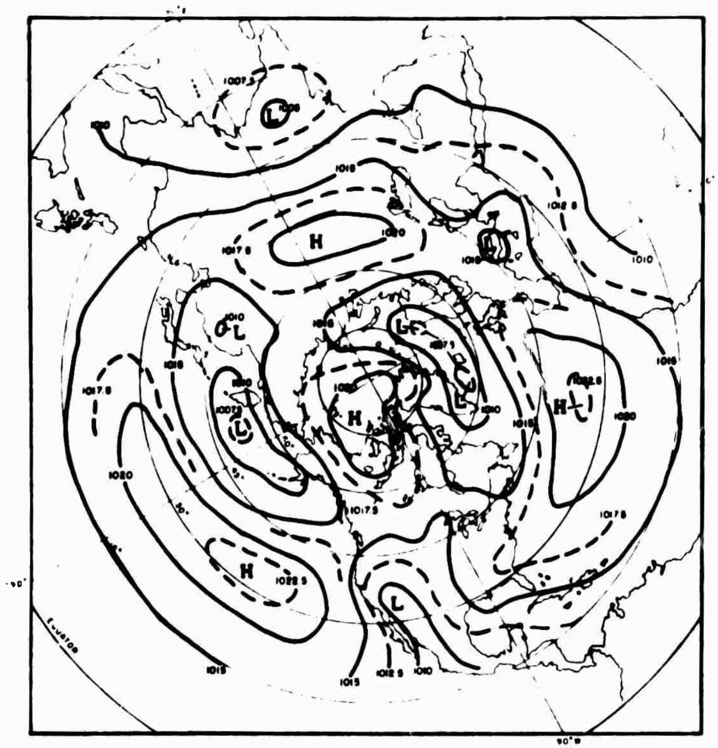


FIGURE V-4a SEA LEVEL PRESSURE (mb), APRIL (12-YEAR MEAN) (O'CONNOR, 1961[21])

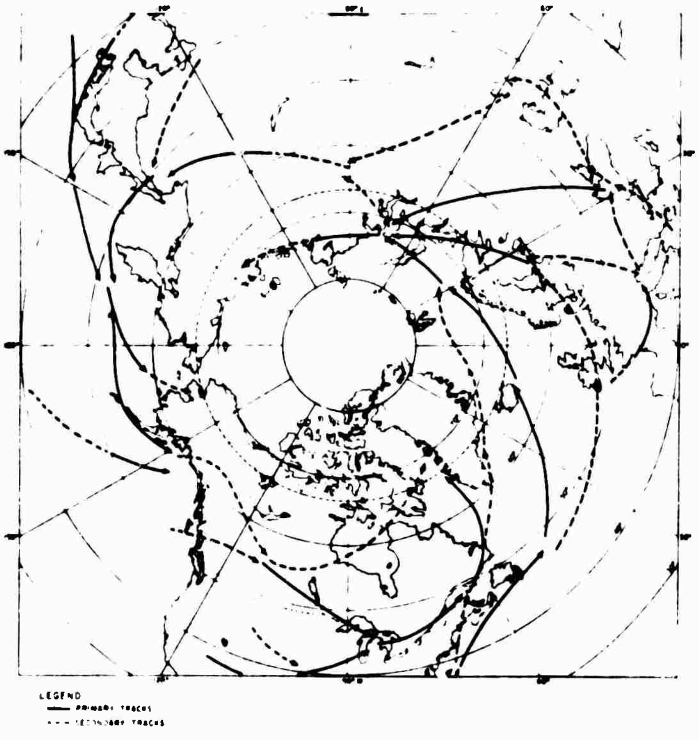


FIGURE V-4b PRINCIPAL TRACKS OF CYCLONES, APRIL (AFTER KLEIN, 1957[15]) toward more frequent blocking than is evident for past years.

(3) Over southern Scandinavia. This is a favored location of the northward extensions of the Atlantic ridge.

(4) Near 90°E at a northern extension of the main west Siberian anticyclone, which generally lies to the south of 60°N. This maximum is linked with the frequent upper-ridge development in the vicinity of the Urals.

A comparison of the mean pressure charts for January and April (Figs. V-3a and 4a) shows the general weakening of the mean systems in spring, the northeast shift in the mean center of the Icelandic low, and the new high-pressure centers which are the dominant feature of the central Arctic at this season. In April the mean 700-mb chart shows a single circumpolar cell, although the axis of the main westerly belt is still far to the south (Fig. V-1b); during May and June, however, it shifts rapidly to the north.

During spring there is a reduction in frequency of the anticyclone tracks over western Canada (Fig. V-4c), while there is a shifting eastward of the primary track toward Hudson Bay. These Hudson Bay highs are characteristic of this season when there is more anticyclonic activity in eastern than in western Canada. Over the arctic portions of North America and Siberia there is a maximum intensification of anticyclonic conditions, while cyclonic activity is at a minimum.

An effect of the increasing insolation is a shifting of the areas of cyclogenesis from water to land regions. In Siberia spring is marked by increasing cyclonic activity which reaches a maximum in May; secondary cyclonic developments over Lake Baikal and northeastern China reach maximum frequency and intensity at this season. The cyclone track over the Arctic Ocean becomes a secondary feature

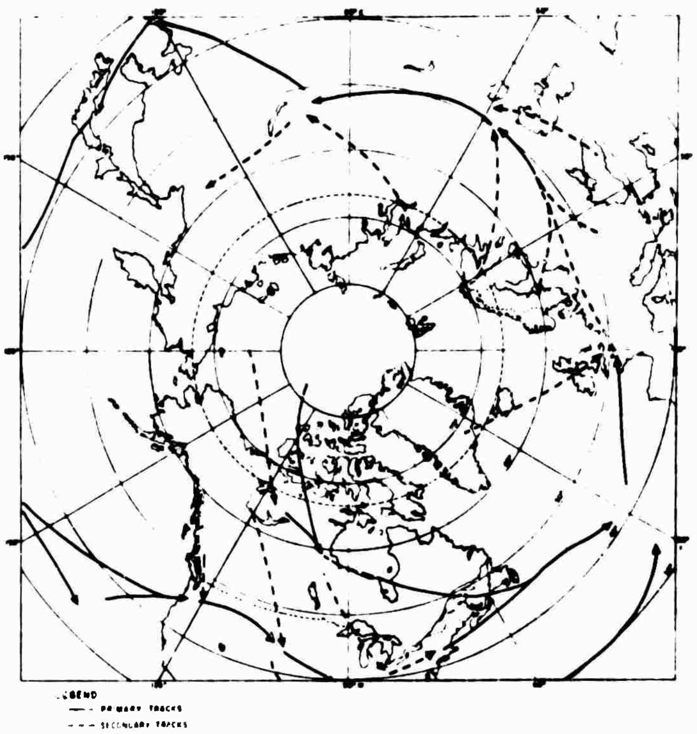


Figure V-4c PRINCIPAL TRACKS OF ANTICYCLONES, APRIL (AFTER KLEIN, 1957[21])

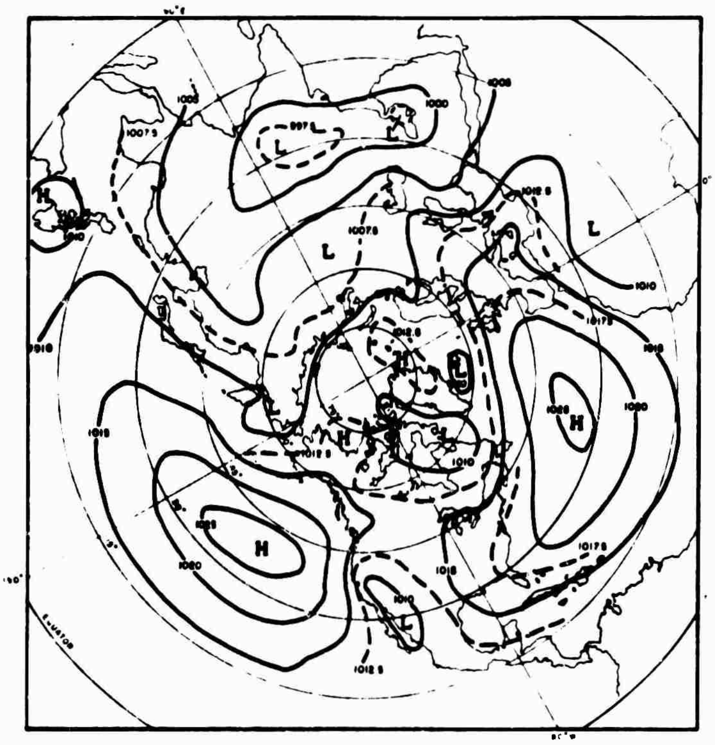


Figure V-5a SEA LEVEL PRESSURE (mb), JULY (12-YEAR MEAN)(O'CONNOR, 1961[21])

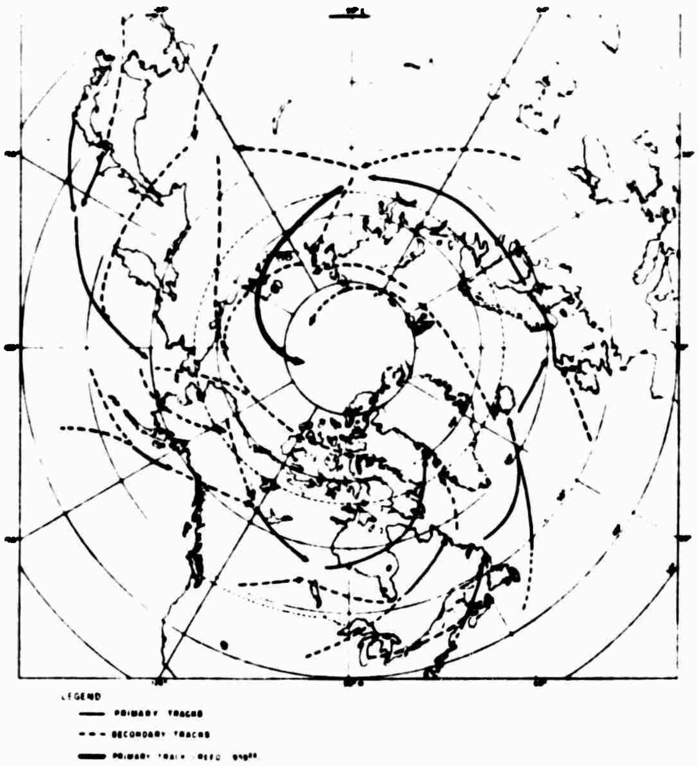


FIGURE V-5b PRINCIPAL TRACKS OF CYCLONES, JULY (AFTER KLEIN, 1957[15])

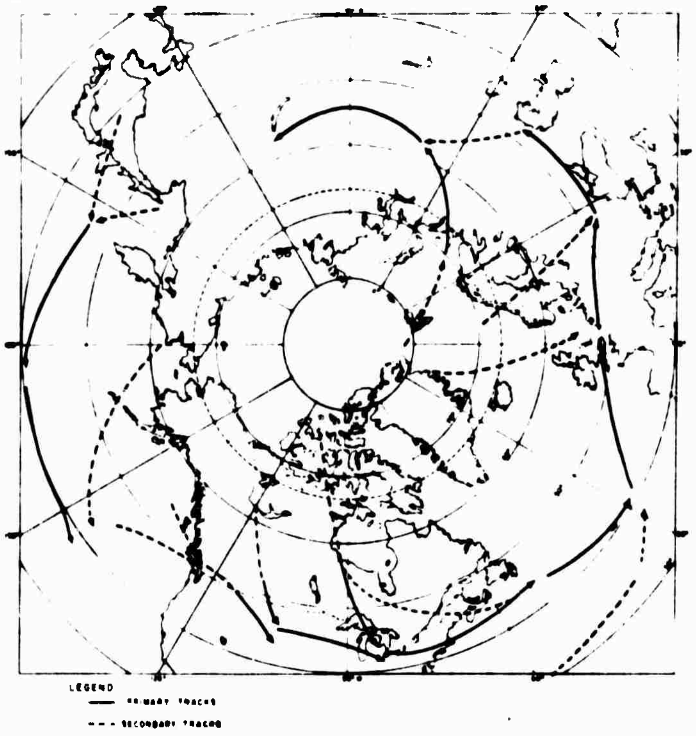


FIGURE V-5c PRINCIPAL TRACKS OF ANTICYCLONES, JULY (AFTER KLEIN, 1957[15]) in May, and primary tracks shift inland over the northeast part of Eurasia. Over North America the cyclone paths shift poleward, to the north of the Great Lakes; and by June cyclonic frequency in Alberta reaches a maximum for the hemisphere.

The mean sea-level pressure chart for July (Fig. V-5a) shows the remarkable fall in mean pressure in summer and the rather amorphous mean pattern which results from the relatively weak, shifting systems over the central Arctic. There appears to be considerable variation in the mean summer pressure from year to year, depending on the frequency and intensity of cyclonic activity. The main characteristics of the mean July pressure chart are the marked east-west extension of the Pacific and Atlantic ridges, the absence of a low center over the Gulf of Alaska, the westward shift of the weak Icelandic center, and the major low center over the heart of Siberia.

A comparison of the January and July mean height charts at 700 mb (Figs. V-1a and c) shows the weakening of the upper westerly circulation in summer and the contraction of the vortex to a single major center over the Arctic Ocean. In July and August the peripheral zone of strong westerlies and the associated surface disturbances are farthest north (Fig. V-5b), and over Asia this zone now lies to the north of the Himalayan massif. The primary cyclone tracks are at approximately 60°N, and a further zone of high frequency occurs between 70 and 80°N.

Over North America the most marked change at this season is the northern displacement of the track of the Alberta low. In summer the Prairie Provinces have one of the highest frequencies of cyclonic activity in the northern hemisphere; most of the storms cross Hudson Bay to the Davis Strait-Baffin Bay area where they stagnate. Over the Atlantic, to the east of the Icelandic graveyard, the primary track is more zonal in July with a drop in the frequency of storms entering the Arctic Basin. A well-defined path now extends from the eastern Atlantic into Siberia at about 60°N.

In the Pacific sector the main track is well to the north of the Aleutians in summer, with storminess reaching an annual maximum over the Bering Sea and Bering Strait. Fewer Atlantic storms enter the Arctic Basin at this season; but when the circulation is more meridional, storms from the Atlantic and the Pacific may penetrate well into the Arctic. There is a high frequency of traveling disturbances along the northern coastal regions of Siberia which appear to be related at least in part to the strong temperature gradients at the margin of the heated land and the melting arctic ice. The storms move mainly to the northeast, spiraling into the central Arctic where they stagnate. Approximately one-third of the lows that enter the Beaufort Sea are from northern Siberia.

The primary anticyclone tracks lie mostly to the south of the subarctic in summer (Fig. V-5c). Anticyclones form to the west of Hudson Bay and move southeast to a region of maximum frequency over the Great Lakes. In August, however, this major track of polar highs shifts westward to the Beaufort Sea-Mackenzie valley area, although some centers still form in Manitoba. Another important primary track in summer extends from the Barents Sea southeast into central Siberia. Anticyclone frequency reaches an annual minimum in eastern Siberia at this time.

The whole region is the scene of considerable cyclonic activity, but maximum cyclonic frequency occurs:

(1) From southern Greenland to Iceland and to the north of Baffin Bay - the graveyards for both continental and east-coast storms. The center west of

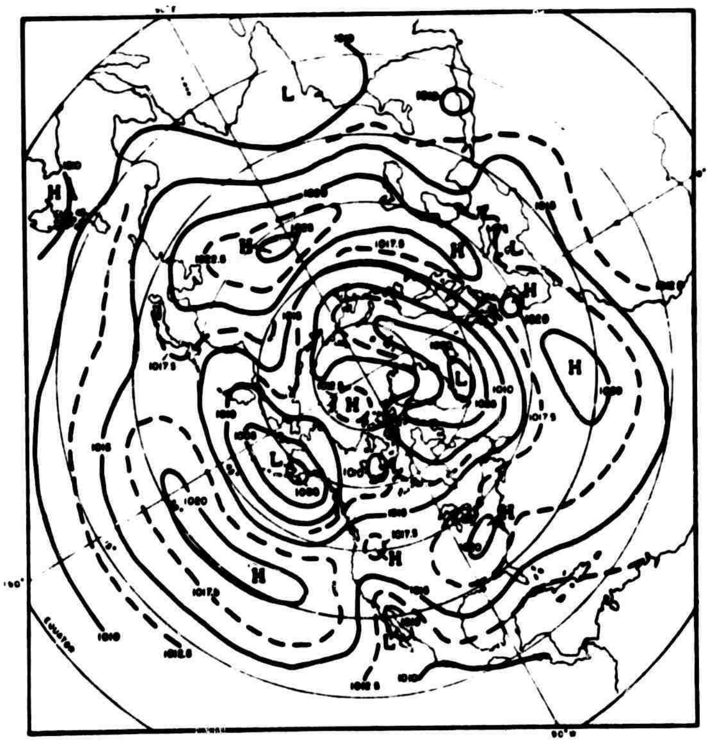


FIGURE V-6a SEA LEVEL PRESSURE (mb), OCTOBER (12-YEAR MEAN) (O'CONNOR, 1961[21])

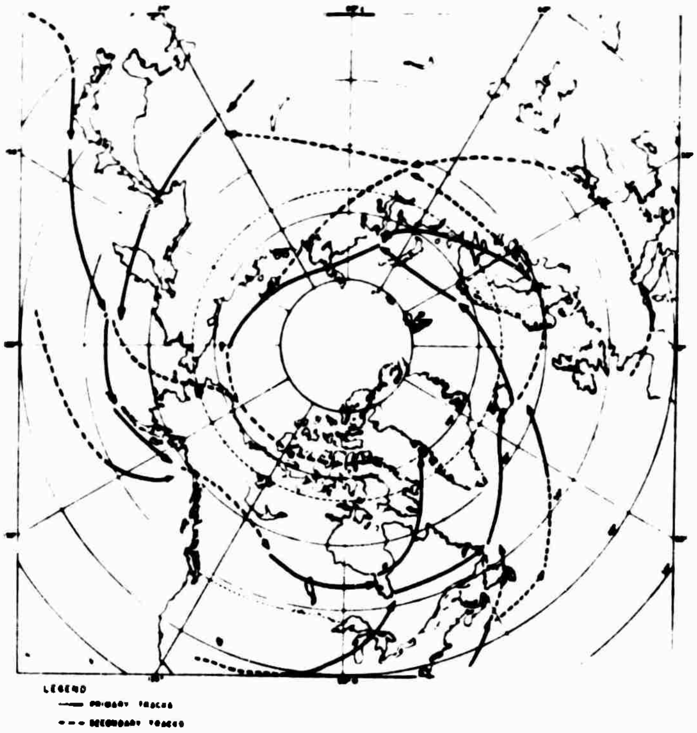


FIGURE V-6b. PRINCIPAL TRACKS OF CYCLONES, OCTOBER (AFTER KLEIN, 1957[15])

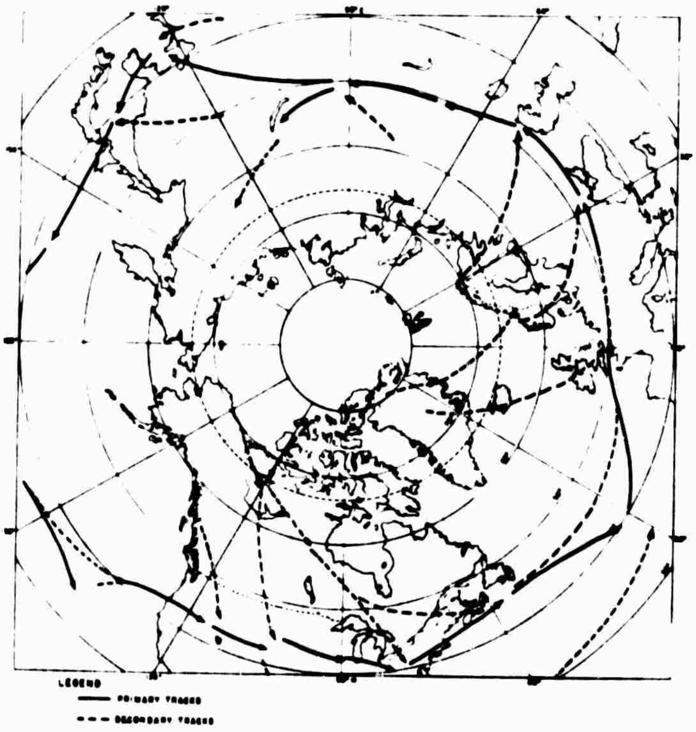


FIGURE V-6c PRINCIPAL TRACKS OF ANTICYCLONES, OCTOBER (AFTER KLEIN, 1957[15])

Greenland is partially explained by the presence of the Greenland ice sheet as a barrier to progress, although the summer records from over the ice sheet reveal that frequent upper air disturbances (700 mb) cross the plateau during the summer as in winter.

(2) Over the Arctic Basin where the high frequency reflects slow-moving, stagnating systems.

Anticyclonic frequency maxima occur over Greenland, the western Canadian Arctic Archipelago, and Pohrosfrov Taimyr - in general, a belt between 70 and 75°N. The decreased frequency of cyclonic centers in this belt is not only due to the tendency for quasi-stable anticyclones to develop, especially over Greenland and near the Beaufort Sea, but also because of the greater mobility of the traveling cyclones in this zone.

A spot-check of the frequency of cyclones during the period from 1952 to 1956 against earlier expedition data for 1894-5, 1904-5 and 1923-4 indicated no significant difference in the number of cyclones. Therefore, these results are probably representative of a longer period of time.

The autumn is marked by the intensification and southward expansion of the upper circumpolar vortex, and the mean October maps both for the 700-mb (Fig. V-1d) and surface levels (Fig. V-6a) already show many of the major features of winter. Following this southern shift in the zone of maximum upper westerlies, the primary storm tracks move south. By November the belt of strongest westerly flow is again in the latitude of the Himalayan massif and the major winter features of the circulation are set up and held during the following 5 or 6 months.

A second feature of autumn is the relative warmth of open water bodies compared with the rapidly cooling land. The October track charts (Figs. V-6b and c) show areas of maximum cyclonic activity over the Sea of Okhotsk, the Barents Sea, the Gulf of Alaska, James Bay, and the Kara Sea. Primary cyclones enter the Arctic Basin from the Atlantic and the Baltic Sea, and the frequency reaches an annual maximum over the Kara Sea from which a primary track follows the coastal margin. In November primary storm tracks from Alberta and Colorado again merge in a center of maximum frequency over the Great Lakes. Over the Atlantic the primary track shifts from the east to the west side of Iceland, and the southern Davis Strait track reappears and intensifies.

During this season anticyclonic frequency increases over eastern Asia, and centers of high pressure appear at sea level over Siberia. Over North America the trajectory of polar highs over western Canada becomes of primary importance by November and remains so until March.

B.5.3 Surface Weather

The surface weather associated with anticyclones and cyclones is profoundly affected by the high degree of static stability of the lowest layer of the atmosphere, caused by the radiational cooling over the snow and ice cover. The surface weather over much of the Arctic and subarctic depends on the local effectiveness of anticyclonic and cyclonic activity in either intensifying or breaking down this very cold stable layer.

In general, the calm clear conditions of the central regions of large persistent anticyclones favor the intensification of the inversion, both by heating and drying the free atmosphere through subsidence, which characterizes these systems, and by not hindering outgoing radiation from the surface. Consequently very low surface temperatures occur and are maintained. The normally excellent visibility can be marred by local ice fog around settlements if supersaturated air at about -40°F is present, and the sharp density gradient in the inversion layer is often responsible for such optical phenomena as mirages. Where leads exist in the ice, the cold air spreading over the warm surface of the water gives rise to arctic smoke or steam fog.

Most of the various surface weather factors - cloud cover, precipitation, winds, and temperature change - are associated with the presence and passage of cyclonic disturbances. Besides the strong winds that frequently accompany the passage of cyclones, the cloud cover and precipitation effect a rise in the surface temperature through longwave radiation. This rise tends to be slow, but persistently overcast skies together with warm air advection can lead to extended periods of above-normal surface temperatures.

The weather associated with these cyclonic systems is modified, especially in winter, by the fact that air temperatures are generally so low that the moisture content even at saturation point is too little to give a dense cloud cover and heavy precipitation. Locally the weather resulting from these storms is influenced by their recent passage over land, water, or ice surfaces. Most of the cyclones originate in more southerly latitudes and by the time they reach the Arctic have lost most of their strong temperature contrast and their frontal structure at the surface with the associated high, middle, and low cloud sequences. They are generally weaker and occluded with the warm air lifted to considerable heights. When fronts do exist they are often ill-defined with no extensive cloud systems or precipitation. Although the cyclonic circulation at the surface may be weak or absent, the sky is generally overcast and occasionally ice needles or traces of snow are formed.

With the general breakup of the ice and the disappearance of the snow cover, the surface inversion weakens and is confined to a shallow layer over the pack ice and ice caps. At this season, the differential heating of the land, water, and ice and the local modification of the temperature profile and moisture content of the moving air masses become a prime factor affecting surface weather over the arctic and subarctic regions.

Over the arctic pack and coastal waters, where the presence of ice holds the surface temperatures near 32°F, the high frequency of fogs and heavy low stratus cloud in the 0 to 5,000-ft layer is closely linked with the inversion itself and to a large extent independent of the type of circulation system. Condensation occurs when warmer air flows over and is cooled by the cold waters or watery ice surface. If the windspeed is low, fog forms close to the surface; if the wind is blowing strongly enough, the fog is lifted to form a low ceiling of stratus cloud. When the wind direction is favorable the fog and cloud drift in over the coastal regions, and hence conditions over coastal airstrips can change rapidly. Drizzle and snow frequently fall from the stratus layer, and temperatures remain low beneath this layer of cloud. Wherever cloud is present within the temperature range of -22 to 32°F there is the possibility of aircraft icing. Moderate to heavy icing has been encountered where cold air passes over open water forming cloud with a large concentration of supercooled droplets; usually this type of cloud does not extend above about 4,000 ft. Inland the inversion and the stratus layer soon disappear over the heated land surfaces.

It is the frequent cyclonic activity over the central Arctic in summer which is responsible for the middle and high clouds, the storms, and much of the precipitation. Cyclones with very great temperature contrasts, strong winds, well-defined frontal zones, extensive cloud systems, and precipitation zones occurred over North Pole IV during the summer of 1955. The vigor of these storms is partly related to the strong thermal contrasts that exist near the arctic coast at this season. In frontal cyclones over the Arctic Basin, the relationship between cloud amount and type is probably similar to that in mid-latitudes, where the large-scale upgliding motion of the warm air produces the sequences of cirrus and cirrostratus, lowering to altostratus and altocumulus, and finally to nimbostratus with the onset of precipitation. However, the different cloud types often lie in discrete overlapping strata rather than in a single sloping layer. The largest amount of cloud occurs in the vicinity of the occlusion, which may extend to over 20,000 ft. Strong cumulus development and thunderstorms are rare over the central Arctic.

Once the snow and ice have cleared from the continental subarctic, surface temperatures generally rise rapidly; however, heavy cloud cover and increased precipitation in areas of high cyclonic frequency can do much to delay the temperature rise and the drying out of the surface. Labrador, for example, has a lower cyclone frequency and a correspondingly higher number of sunshine hours. The amount of weather associated with cyclonic activity over the subarctic will again vary according to the temperature contrasts and moisture content of the air masses involved. These in turn are influenced by surface conditions, from the strong contrasts over land and ocean to the lesser variations over local water bodies such as the lakes in the Mackenzie valley.

Under an anticyclonic regime the long hours of daylight can produce warm clear weather with inland temperatures north of the Arctic Circle rising above 70 or even 85°F. With continued heating of the surface, convection leads to cumulus development and thunderstorms.

Over both the central Arctic and much of the continental subarctic the precipitation maximum occurs in the warmer half of the year and is associated with the passage of frontal cyclones. The low stratus give some additional precipitation over the basin, and convective showers occur south of 75°N where continental heating is strong. The amount and type of precipitation depend on the temperature and moisture content of the air. Rain can fall anywhere in the Arctic in summer, and it has been suggested that nearly 50% of the annual total precipitation over the Canadian Arctic Archipelago may fall as rain. Where temperatures are in the vicinity of 32°F, it is not easy to predict a type of precipitation without a detailed knowledge of local conditions which are often unknown. Furthermore, since the greater part of the high Arctic is virtually a desert, a single storm can represent a high percentage of the total annual precipitation. At Thule on July 22, 1957, 1.9 in - almost the whole season's rainfall - fell in one day. There is, therefore, considerable variability in the amount from year to year. Local terrain also ensures considerable variation from place to place in the amount of precipitation received from any one storm.

Over coastal regions, the bulk of the precipitation frequently falls at the end of the warm season when the land surfaces are cooling more rapidly than the open water. The latter serves as a heat, as well as moisture, source in modifying the air masses.

B.5.4 Wind

Surface wind velocities usually are not very high over the Arctic Ocean. On the average the force is between 8 and 10 knots. When an inversion is present the surface layer is effectively isolated from the faster-moving air above. It is this fact, in combination with the lack of topographic effects, which results in the low number of occurrences of strong winds.

The annual mean wind speeds are greatest at exposed coastal stations near cyclone tracks, and it is in these same locations that storms are most frequent (Jan Mayen, northern Norway to Dikson Island, Bering Strait). A study of wind conditions over the central Arctic Ocean showed that the layer of the atmosphere from the surface to 1,600 ft is characterized by a minimum interdiurnal variability of the wind speed near the ice surface (4 to 6 kt), and that the seasonal variability of wind speed is slight.

The interdiurnal variability of wind direction was found to be greatest near the ice surface (50 to 70 ft) and less at greater heights. This is explained by the low surface wind speed and the consequent instability of the wind direction. Observations from drifting ice stations have been used to prepare the following table of surface wind speed.

TABLE 5-I

Frequency Distribution of Wind Speed Over Central Arctic Ocean (Percent)

Knots	0	2	4	6	8	10	12	14	16	18	20-28	29-37	38	No. of obs.
J	11	7	8	10	13	12	8	8	6	4	10	2	1	664
F	10	6	11	15	17	15	9	5	5	1	6	0	0	548
M	6	6	8	18	22	15	7	6	4	2	5	1	0	585
A	6	5	15	15	17	15	8	7	5	4	3	0	0	479
M	7	5	11	16	16	15	11	7	6	3	3	0	0	744
J	5	5	9	15	13	15	11	9	6	4	7	1	0	669
J	4	3	9	12	13	16	10	9	9	6	8	1	0	609
A	4	4	7	11	11	15	11	11	8	5	11	2	0	570
S	8	4	9	13	15	15	8	10	5	4	7	2	0	545
O	7	5	9	12	12	12	10	9	7	4	11	2	0	586
N	9	7	10	13	16	15	7	6	6	4	6	1	0	607
D	11	10	14	14	17	12	6	5	4	3	4	0	0	607

There are certain regional characteristics in the arctic wind systems over land. Eastern Siberia, beyond the Verkhoyansk Mountains, experiences a monsoonal change in wind direction. During winter southerly winds prevail, being strongest over the coast and light or calm in the interior. At Verkhoyansk there is practically no surface wind from February to May. The frequency of storms there is greatest during the months of June, July, and August, with an average of two or three per month. During these months, the thermal low-pressure system shows greatest development over the interior. Along the coast the frequency of storms is least during the same summer months, Tiksi Bay having an average of one per month from April through September. During the colder months, when the anticyclone is well-developed over the continent, there are no gales in the interior at Verkhoyansk; but during these same months at Tiksi Bay the frequency is greatest. From October through May there is an average of four storms each month.

The Bering Strait and the land areas immediately to the east and west of it, being low in elevation, present no obstacle to the free interchange of air between the Arctic Basin and the region of the Bering Sea. During the winter the prevailing wind through this region is northerly, and this flow of cold air replenishes the air moving from the Bering Sea into the Pacific Ocean. During this season the Aleutian low is well-pronounced, and the surface winds and upper winds at Nome, Alaska confirm this. Sixty percent of the surface winds at Nome are from the north to northeast at about 9 knots, and continue from that quadrant up to heights of 6,000 ft.

The wind regimen of Alaska, like that of eastern Siberia, is to some extent monsoonal in character. In the warmest months, a thermal low develops over the interior of Alaska whereas, in winter, high pressures are prevalent. As a result, winter winds at Fairbanks are regulated by the polar high-pressure ridge and by the proximity of the Aleutian low-pressure system to the south. Surface winds at Fairbanks itself are light because the town is located in a sheltered valley in which the average winter surface wind speed is about 4 knots. Immediately above the surface 71% of the winds are east-northeast to east-southeast; but with increasing height the wind veers, and at 10,000 ft 43% are from the south to the west-southwest. In summer the air pressure increases over the Gulf of Alaska, and a local thermal low is established over the interior. The prevailing winds at Fairbanks are southwesterly at about 5 knots.

The frequency of storms in Alaska is similar to that of eastern Siberia. In the interior the season of greatest frequency is during the summer, whereas along the coast it is during the winter. Point Barrow, on the north coast, has an average of 19 storms per year whereas Fairbanks in the interior has only 1. Wrangel Island has an annual average of 50 storms, and from May through August there is an average of 2 each month. Summer is the season of low storm average at Wrangel Island, at which time the Aleutian low is dissipated. During the colder months, between September and April, the average monthly frequency is between 5 and 6. At this season the Aleutian low is well-developed.

The maps of air pressure show that a north and northwesterly air flow is directed over the Canadian Arctic Archipelago throughout most most of the year. The flow is strongest during winter. Winds across the Hudson Bay region are northwesterly in winter and only disturbed in summer by the passage of individual storms. Such cyclones invade the region and give frequent summer storms.

In central Greenland, unlike the interior of Eurasia and North America, winter winds are strongest. The temperature inversion persists through the year in Greenland, and the pressure gradient controls the annual wind cycle.

At several arctic sites, the high surface winds associated with strong pressure gradients are often enhanced by local effects. Many stations show locally preferred wind directions due to topography, and surface wind data from

Table 5-II MEAN WINDSPEED (mph) AND DIRECTION, NORTHERN HEMISPHERE

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Year*
Canada (1951-60)[11] [160]													
Aklavik													
mean speed	6.0	5.6	6.5	7.4	7.3	7.8	7.0	7.0	7.0	6.2	5.2	5.7	6.6
prevailing direction	S	S	NW	N	N	N	NW	NW	NW	NW	NW	NW	
Alert													
Mean speed	5.0	5.1	4.6	4.9	5.1	6.5	7.6	6.0	6.4	6.9	5.8	4.6	5.7
prevailing direction	W	W	W	W	WNW	NE	NE	NE	W	W	W	W	
Baker Lake													
mean speed	14.6	14.1	13.5	14.1	14.2	12.0	11.3	12.7	13.6	15.0	14.6	14.8	13.7
prevailing direction	NW	NW	N	N	N	N	N	N	NW	N	N	N	
Cambridge Bay													
mean speed	12.3	10.7	10.6	12.3	12.7	12.9	12.9	12.8	13.4	14.0	12.0	10.8	12.3
prevailing direction	W	W	W	N,NW	NW	N	N	E	E,NW	NW	W	W	
Chesterfield													
mean speed	15.3	14.2	12.9	13.2	14.9	11.8	11.7	13.0	15.2	17.1	14.9	15.6	14.2
prevailing direction	N	N	N	N	N	N	N	N	N	N	NW	N	
Clyde*													
mean speed	4.6	7.4	4.9	4.7	6.4	8.0	8.5	6.4	8.1	10.3	7.0	3.8	6.7
prevailing direction	NW	NW	NW	NW	NW	NW	NW	NW	NW	NW	NW	NW	
Coppermine													
mean speed	12.2	10:5	9.2	8.7	8.5	8.5	9.6	10.0	11.0	12.0	11.0	10.3	10.1
prevailing direction	SW	W	SW	W	W	N	NE	NE	N	SW	W	SW	
Coral Harbour													
mean speed	12.1	12.3	10.5	13.1	13.2	12.3	12.2	12.9	13.3	13.3	13.2	13.3	12.6
prevailing direction	NW	N	N	N	NW	N	N	N	N	NW	N	N	
Eureka													
mean speed	7.2	6.6	5.4	5.8	8.4	10.9	11.3	9.6	7.8	6.3	6.2	5.4	7.6
prevailing direction	E	E	E	E	NW	NW	NW	NW	NE	E	E	E	
Frobisher Bay													
mean speed	9.1	9.5	9.7	11.0	13.4	11.7	9.7	8.9	11.5	14.6	12.3	11.0	11.0
prevailing direction	NW	NW	NW	NW	NW	SE	SE	SE	NW	NW	NW	NW	
Holman													
mean speed	9.1	7.9	9.5	11.5	10.3	9.5	8.7	8.9	11.7	13.1	11.6	11.2	10.3
prevailing direction	E	E	E	E	E	E	W	E	E	E	E	E	
Isachsen													
mean speed	10.6	7.9	7.0	7.5	10.3	9.9	10.9	10.0	9.9	11.0	9.0	9.7	9.5
prevailing direction	N	N	N	N	N	N	NW	N,SW	N	N	N	N,NW	
Mould Bay													
mean speed	10.5	8.6	7.9	8.4	11.7	13.0	12.2	11.2	11.6	11.3	9.9	8.5	12.5
prevailing direction	NW	NW	N	N,NW	NW	NW	NW	S,NE	NW	NW	NW	NW	
Nottingham I.													
mean speed	10.9	11.5	9 9	11.5	11.2	10.7	9.8	11.1	10.9	14.5	13.0	11.6	11.4
prevailing direction	NW	NW	NW	NE	NE	NE	W	NE	NW	NW	N	N	

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* 2 years only.

Table 5-II MEAN WINDSPEED (mph) AND DIRECTION, NORTHERN HEMISPHERE

	J	F	M	A	M	J	J	A	S	O	N	D	Year
							Canada (cont'd)						
Resolute													
mean speed	11.9	11.5	10.4	10.9	11.6	12.6	12.1	12.3	12.5	12.5	11.0	10.4	11.6
prevailing direction	NW	E	NW	NW	NW	NW	NW	NW	NW	NW	NW	NW	
Resolution I.													
mean spead	20.5	20.5	16.1	16.2	14.4	13.2	12.5	13.5	13.9	18.2	17.9	20.4	16.4
prevailing direction	W	SW	W	NE	W	NE	E	E	W	W	W	W	
Sachs Harbour													
mean speed	13.2	11.6	10.7	13.2	12.8	12.8	13.1	13.6	14.7	15.2	13.0	12.0	13.0
prevailing direction	N	E	SE	E,SE	E	N,E	NW	SE	E	E	E	E	
Knob Lake*													
mean speed	10.1	10.6	10.6	10.0	10.6	10.3	10.2	10.7	12.7	12.5	12.1	11.8	11.0
prevailing direction	NW	NW	NW	NW	NW	SE	NW	NW	NW	NW	NW	NW	
Churchill†													
mean speed	14.0	14.2	13.9	14.4	13.1	12.0	12.3	12.6	14.8	16.1	15.1	16.0	14.0
prevailing direction	NW	NW	NW	NW	N	N	N	NW	N	NW	NW	NW	
Fort Simpson†													
mean speed	6.6	7.8	7.9	8.1	8.0	7.5	7.1	6.8	8.1	7.2	7.8	6.6	7.5
prevailing direction	NW	NW	NW	NW	NW	NW	NW	SE	SE	SE	NW	NW	
Yellowknife†													
mean speed	8.0	9.2	10.7	11.6	10.5	10.4	10.5	10.2	10.4	11.7	9.3	8.6	10.1
prevailing direction	N	E	E	N	E	NE	N	SE	N	E	E	E	
Fort Nelson†													
mean speed	3.9	4.7	5.6	6.5	6.4	6.1	5.6	5.4	5.2	4.8	4.0	3.5	5.1
prevailing direction	S	N	N	N	N	N	S	S	S	S	S	S	
Whitehorse†													
mean speed	8.6	8.8	9.1	8.7	8.7	8.0	7.4	7.8	9.1	10.4	9.0	8.7	8.7
prevailing direction	S	S	S	S	SE	SE	SE	SE	S,SE	S	S	S	
							Alaska[152]						
Anchorage													
mean speed (31)**	5.2	5.9	5.8	5.7	6.4	6.2	5.6	5.2	5.2	5.3	5.1	4.9	5.5
prevailing direction (8)	NE	N	N	N	S	S	S	NW	NNE	N	N	NE	N
fastest mile (10)	60	62	49	66	31	33	32	45	49	59	66	56	66
direction (10)	NE	S	N	S	S	S	S	S	SE	S	NE	SW	NE
Barrow													
mean speed (30)	11.0	11.3	10.9	11.5	11.8	11.4	11.8	12.7	13.7	14.0	12.5	10.9	12.0
prevailing direction (7)	ESE	ENE	NE	E	NE	E	SW	E	ENE	NE	NE	ENE	NE
fastest mile (30)	56	51	48	52	43	38	41	47	56	51	63	70	70
direction (9)	E	SW			SW	W		E	W		W	W	W
Barter I.													
mean speed (7)	14.7	15.2	11.7	12.7	11.9	11.0	10.3	11.0	12.8	14.7	14.9	11.9	12.7
prevailing direction (7)	W	W	W	ENE	E	ENE	ENE	E	E	E	ENE	ENE	ENE

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* Average values 1954-62:[127] prevailing direction.[6]

† Periods prior to 1951-60.[6]

** Number of years of record.

Table 5-II MEAN WINDSPEED (mph) AND DIRECTION, NORTHERN HEMISPHERE

	J	F	M	A	M	J	J	A	S	O	N	D	Year
Alaska (cont'd)													
Fairbanks													
mean speed (27)	3.2	3.8	4.7	5.9	6.8	6.4	5.8	5.7	5.5	4.9	3.7	3.0	5.0
prevailing direction (8)	N	N	N	N	N	SW	SW	N	N	N	N	N	N
fastest mile (11)	41	57	60	35	42	40	50	34	49	50	50	34	60
direction (11)	SW	W	SW	SW	SE	SW	SW	W	SW	SW	SW	E	SW
Kotzebue													
mean speed (13)	15.3	14.3	13.2	13.4	10.3	12.2	12.9	13.9	13.0	13.6	13.9	12.9	13.3
prevailing direction (14)	E	E	E	ESE	W	W	W	W	ESE	NE	ESE	NE	W
McGrath													
mean speed (8)	2.7	3.5	4.0	5.0	5.6	5.7	5.4	5.1	4.9	3.8	2.5	2.4	4.2
prevailing direction (9)	NW	NW	NW	N	E	S	S	S	N	N	ESE	NW	NW
Nome													
mean speed (10)	12.0	11.1	11.2	10.9	10.0	9.5	9.9	10.7	11.4	10.9	11.6	9.8	10.8
prevailing direction (9)	E	ENE	E	ENE	NE	SW	WSW	SW	N	NE	N	E	E
fastest mile (10)	75	74	72	53	68	40	38	50	57	57	73	72	75
direction (10)	E	W	NE	S	NE	E	SE	SW	E	SE	SW	E	E
Greenland[3] [57]													
Thule													
mean speed (11)	9	9	7	7	7	7	7	6	8	10	9	8	8
prevailing direction (11)	E	E	E	E	E.W	W	W	W	E	E	E	E	E
Eismitte													
mean speed (2)	11	9	13	12	9	9	9	8	11	10	9	14	10
Angmagssalik													
mean speed (30)	6	6	5	3	3	3	3	3	3	4	5	5	4
Scoresbysund													
mean speed (12)	5	5	4	4	3	3	3	3	3	4	4	4	4
Nanortalik													
mean speed (41)	12	11	11	10	7	10	7	6	8	9	10	10	10
Eurasia[98]													
Jan Mayen													
mean speed (mph)	21	21	18	17	13	14	12	14	16	18	19	18	17
no. of days with gales*													
Vardo													
mean speed (28)	22	21	21	19	16	16	13	14	17	19	21	21	18
no. of days with gales*(28)	5	5	4	3	2	1	1	2	2	3	5	5	38
Ostrov Diksona													
mean speed (19)	19	19	16	17	16	16	15	16	17	16	18	17	17
no. of days with gales*(19)	11	19	8	7	5	3	1	3	5	6	9	8	75
Bukhta Tikhaya													
mean speed (10)	16	15	12	12	11	10	8	9	13	15	15	17	13
no. of days with gales*(10)	10	9	7	5	3	3	2	2	4	6	8	11	69

• Gale defined as: windspeed $\geq$ 32 mph.

Table 5-II MEAN WINDSPEED (mph) AND DIRECTION, NORTHERN HEMISPHERE

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Year*
Eurasia (cont'd)													
Russkoye Ust'ye													
mean speed (9)	8	8	9	9	9	12	12	12	11	9	9	9	10
no. of days with gales*(9)	<0.5	<0.5	2	1	1	2	2	2	1	2	1	1	15
Mys Shmidta													
mean speed (3)	13	11	12	11	11	11	10	13	14	17	14	12	12
no. of days with gales*(3)	10	6	8	5	5	4	2	4	5	7	4	8	68
Anadyr'													
mean speed (14)	13	14	12	11	9	10	11	11	11	12	13	13	12
no. of days with gales*(14)	4	4	4	2	1	< 0.5	2	1	2	2	4	3	29
Verkhoyansk													
mean speed (21)	1	2	2	4	6	7	6	5	4	3	2	1	3
no. of days with gales*(21)	<0.5	<0.5	<0.5	<0.5	2	2	1	1	<0.5	<0.5	0	7	17
Yakutsk													
mean speed (21)	3	3	3	5	6	5	5	5	4	4	3	3	4
no. of days with gales*(21)	<0.5	<0.5	1	1	2	2	1	1	1	1	<0.5	<0.5	10
Polar Ocean													
"Sedov"													
mean speed (1)	11	9	10	11	13	14	12	11	13	14	14		
"Fram"													
mean speed (1)	11	9	8	8	11	12	12	11	11	10	9	10	10

Ice Caps[59]
(summer)

	Upper Ice I 1960	*Penny Ice Cap 1953*	*Barnes Ice Cap 1950*	*Frøys Glacier 1939*	*Isachsen Plateau 1934*
Height above surface (cm)	200	183	213	200	200
Mean speed (m/sec)†	5.1 (11.2)	4.4 (9.8)	4.3 (9.6)	2.8 (6.2)	2.7 (6.0)
Maximum speed (m/sec)†	20.0 (44.5)	16.1 (35.9)	16.5 (36.8)	6.2 (13.8)	8.1 (18.0)

* Gale defined as windspeed ≥ 32 mph.

† Mph in parentheses.

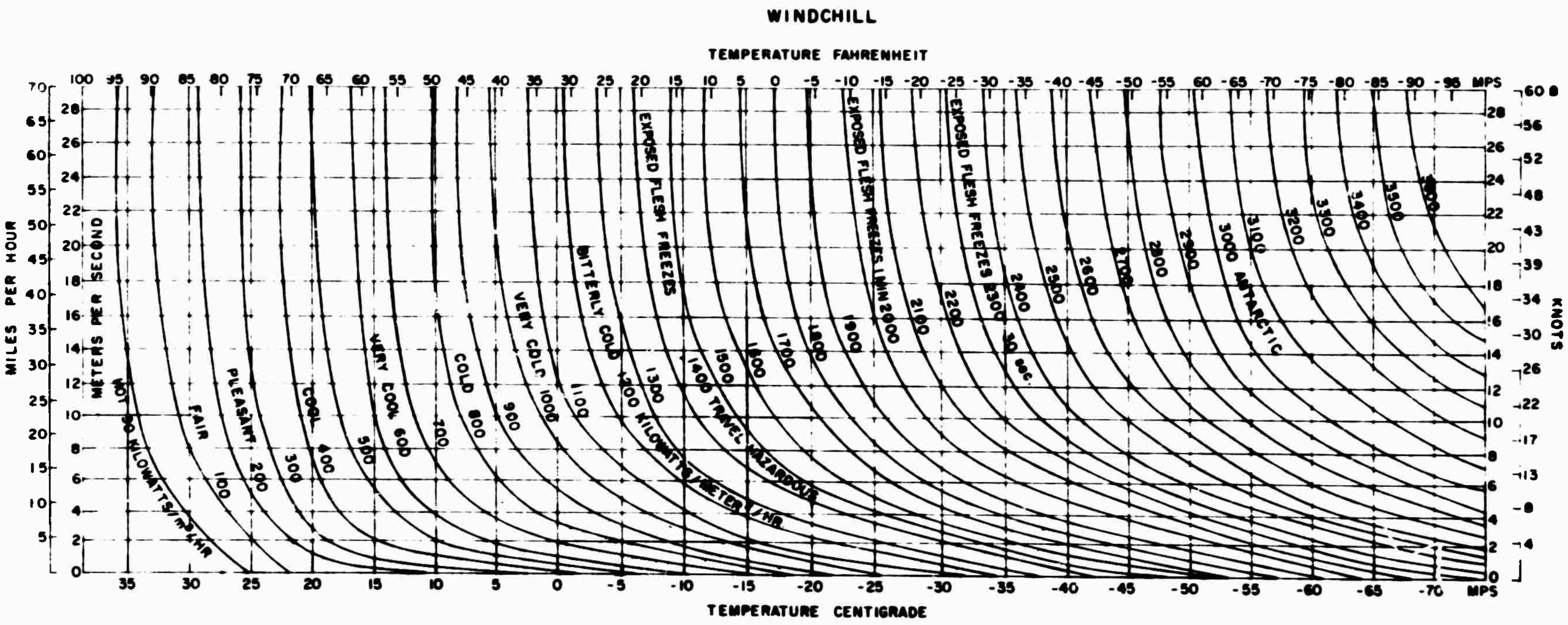
arctic land stations must not be taken as typical of the wind speed and direction over a larger area. Certain locations in Greenland are good examples of this - Thule in the northwest and Narssarssuak in the southwest both have local strong winds partly associated with descent of the sir from the inland ice sheet. Alert, Barter Island, and Juneau all experience local strong winds. Almost any fiord in the Arctic may have peculiar local wind conditions.

B.5.5 Windchill

It must be remembered that in cold climates the combination of temperature and wind produces the greatest heat loss and discomfort (Fig. V-7 and Table 5-III). With a wind speed of 17 knots, a temperature of 5°F is just as effective in chilling a man as a temperature of -40°F with a very slight wind of less than 2 knots. In the system shown, exposed flesh freezes at a wind-chill value of 1,400 kilogram calories an hour per square meter; however, it should be noted that the wind-chill concept has not yet been rigorously defined by physiologists, and its usage and usefulness are consequently still open to considerable controversy. Some mean monthly wind-chill values for the extreme month (physiologically coldest) for several locations are:

Baker Lake, Canada	2,030	Resolute, Canada	1,940
Barrow, Alaska	1,705	Thule, Greenland	1,605
Churchill, Canada	1,820	Verhkoyansk, Siberia	1,471

Although the Siberian Arctic has low mean and extreme temperatures, the Alaskan-Canadian Arctic, particularly in the barren lands northwest of Hudson Bay, has higher winter wind speeds. Combined with low temperatures, these produce greater physiological stress. But physiologically the coldest place in the Arctic is probably the interior of the Greenland ice sheet.



Notes:

1. THE RATE OF WINDCHILL IS BASED ON THE AMOUNT OF ENERGY IN KW/M^2/ HR NECESSARY TO MAINTAIN A CONSTANT INTERIOR BLACK-BODY TEMPERATURE OF 33°C (91.7°F) SIMULATING A HUMAN BODY TEMPERATURE IN ATMOSPHERIC ENVIRONMENT.
2. UNDER CONDITIONS OF BRIGHT SUNSHINE SUBTRACT 200 UNITS WINDCHILL KW/M^2/HR.
3. UNDER CONDITIONS OF HIGH HUMIDITY - ABOVE 60% - ADD 50 UNITS OF WINDCHILL KW/M^2/HR FOR EACH 10% INCREASE.

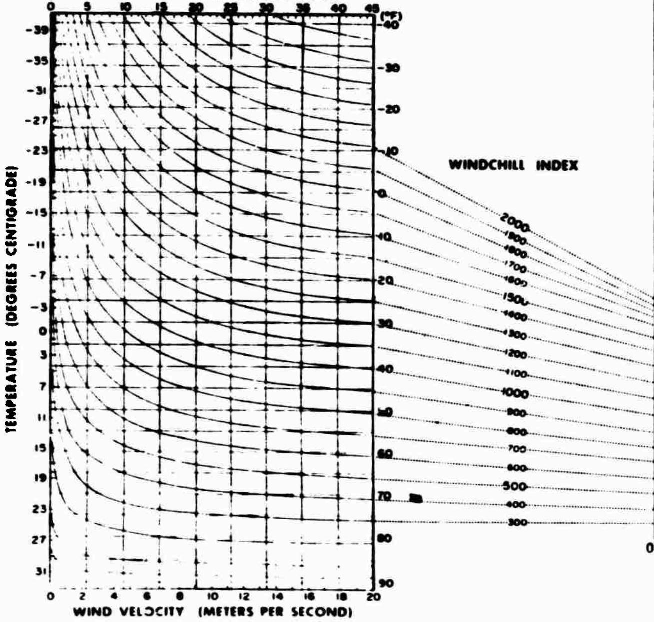
Figure V-7 WINDCHILL NOMOGRAM AND/OR T-W-P-R FACTOR WINDCHILL EXPRESSED IN KILOWATTS PER METER SQUARED PER HOUR

EXPLANATION ON USE OF CHART

BY A SIMPLE TECHNIQUE, IT IS POSSIBLE TO ESTIMATE THE PROBABILITY OF A SPECIFIED LEVEL OF WINDCHILL. DATA REQUIRED (MEAN MONTHLY AIR TEMPERATURE AND WINDSPEED) ARE ENTERED IN THE SIPLE WINDCHILL NOMOGRAM AT THE LEFT AND A WINDCHILL INDEX OBTAINED. THIS INDEX IS TRANSFERRED TO THE PREDICTION CHART AT THE RIGHT AND FOLLOWED TO THE PREDETERMINED LEVEL DESIRED (READ ON ACTUAL WINDCHILL SCALE AT THE EXTREME RIGHT). PERCENTAGE FREQUENCY CAN BE READ ON THE PROBABILITY SCALE AT EITHER TOP OR BOTTOM OF THE PREDICITION CHART.

<u>EXAMPLE</u>: AT FORT CHURCHILL, JANUARY MEAN TEMPERATURE (-18°F) AND WINDSPEED (14.9 MPH) ENTERED IN THE NOMOGRAPH AT THE LEFT GIVE AN 1,800 WINDCHILL INDEX. THIS 1,800 INDEX INTERSECTS THE 1,400 ACTUAL WINDCHILL (CONDITION AT WHICH EXPOSED FLESH FREEZES) AT 72 PERCENT ON THE UPPER SCALE OR 28 PERCENT ON THE LOWER SCALE, INDICATING THAT DANGER OF FREEZING IS A PROBABILITY 72 PERCENT OF THE TIME AT CHURCHILL DURING JANUARY SAFETY FROM FREEZING IS A PROBABILITY 28 PERCENT OF THE TIME. THE POSSIBILITY OF THE SITUATION BECOMING DANGEROUS FOR TRAVEL OR LIVING IN TEMPORARY SHELTERS (2,000 ACTUAL WINDCHILL) IS A PROBABILITY 16 PERCENT OF THE TIME.

SIPLE WINDCHILL NOMOGRAM



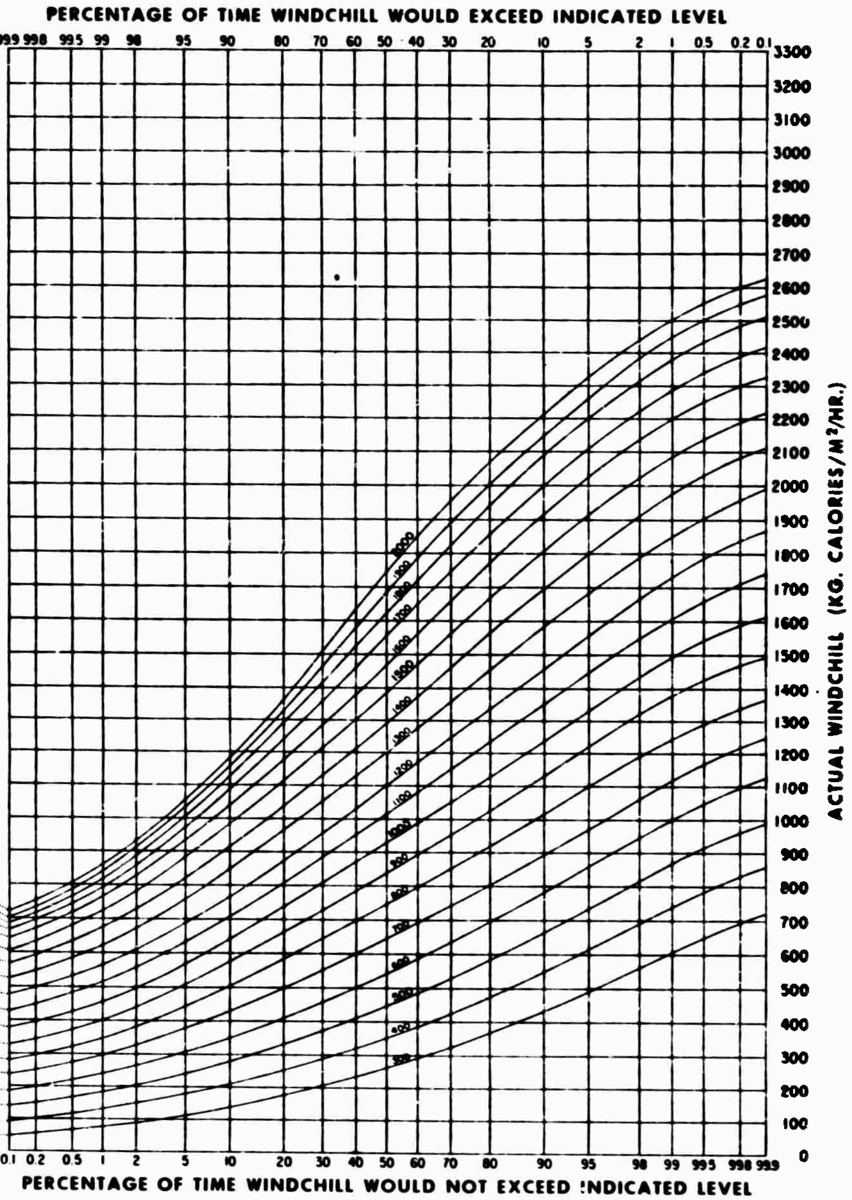


Figure V-7a WINDCHILL PREDICTION CHART

Table 5-III DESCRIPTIVE TERMS WITH WIND-CHILL VALUES APPROPRIATE TO DIFFERENT TEMPERATURES AND WIND SPEEDS

Windspeed		*Temp*		*Windchill*	*In sunshine*	*Light cloud*	*Thick cloud*
(mph)	*(m/sec)*	*(°F)*	*(°C)*				
5	2.5	90	32	18	Unbearably hot	Unbearably hot	Very warm
		80	27	146	Unbearably hot	Very warm	Warm
		70	21	269	Hot	Warm	Pleasant
		60	15.5	398	Warm	Pleasant	Pleasant
		50	10	528	Pleasant	Cool	Cool
		40	4.5	658	Cool	Cool	Very cool
		30	-1	788	Cool	Very cool	Cold
15	7	90	32	23	Unbearably hot	Very warm	Very warm
		80	27	187	Hot	Warm	Pleasant
		70	21	343	Warm	Pleasant	Cool
		60	15.5	508	Pleasant	Cool	Cool
		50	10	674	Cool	Very cool	Very cool
		40	4.5	840	Very cool	Cold	Cold
		30	-1	1005	Cold	Very cold	Very cold
25	11	90	32	25	Unbearably hot	Very warm	Warm
		80	27	206	Very warm	Pleasant	Pleasant
		70	21	378	Pleasant	Cool	Cool
		60	15.5	561	Cool	Very cool	Very cool
		50	10	744	Very cool	Cold	Cold
		40	4.5	927	Cold	Very cold	Very cold
		30	-1	1109	Very cold	Bitterly cold	Bitterly cold

B.5.6 Temperature

The most common misconception of the Arctic is that the land areas are perrenially covered with ice and snow and that winter is continuous and intensely cold. A large part of Greenland is a striking example of an ice-covered land possessing these qualities, and from it the rest of the north has been pictured by analogy. The high average elevation of Greenland and the precipitation it receives are two factors which strongly favor glaciation. In general, however, other arctic lands posses neither of these characteristics, and over a large portion of the Arctic the scanty snows melt rapidly with the approach of summer. Most of the little snow that does fall is soon swept by the wind into gullies and the lee of hills or structures, so that more than three-quarters of the arctic lands are comparatively snow-free at all seasons. With respect to the intense cold: at the North Pole the lowest temperatures generally do not fall below -65°F, a figure that is reached equally often in some parts of the North American prairies. The coldest observed temperature on the surface of the earth before 1957 was recorded at Verkhoyansk, near the Arctic Circle in eastern Siberia and well over 1,200 mi from the geographic pole. However, the temperatures recorded at Oimyakon, about 300 mi southeast of Verkhoyansk, have consistently averaged lower than those recorded at the same time at Verkhoyansk. The official minimum at Oimyakon is -90°F. Only the Antarctic, occupied since 1957, is colder (-128°F).

A great variety of climatic conditions are encountered in the arctic regions (Fig. V-8). Areas adjacent to one another are found to have widely differing climatic characteristics determined by latitude, marine influence, and topography. A study of annual temperature curves reveals three well-defined types - maritime, coastal, and continental:

(1) The Arctic Ocean has a flat temperature curve during June, July, and

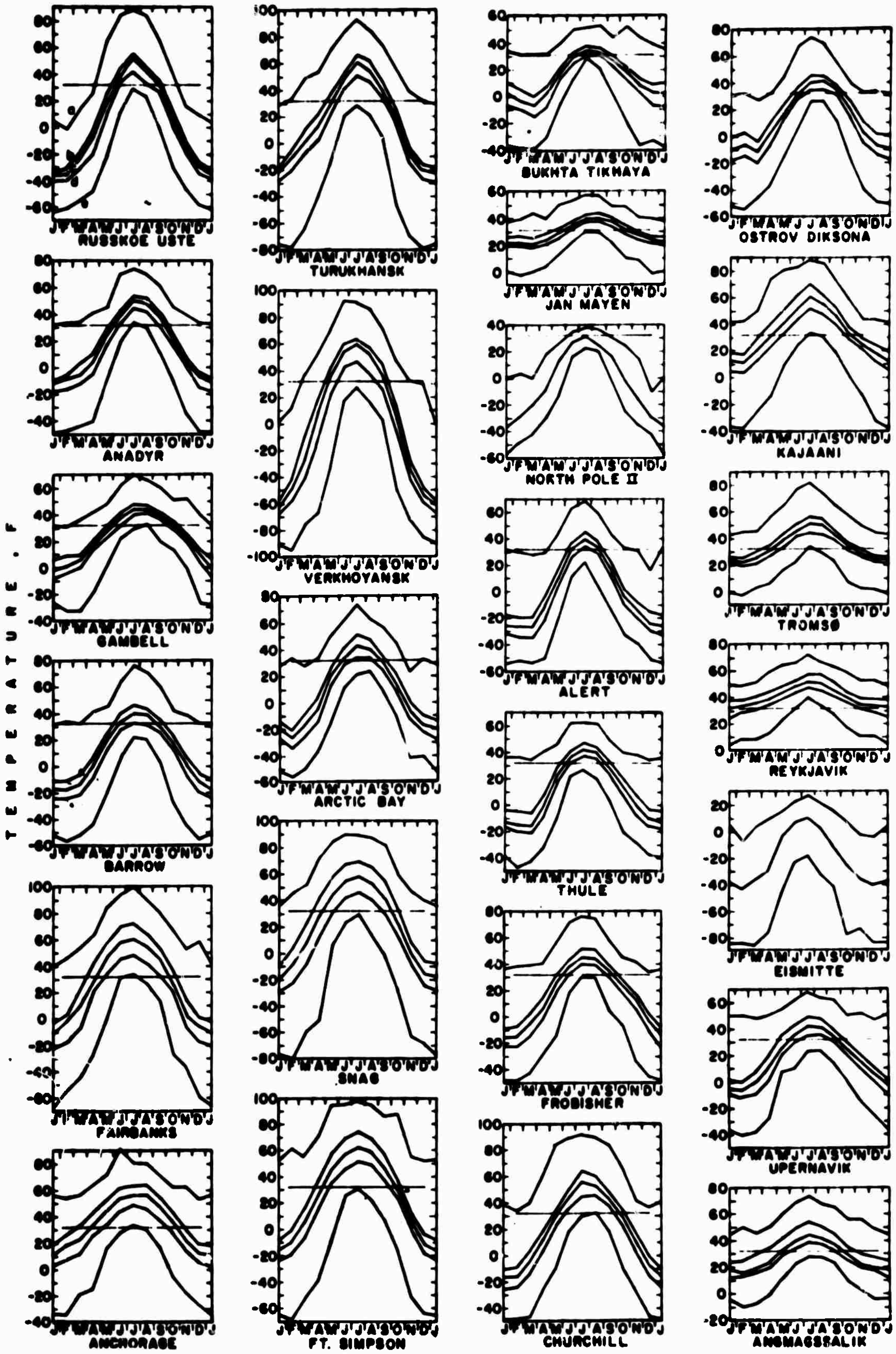


Figure V-8 MONTHLY TEMPERATURE DISTRIBUTION AT SELECTED STATIONS[166]
[(a) Absolute maximum; (b) Mean maximum; (c) Mean; (d) Mean minimum; (e) Absolute minimum]

August with small deviations from the freezing point. The winter curve is again flat, but at this time the temperatures stay around -29°F.

(2) The coastal climate is quite similar to the maritime one. The year consists of a long cold winter and a short cool summer similar to the spring or autumn experienced on the continent. The annual temperature curve has the same winter characteristics as that of the Arctic Ocean but there is a seasonal maximum in July. The mean summer temperature, however, remains below 50°F.

(3) The arctic continental climate is characterized by very low winter temperatures with a pronounced winter minimum and high summer temperatures with a very pronounced summer maximum. Here the annual range of mean temperature may be as much as 110°F.

In Eurasia, winter temperatures (Fig. V-9a) show a marked decrease from the warm coastal areas of Scandinavia to northeastern Siberia and the region around Verkhoyansk where the January mean is -54°F. The reason this area has the most extreme winters in the northern hemisphere is readily understood from its geographic location. Heat from the North Atlantic cannot penetrate into or over the frozen continent; the Pacific Ocean has no appreciable effect, since ranges of mountains intervene and the winds are prevailingly offshore; the ice-covered Arctic Ocean cannot modify the conditions; and the ranges of central Asia check any influence from the warm Indian Ocean far to the south. The characteristic lag of seasons, so highly pronounced on the arctic coasts and conspicuous at any locality with maritime influence, is almost totally absent. The temperature curve follows more nearly that of insolation and April (Fig. V-9c) is as warm as October (Fig. V-9f). In the interior of Canada the same characteristics are evident, but since the area is not comparable to the vastness of the Siberian land mass, the feature is less pronounced. The Siberian winter is by no means as unpleasant as its extreme temperatures might suggest - the air is often calm and the skies clear. Danger to man and beast occurs only when the wild

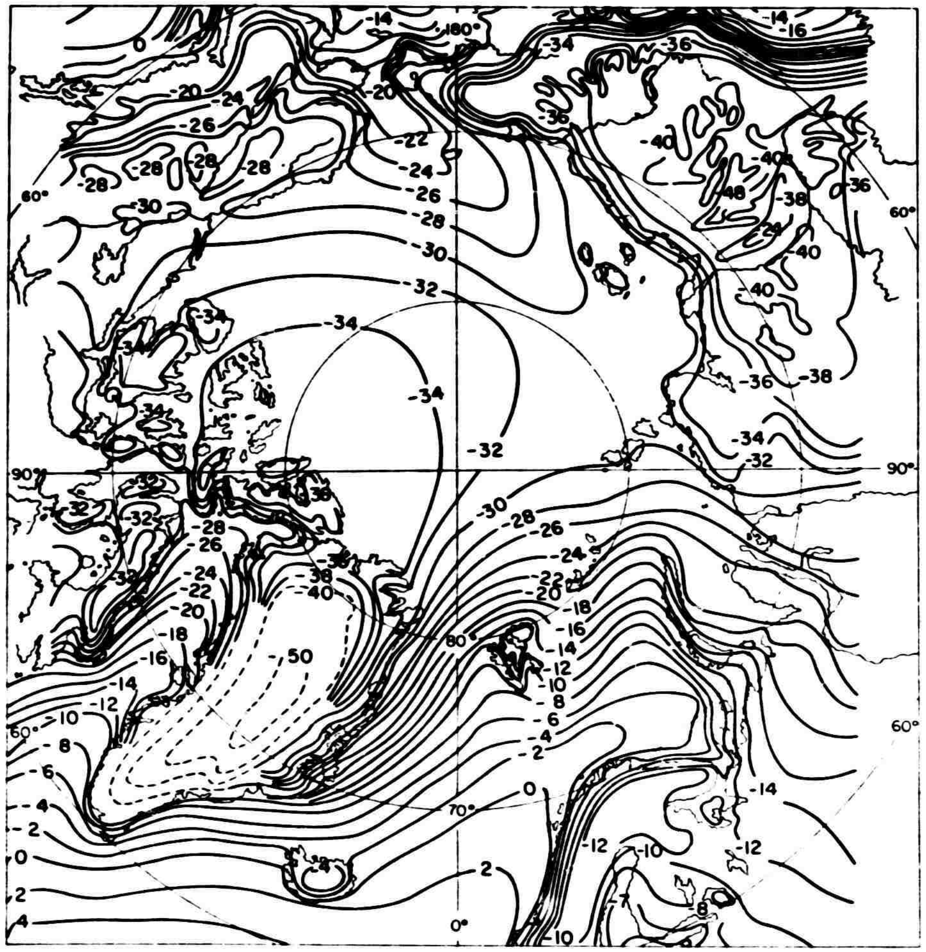


Figure V-9a Mean Air Temperature (°C), January

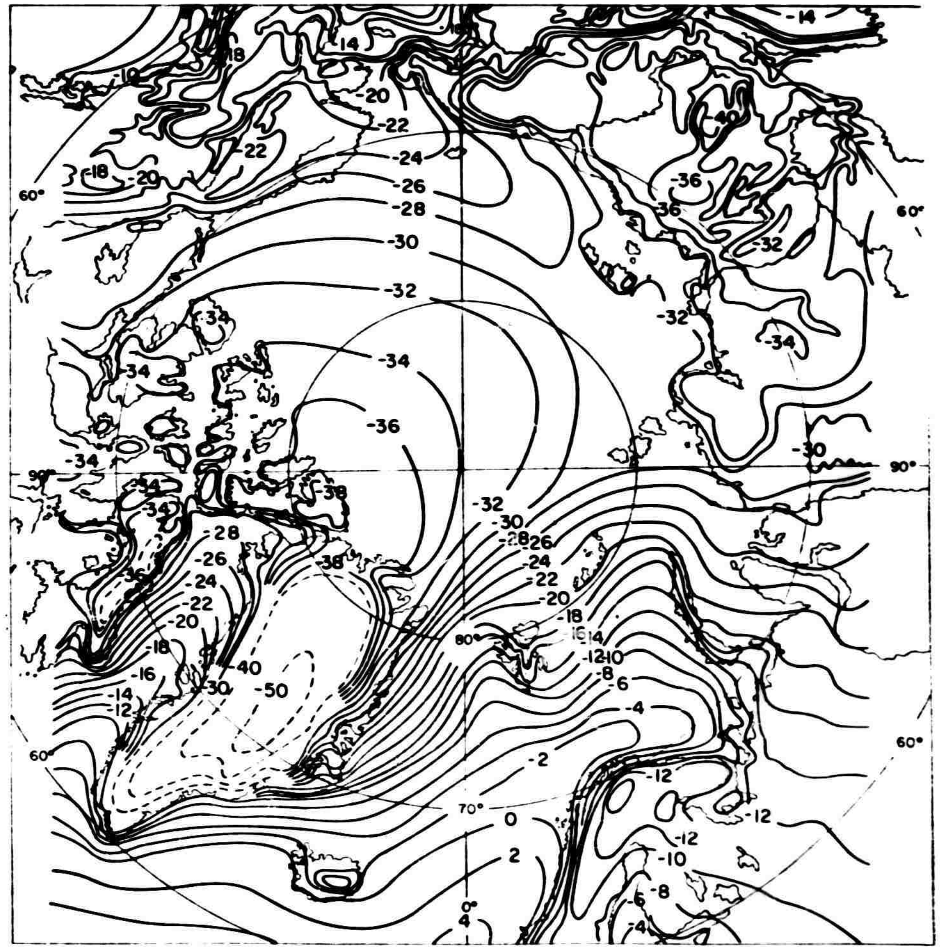


Figure V-9b Mean Air Temperature (°C), February

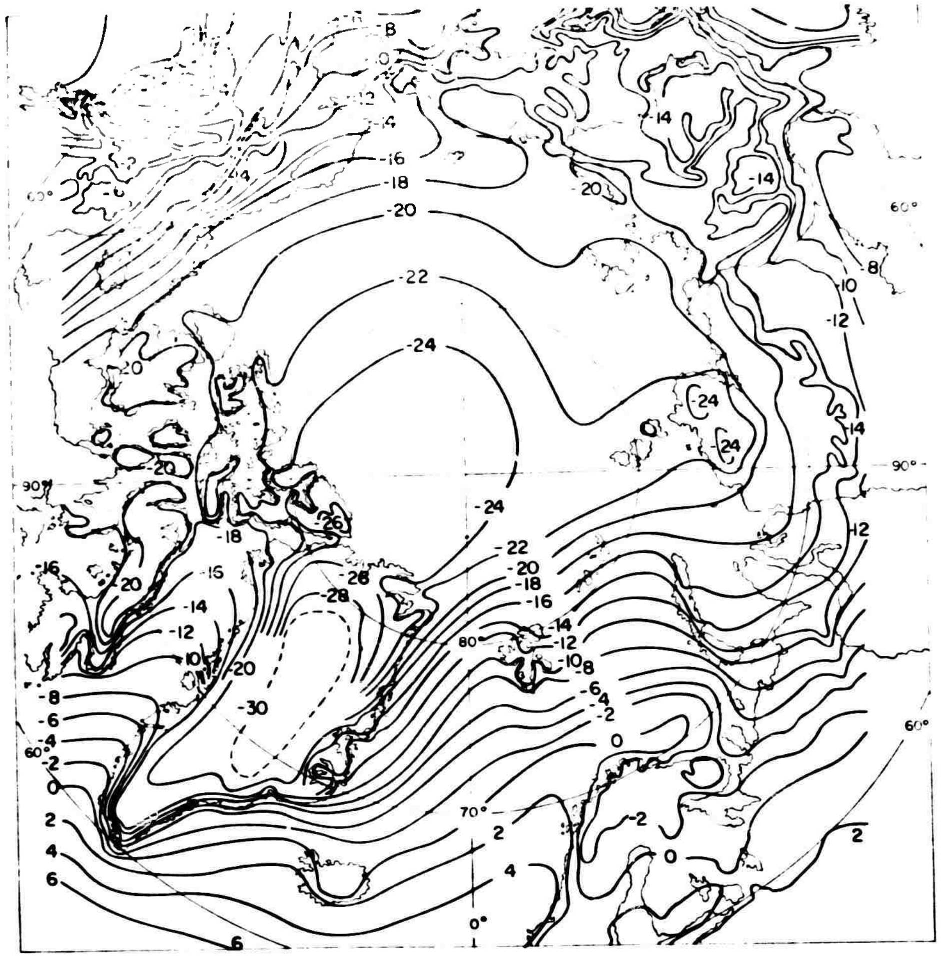


Figure V-9c Mean Air Temperature (^{o}C), April

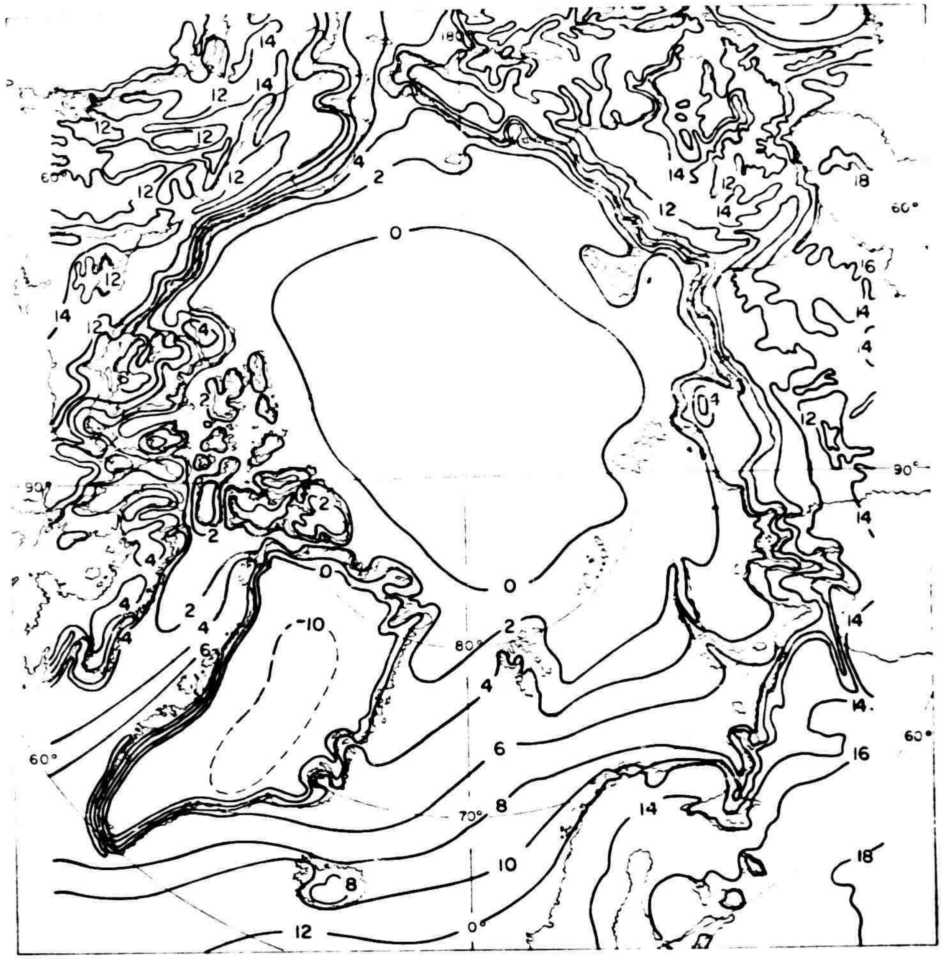


Figure V-9d Mean Air Temperature (^{o}C), July

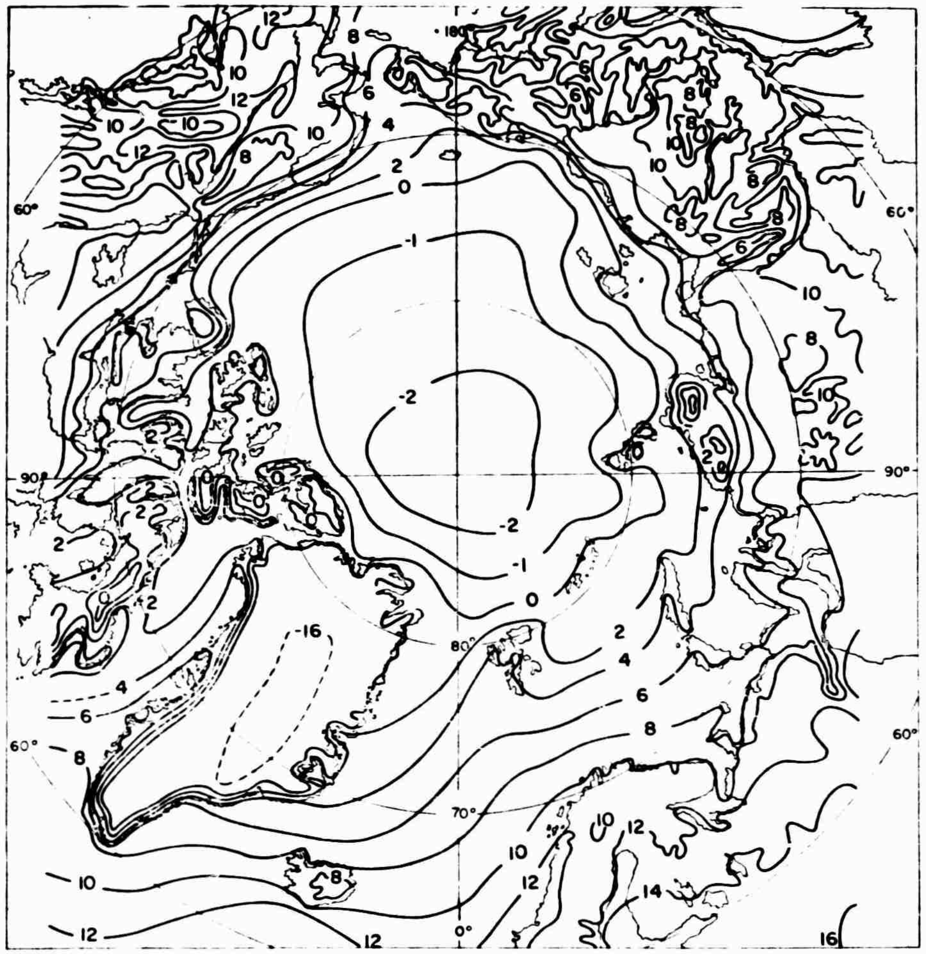


Figure V-9e Mean Air Temperature (°C), August

Figure V-9f Mean Air Temperature (°C), October buran blows. Similar blizzards also occur in the interior of Canada. The probable occurence of extremely low temperatures in the North has recently been computed by U.S. Air Force personnel and are presented in Figures V-10a through d. The most conspicuous element of these calculations is the rapid diminution of the likelihood of extremely low readings in North America as the temperature decreases.

The irregularities of the winter isotherms over the Arctic Ocean are produced by the influence of the Atlantic and Pacific oceans. In the North Atlantic the isotherms are pushed far northward, and the greatest positive anomaly of temperature (the greatest departure from latitudinal average) in the world occurs here. The winter temperatures in the vicinity of Bering Strait are influenced by the open waters of the Bering Sea. North of the Strait the influence is limited to a small area in which large local differences occur with different wind directions. Large and rapid temperature fluctuations are otherwise not a common feature over the Arctic Ocean. The air temperature near the surface is primarily dependent on the temperature of the ice surface itself. On occasion, however, warm air from the Atlantic may raise the winter temperature near the North Pole by as much as 54°F. After the warm air import has ceased, radiation control again dominates the surface temperature which will show an equally drastic drop. Strong winds may also cause higher surface temperatures by breaking down the temperature inversion and mixing the warmer air from above with that near the surface. Renewed cooling follows every slackening of wind.

The winter temperatures over the pack ice remain nearly constant for a considerable time. This flat minimum represents the temperature at which the loss of heat by radiation from the snow and ice surface balances the heat which is conducted to the surface from the water under the ice plus the heat trans ported into the Arctic by intense warm air advected in cyclones. In periods of little cyclonic activity, heat transfer to the surface from the atmosphere is

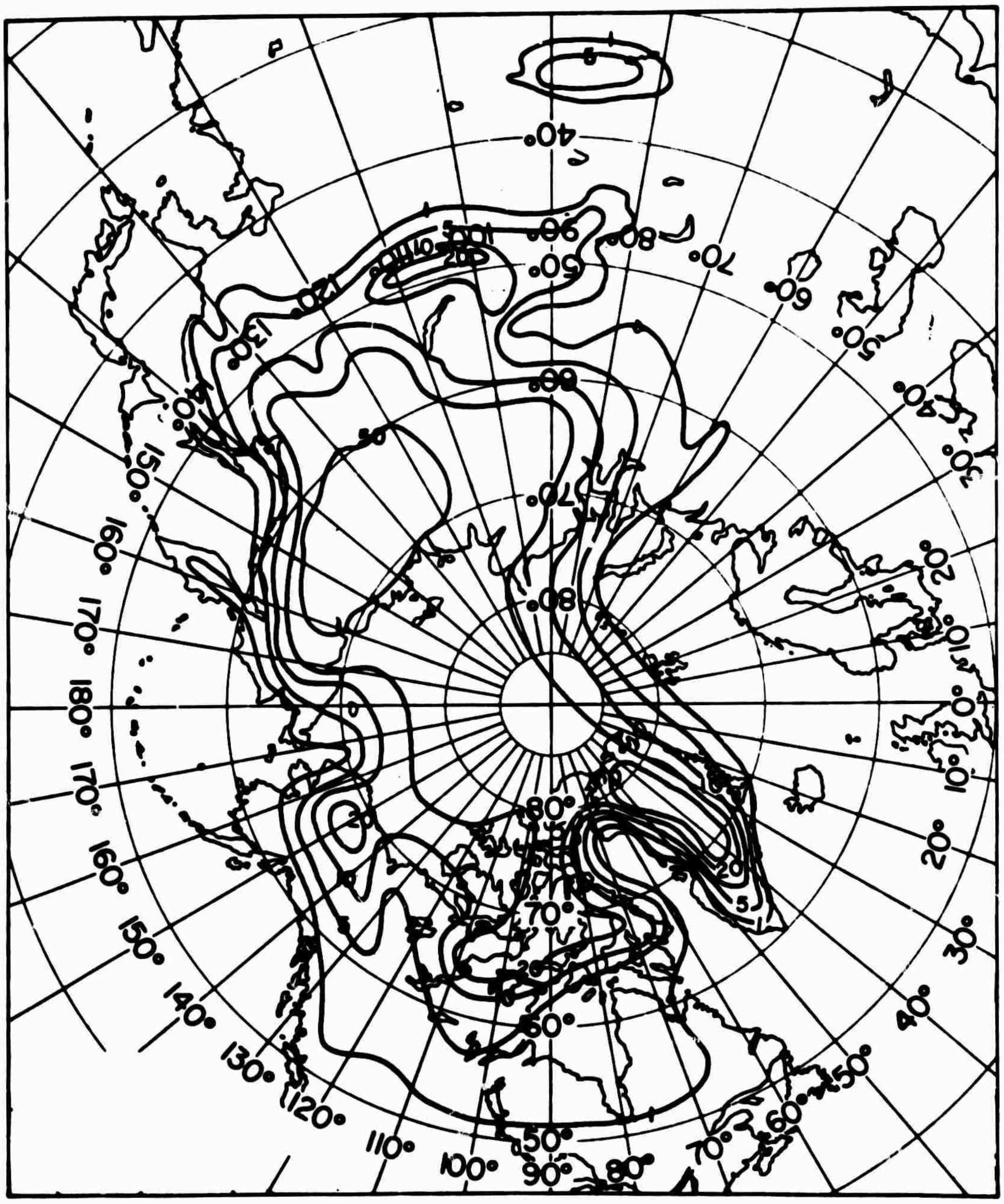


Figure V-10a ESTIMATED RISK (PERCENT OF COLDEST MONTH) TEMPERATURE BELOW -40°F

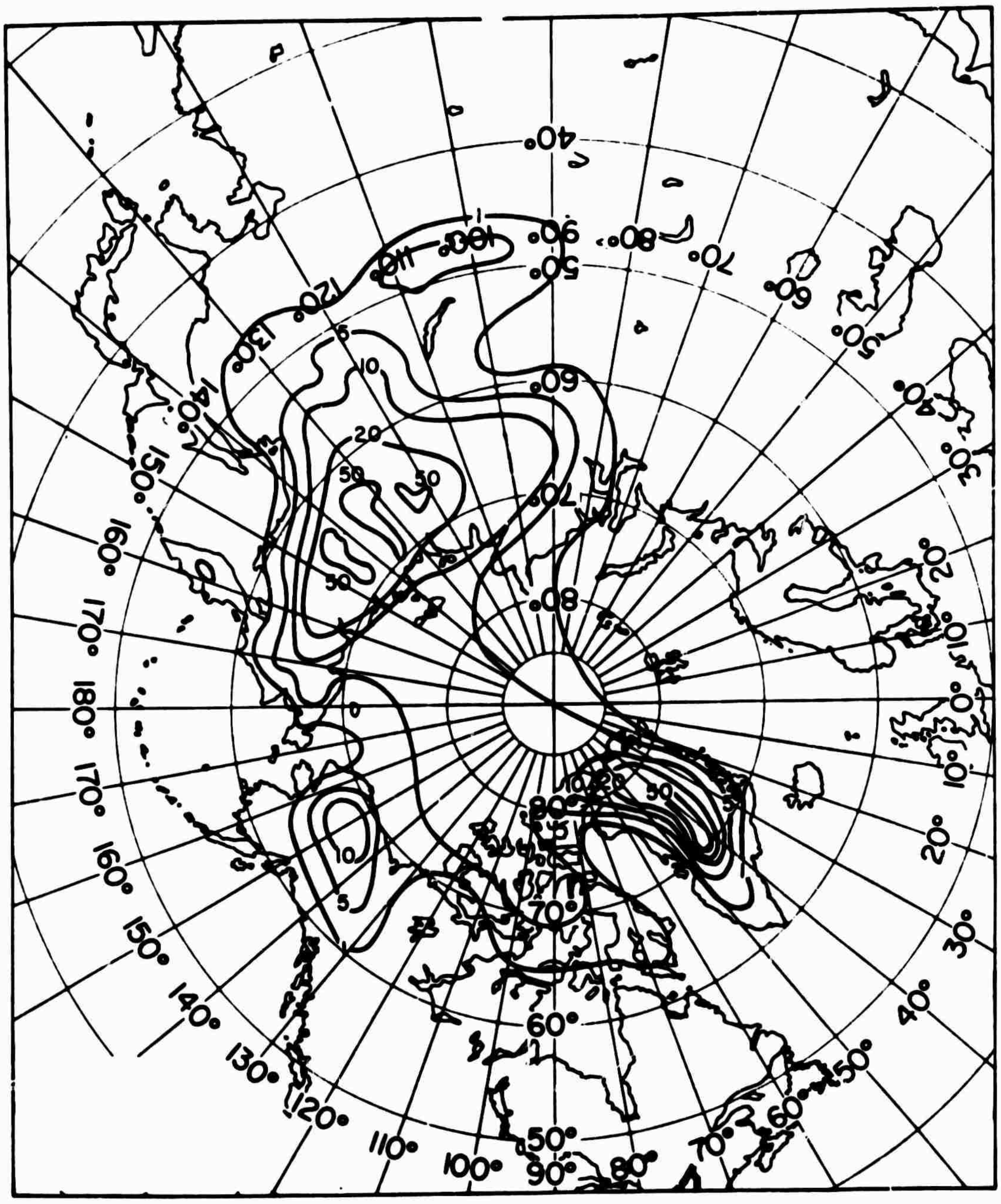


Figure V-10b ESTIMATED RISK (PERCENT OF COLDEST MONTH) TEMPERATURE BELOW -50°F

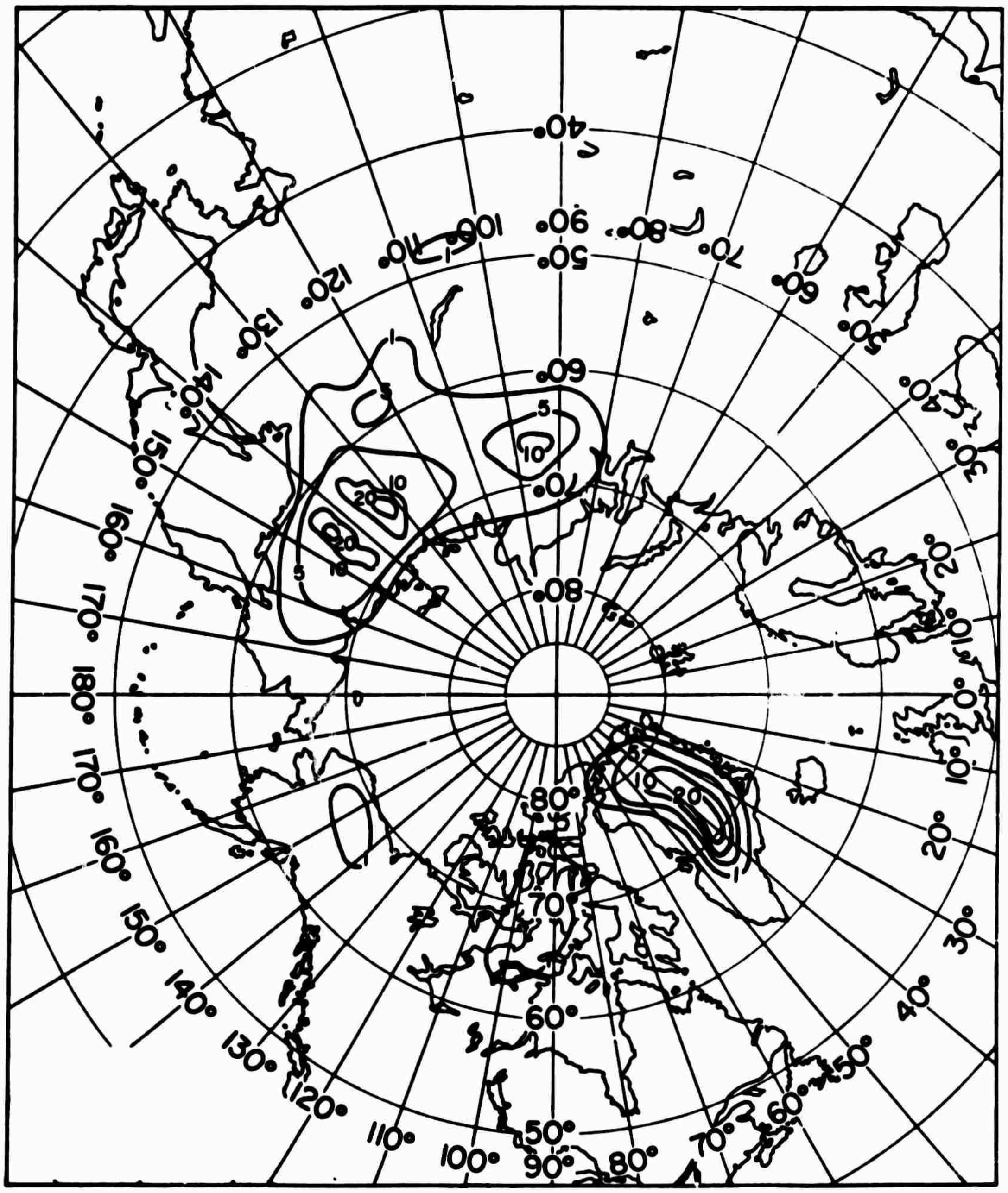


Figure V-10c ESTIMATED RISK (PERCENT OF COLDEST MONTH) TEMPERATURE BELOW -60°F

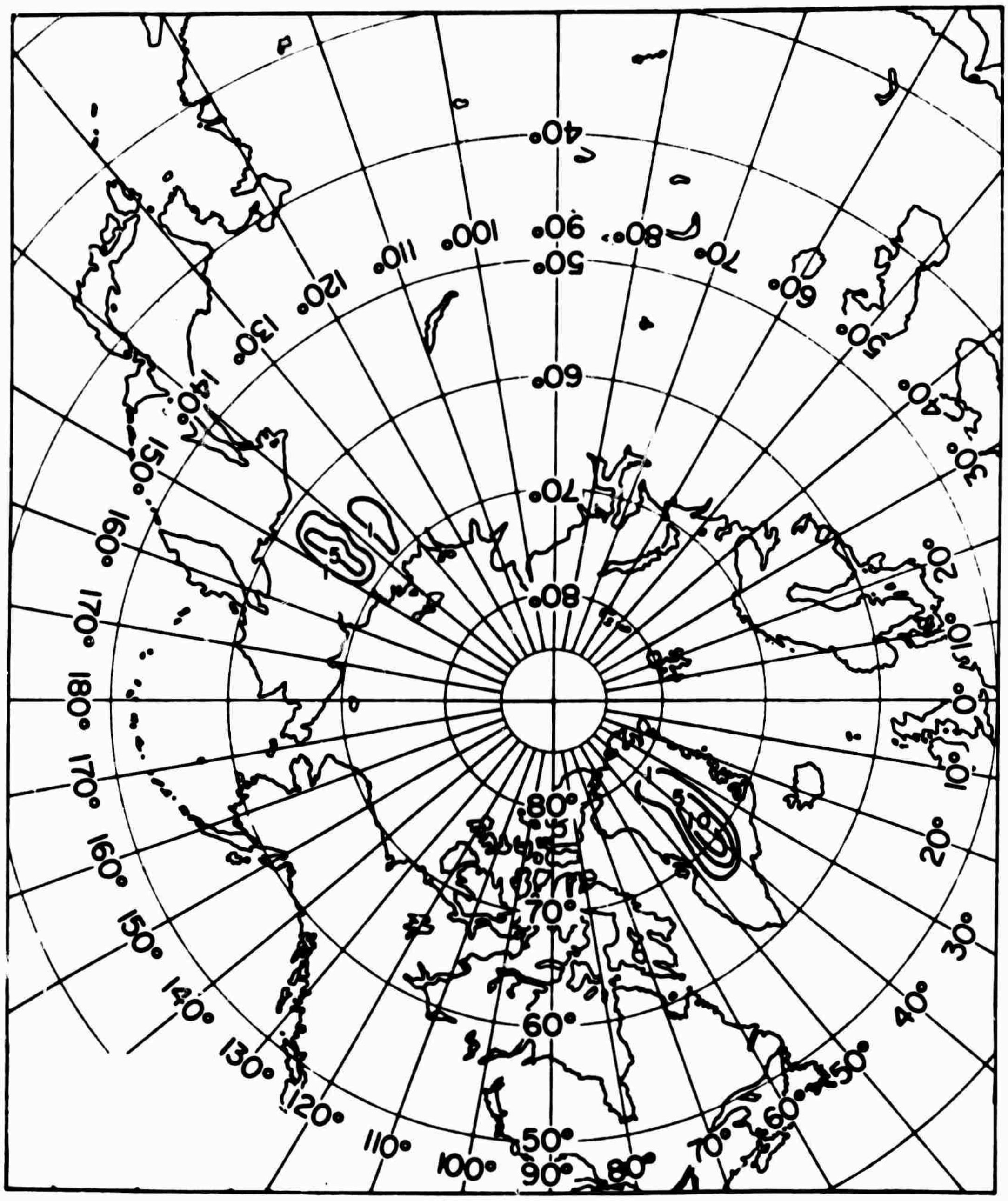


Figure V-10d ESTIMATED RISK (PERCENT OF COLDEST MONTH) TEMPERATURE BELOW -70°F small since the eddy conductivity is very small in the inversion layer. The winter air over the Arctic Ocean is, as a rule, nearly saturated with water vapor. Surface temperatures are mainly governed by net radiation and conduction of heat to the surface from below.

Minimum air temperatures (Fig. V-11) must occur when the net radiation loss is least balanced by transport of heat from the water. This will give a minimum temperature at the surface of about -40°F down to -58°F over thick ice. The maximum temperature in overcast calm weather is reached at about -13°F. The extreme maximum temperatures are nearly linear functions of the wind speed, increasing with increasing wind speed due to the transport of heat from above the inversion.

The presence, or nearness, of open water is seen in the temperature distribution in the winter half of the year when broad tongues of mild temperatures extend into Barents Sea and Baffin Bay and across Bering Strait. The strongest influx of heat by cyclonic activity occurs in the Atlantic section of the Arctic Ocean in January and February; and, as a consequence, the lowest temperatures are observed in that section in March. In the other regions the lowest winter temperatures usually appear in January or February. From April until June the temperature rises quite rapidly until the ice begins to melt. The mean summer surface temperatures (Figs. V-9d and e) remain fairly constant, conforming to the nature of the underlying surface. Almost all regions are warmest in July. Temperatures close to the melting point prevail over the pack ice and along the fringes of the Greenland ice sheet. Cool temperatures dip southward across Bering Strait and Baffin Bay. Maximum temperatures in the Arctic Basin do not exceed 41°F. An examination of large departures of monthly mean temperatures from normal values for 20 coastal or island stations, from 20 or more years of observations, showed that the greatest departures are found in winter (from

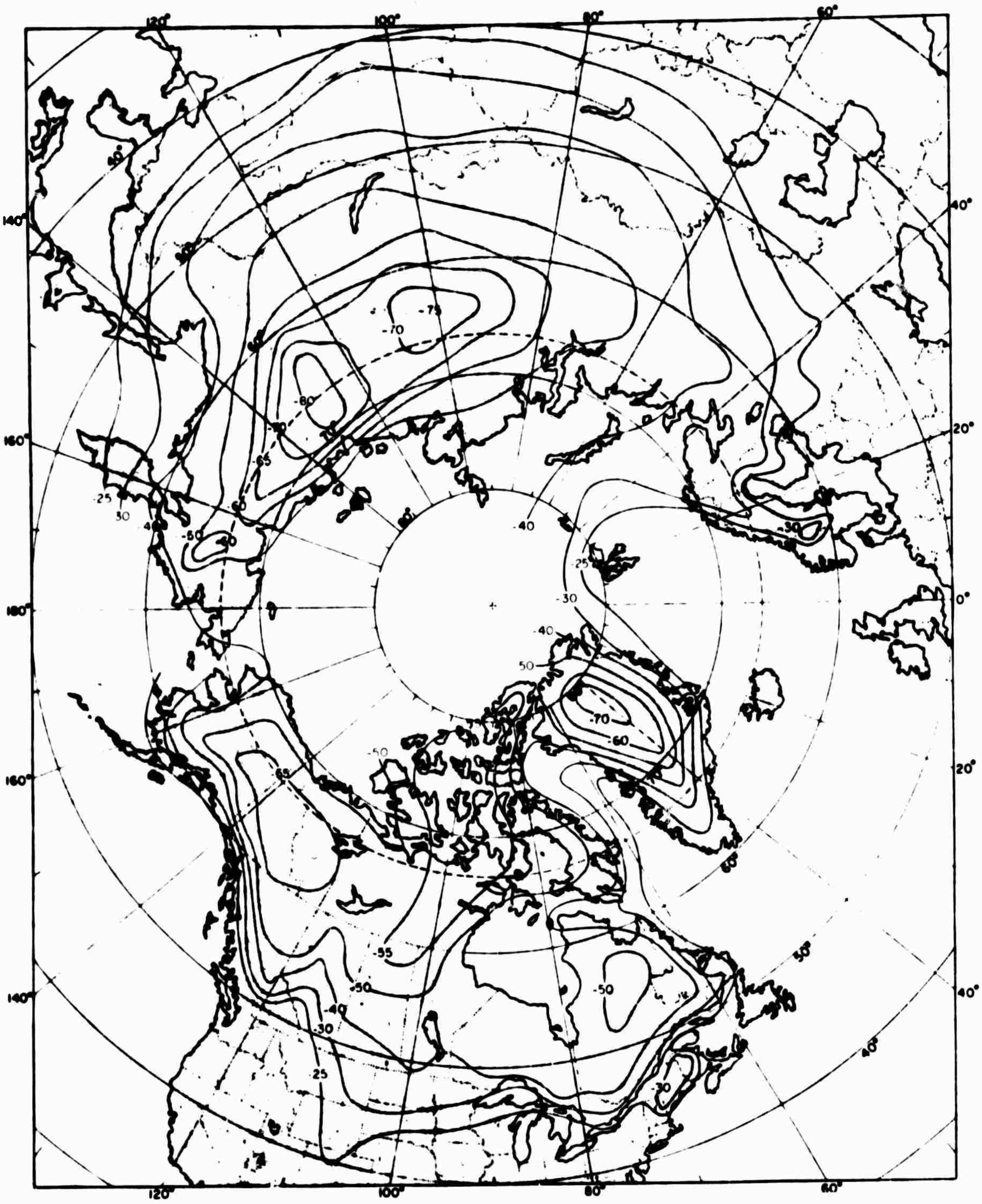


Figure V-11 AVERAGE OF THE LOWEST TEMPERATURES (°F) OF EACH YEAR[232] mount to 0.8 inches - about 15% of the annual total precipitation. However, the deposition is less on a horizontal surface, and hoarfrost is of small importance to the total snow cover in winter.

B.5.7 Visibility

The melting of the pack ice in summer leads to formation of persistent fog and low cloud. Arctic Ocean stations experience fog more than 100 days per year, most frequently in summer and least in winter. The summer fog is generally caused by advection of relatively warm and moist air over the melting ice or cold water, It is patchy and of fairly short duration. Advection fog is particularly prevalent from June to September, occurring on 10% of all observations in June, 15% in July, 25% in August, and 7% in September. The fog does not occur at wind speeds above 20 knots. During the cold season, small patches of steam fog form over open water leads in the pack ice. This type of fog, sometimes called arctic sea smoke, develops when cold air blows over open water causing rapid discharge of moisture and heat to the air.

The types mentioned above are water fogs. There also occurs during the arctic winter a phenomenon called ice fog. This may form where large quantities of water vapor are added to very cold air, preferably about -22°F. Lightweight ice particles, with a slight fall velocity, remain suspended in the stagnant air near the surface for a considerable period. Such fog occurs locally in the vicinity of human habitation. The second type that has major importance in the Arctic is the radiation fog of winter. These fogs form readily under an inversion in very cold weather and are caused by the cooling of the lowest layer of air by contact with the surface. Cooling by radiation takes place most rapidly over the land areas and least rapidly over open water. The fogs that are formed are generally thin and shallow; they occur most frequently along river bottoms where the air drainage is poor and the air movement sluggish. The locale of most frequent occurrence appears to be in the lower Lena River valley and the lower

November to April) and small departures are most probable in July. Of the departures in winter, the greatest were observed in the Kara Sea and the smallest were noted in the East Siberian Sea. In summer the greatest observed departures from normal temperatures were on the coasts of the peripheral seas and decreased toward the north.

The temperature inversions observed in all arctic regions are especially prominent over the pack ice. As would be expected, the diurnal variation is somewhat irregular and the maximum occurs during the night hours and the minimum during the middle of the day. Again, the wind accounts for the irregularity; any slight increase in wind, regardless of direction, produces a considerable rise in temperature by the process of mixing.

In summer the melting of snow and ice keeps the surface temperature close to the freezing point. The number of days with a maximum temperature slightly above the freezing point is very nearly the same all along the latitude of 75°N - about 40 days. Nearer the Pole, positive temperatures are usually observed in the second half of July on some 10 to 15 days.

The small temperature variability in summer is a characteristic common to all stations on the central Arctic Ocean. The temperatures over the pack ice never deviate far from the freezing point. The diurnal ranges are smaller and the interdiurnal changes are less in summer than in any other season. Warm air from inland is often transported a considerable distance from the coast; but, in passing over the ice-filled waters, it is cooled so effectively by contact with the ice that a sharp inversion is formed. Hence, the warm air current flows at some distance above the ground and has practically no influence on the surface temperature. Considerably higher temperatures are found on the coast and the change from coastal to maritime conditions takes place within a very short distance. However, this characteristic holds good for the surface layer only and does not apply at higher levels. The discontinuity, therefore, represents only a quasi-front and is of no consequence to the circulation of the atmosphere.

B.5.8 Humidity

The relative humidity of the air over the Arctic Ocean in the winter always remains near 100%. During this period it should be calculated as the relative humidity over ice, i.e. with respect to the saturation vapor pressure over ice, which is lower than over water. This permits the air to be slightly supersaturated in respect to ice, while it is not saturated in respect to water. It will cause water droplets to evaporate while water vapor condenses on ice crystals. Supersaturation occurs frequently near the surface in winter, as long as no condensation takes place. Normally, however, hoarfrost formation begins when a value slightly over 100% has been reached. The relative humidity over ice has a maximum in midwinter (103%) and a minimum in June (95%), while the relative humidity over water has a maximum in August (96%) and a minimum in midwinter (73%). The actual pressure of the water vapor is very small in midwinter (0.2 mb), and it has a maximum in July-August (6 mb).

The amount of hoarfrost which is deposited is proportional to the wind speed and to the difference between actual pressure of the water vapor and the saturation vapor pressure over ice. Hoarfrost is formed only when the relative humidity over ice is greater than 100%. The amount of frost deposition is greatest at a temperature of around -20°F while the probability of hoarfrost has a maximum at about -26°F. In July and August there is relatively little hoarfrost since air temperatures usually are around freezing point. Otherwise, the smallest amounts occur in April and the greatest in September. In the latter month there is, over the entire Arctic Ocean, a surplus of water in the air since open leads in the sea ice are frequent. Hoarfrost is considerable on the vertical surfaces and may a-

Mackenzie River valley. They are also quite frequent in the Yukon valley and the valleys of the principal northward-flowing rivers of central and eastern Siberia. Radiation fogs are also common over the pack ice, these fogs may be expected from 8 to 12 days per month during the winter, and in adverse years the number of days of radiation fogs may be doubled. In a few localities, such as the lower Mackenzie in the vicinity of Aklavik, early morning radiation fogs may be expected on 20 to 25 days monthly during the coldest months. It is stressed that these frost fogs do not present as serious an obstacle to aviation as do water fogs. They are generally thin fogs, so that contrasting objects can be distinguished on the surface directly beneath the aircraft. It is only in the brief interval during which the aircraft is passing through the thin fog layer that visibility is seriously reduced. With experience, landings and take-offs can be made with safety through the shallow frost fogs.

As a rule, fogs are not to be expected in summer with offshore winds. The offshore winds are relatively dry and are not cooled sufficiently to cause the formation of fog. Summer fogs are most frequent with light winds but under favorable conditions they may occur with high winds. Since they are advection fogs, they do not form readily in calm weather. The opposite is true of the mists or frost fogs of the winter season. The latter are radiation fogs and form most readily in calm weather or with very light winds. High winds are very unfavorable for the formation or continuation of fog in the colder months of the year.

Fogs in the Arctic are more frequent in the late-night and the early-morning hours than at any other time. In the Kara Sea region the maximum frequency of fogs occurs at midnight or in the hours immediately following and the minimum frequency occurs at noon or in the early afternoon. These observations agree closely with experience in the East Siberian Sea. Data on the diurnal distribution of mists or frost fogs elsewhere are not so conclusive but indicate a similar distribution for the summer advection fogs. Frost fogs are more frequent in the early morning than at noon or in the evening.

It will be noted from the figures shown in the following table that the summer fogs are less frequent over the Barents and Norwegian seas. This is due to the more turbulent nature of the lower atmosphere over these seas, resulting in less fog and more low clouds.

The matter of visibility in the Arctic is a complex one due to the extreme clarity of the air. The records contain many accounts of the extreme range of visibility - it is not uncommon to see dark mountains 60 mi distant. On the other hand, the lack of contrast, particularly where all surface objects are covered with new snow, results in the inability to distinguish some objects close at hand. As the arctic atmosphere is fairly uncontaminated by impurities, the visibility will be reduced only, apart from fog, by precipitation or blowing snow.

Wind speeds of about 5 knots will cause unconsolidated snow to drift along the surface. At speeds of 10 to 15 knots the snow is lifted into the air, and when it reaches a height of 6 ft the term 'blowing snow' is used. Coastal stations around the Arctic Ocean may experience more than 100 days of blowing snow per year - in winter on more than half the days.

In late autumn and winter, falling ice crystals may be observed when the moisture of advected air condenses over the cold ocean. Fine ice needles form a slight haze as they settle slowly from an otherwise-clear sky. Such falls of ice crystals add little to the snow cover, however.

The frequent well-marked temperature inversions of the Arctic explain the many accounts of mirages. Objects that are known, beyond any doubt, to be

Table 5-IV AVERAGE NUMBER OF DAYS WITH FOG

PLACE	JAN.	MAR.	MAY	JULY	SEPT.	NOV.	ANNUAL
Dikson Island	5.5	7.9	4.9	22.9	10.5	5.2	152.2
Kola	0.7	0.2	1.0	3.1	1.9	1.6	16.1
Tiksi Bay	3.7	4.4	5.3	9.0	4.6	1.0	66.1
Verkhoyansk	3.4	0.6	0.6	1.5	4.1	1.8	25.9
Wrangel	1.6	4.6	12.2	21.4	10.8	3.8	113.5
Yugor Strait	4.5	6.2	9.1	19.6	14.1	4.5	123.8
Aklavik	19.6	27.6	2.0	2.2	21.2	15.6	165.6
Tromsø	1.0	0.8	1.0	2.0	2.0	1.0	14.0
Vardo	0.0	0.0	1.0	6.0	1.0	0.0	18.0
Green Harbor	0.1	0.1	0.4	4.0	1.0	0.0	13.0
Bear Island	1.1	2.7	6.7	18.7	12.0	3.3	82.0
Jan Mayen	1.9	3.2	3.3	12.8	5.9	1.8	58.3
Angmagssalik	1.0	1.0	7.0	9.0	4.0	2.0	48.0
Godhavn	9.0	11.0	15.0	17.0	8.0	3.0	127.0
Godthaab	1.0	1.0	7.0	13.0	7.0	1.0	61.0
Ivigtut	0.2	0.4	2.0	6.0	3.0	1.0	26.0
Upernivik	2.0	2.0	5.0	11.0	2.0	1.0	47.0
Churchill	1.0	*	1.0	2.0	1.0	*	13.0
Craig Harbour	*	1.0	0.0	2.0	1.0	*	9.0
Lake Harbour	2.0	1.0	2.0	5.0	2.0	3.0	30.0
Pond Inlet	1.0	*	*	*	*	2.0	9.0
Dutch Harbor	0.9	0.9	2.5	3.5	1.8	0.0	18.4
Fairbanks	17.2	11.6	0.8	2.4	3.2	6.1	68.0
Nome	5.3	5.2	5.0	8.1	4.4	3.2	67.2
Point Barrow	5.0	2.3	7.5	15.0	5.4	2.7	82.6
St. Paul's Island	1.7	2.8	6.9	7.0	2.8	1.5	52.2

* Less than one day

below the horizon are not infrequently visible as mirages and the periods of daytime and twilight are lengthened as the normal index of refraction is altered. The inversions may also interfere with the identification of landmarks through distortion while estimation of vertical distances is made much more difficult.

Whiteout is a very simple phenomenon with easily understood causes and only one direct effect: the loss of depth perception. This results in quite unexpected and sometimes disastrous consequences. The only recourse on the surface is to shuffle along at a slow pace. In the air, the only safe procedure is instrument flying rules, making landing difficult, if not impossible, at many arctic airfields.

Only 2 conditions are necessary to produce a whiteout: (1) a diffuse, shadowless illumination; and (2) a uniformly monocolored, white surface. In polar regions thses conditions occur frequently. Large unbroken expanses of snow are illuminated by a sky overcast with dense, low, stratus clouds that blot out all traces of surface texture or shadow and merge bumps and hollows and snow-covered objects into a flattened, white background. Those who have not been exposed to whiteout are often skeptical about the inability of those who have experienced it to estimate distances under these conditions. In the normal environment there are so many direct and indirect clues to depth that it is hard to understand how one who can see clearly can still fail to guage distances accurately.

The primary clues are those that result directly and without the intervention of thought or logic in a "feeling" of depth. These are unaffected by whiteout. The remaining clues to depth - and they are of the greatest importance in everyday life - are secondary: that is, the mental estimate is not directly in terms of distance but of appearance relative to some standard. Normally, there is a continuous succession of familiar objects, starting from somewhere well within the range of binocular depth perception (where primary clues are effective) and extending out to and beyond the object of attention. Both foreground and background combine to furnish many clues to the distance of any object, even objects of unknown size. This is the crux of the matter, because the entire visual continuum is erased by whiteout, and in such an unfamiliar situation all vestiges of depth perception may be lost without one's even being aware of impairment. Naturally, this frequently leads to extreme confusion and even disorientation.

Most concern about whiteout is in relation to flying safety. Whiteout is generally a problem to the pilot in landing, takeoff, and taxiing. However, low flying, at altitudes less than 1,000 ft above the terrain, is particularly hazardous under these conditions and has resulted in numerous wrecks that are strewn across the entire Arctic.

Ground operations are also impaired, whether they involve walking or the use of vehicles. On an overcast morning following a new snow, the roadsides may be littered with vehicles, frequently including the snow-removal machinery. Often roads are closed - not because of snow depth, but because of whiteout; and the whiteout need not be total.

Solutions may result from an approach involving 2 possible avenues - the physical and the psychological. Physically, the point in question is whether the quality of light reflected from snow-covered objects may vary in some subtle way with the angle presented by the surface or from the incident sky-light, so that contrast might be enhanced by the use of appropriate filters. Yellow nickel-coated glasses seem to be helpful to pilots. While the use of intense directional artificial light would provide adequate shadows and contrast, it is not a solution of general value because of its impracticability during normal daylight hours. The use of smoke bombs on landing strips has been suggested, particularly if there is a little wind to move the smoke across the snow in such a fashion as to create stained "shadows." However, many objects are required in the visual field to provide adequate clues to depth, and some 250 to 500 points are required for good motion parallax. The most suitable method yet found in the latter connection is to line each side of the airstrip with empty fuel barrels.

In addition to the clear air whiteout, there is an increasing tendency to speak of obscuration-type whiteouts which may be caused by water fog, ice fog, blowing snow, and/or precipitation. In these instances vision is not only impaired by loss of depth perception, but also by the presence of hydrometeors that limit the seeing range. Although considerable study has been made of these conditions, standard weather stations report weather and obstructions to vision in terms of the causal agent (Tables 5-V and VI) while climatic tabulations group all the conditions together (Fig. V-12 and Table 5-VII).

In blowing or falling snow, fog, or sea smoke - conditions that may be concomitant with whiteout - and in arctic twilight with freshly fallen or blowing snow, in all probability nothing will help the aviator; but men on the surface can follow well-flagged trails. The training of pilots to cope with the whiteout phenomenon presents difficulties, for whiteout is not always available and hence establishing a training program would be rather haphazard; but the ability of bush pilots to operate with relative impunity in whiteout shows that experience in flying under these conditions is useful.

B.5.9 Clouds

B.5.9.1 Causes and Types

Two basically different causes for clouds and their distribution can be distinguished:

(1) Dynamic causes. The vertical motion associated with pressure

TABLE 5-V HOURLY OBSERVATIONS

BARROW, ALASKA
6(BPOB) 636710B
DECEMBER 1964

REFERENCE NOTES FOR TABLE I

SKY CONDITION COLUMN –
Columns S_1, S_2, S_3, and S_4 present coverage of cloud layers in ascending order, in code. Refer to code table on following page.

CEILING COLUMN –
UNL indicates ceiling unlimited.
CIR indicates a cirriform cloud ceiling with height unknown.

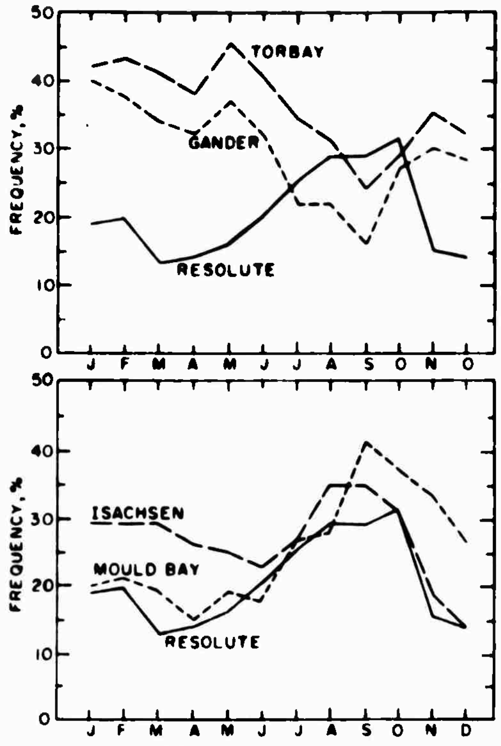
WEATHER AND/OR OBSTRUCTIONS TO VISION COLUMNS –
Refer to the corresponding columns in the code table on following page.

STATION PRESSURE COLUMN –
Station pressures apply to the official elevation of 13 feet above mean sea level.

WIND SPEED COLUMN –
Expressed in knots; multiply by 1.15 to convert to miles per hour.

Table 5-VI CODE TABLE FOR SKY CONDITIONS, WEATHER AND OBSTRUCTION TO VISION COLUMNS

CODE NO.	SKY CONDITION	W-1	W-2	W-3	W-4	W-5	W-6	O-1	O-2
1	Thin Scattered	Thunder	Light Rain		Light Snow	Light Snow Showers	Light Sleet	Fog	Smoke
2	Scattered	Heavy Thunder	Moderate Rain		Moderate Snow	Moderate Snow Showers	Moderate Sleet	Ice Fog	Haze
3		Tornado	Heavy Rain		Heavy Snow	Heavy Snow Showers	Heavy Sleet	Ground Fog	Smoke & Haze
4	Thin Broken		Light Rain Showers	Light Drizzle	Light Snow Pellets			Blowing Dust	Dust
5	Broken	Squall	Moderate Rain Showers	Moderate Drizzle	Moderate Snow Pellets		Hail	Blowing Sand	Blowing Snow
6			Heavy Rain Showers	Heavy Drizzle	Heavy Snow Pellets				Blowing Spray
7	Thin Overcast		Light Freezing Rain	Light Freezing Drizzle	Light Ice Crystals	Light Snow Grains			
8	Overcast		Moderate Freezing Rain	Moderate Freezing Drizzle	Moderate Ice Crystals	Moderate Snow Grains	Hail Pellets		
9			Heavy Freezing Rain	Heavy Freezing Drizzle	Heavy Ice Crystals	Heavy Snow Grains			
X	Obscuration								
-	Partial Obscuration								



Ceiling< 1000 feet and/or visibility < 3 miles.

Figure V-[illegible] FREQUENCY OF BELOW-VFR CONDITIONS AT FIVE ARCTIC STATIONS, 1954-1960[202]

Table 5-VII PERCENTAGE OF BELOW-VFR CONDITIONS IN VARIOUS WINDSPEED GROUPS AT RESOLUTE, N.W.T., 1954-1960[202]

	0-9 mph (0-4 m sec)	*10-19 mph (4½-8½ m 'sec)*	*20-29 mph (9-13 m 'sec)*	*≥30 mph (≥14 m/sec)*
Jan	1	8	39	87
Feb	5	18	39	80
Mar	4	8	40	87
Apr	4	12	29	67
May	14	13	20	42
June	22	21	13	15
July	28	26	15	12
Aug	33	30	18	20
Sept	32	28	26	13
Oct	20	25	41	72
Nov	5	10	36	80
Dec	2	7	38	79

systems favors cloudiness under cyclonic conditions and clear sky under anticyclonic conditions.

(2) Geographic causes. These usually work through the water or radiation balance. Moist surfaces with high evaporation rates favor cloud formation due to the high moisture content of the air, even under anticyclonic conditions. The opposite holds for dry surfaces. A strong positive radiation balance favors cumulus development independent, to a large degree, of the dynamic conditions.

A variety of cloud patterns will be caused by the interaction of these 2 influences. The main Arctic types are described below.

1. *Norwegian Sea Type* - In the zone of westerlies over the ocean all influences favor cloud formation. The cloudiness is high all through the year with a slight maximum in summer. It is noteworthy that, even in this area which is so clearly governed by cyclones, the maximum cloudiness is experienced in summer, the season with the lowest cyclone frequency (Fig. V-13b). The cloud types of this area also show peculiar behavior. In winter cumulus cover reaches a frequency of 30 to 40% - more than that observed in any season anywhere else in the Arctic (Fig. 5-14 and 15). Observations from the Norwegian coast indicate that the most important cloud in this group is cumulonibus. In summer this dominance disappears mainly in favor of altocumulus and altostratus (Figs. V-16 and 17).

The reason for this development is the frequent outbreaks in winter of arctic air masses over this area containing the warm water of the North Atlantic Drift. Extreme instability is created with a consequent development of convection cloud. As convectional clouds produce less cloudiness than do stratiform clouds, the cloudiness tends to diminish. Warm water and cold air masses combine to reduce the winter cloudiness below the conditions typical for the west wind zone.

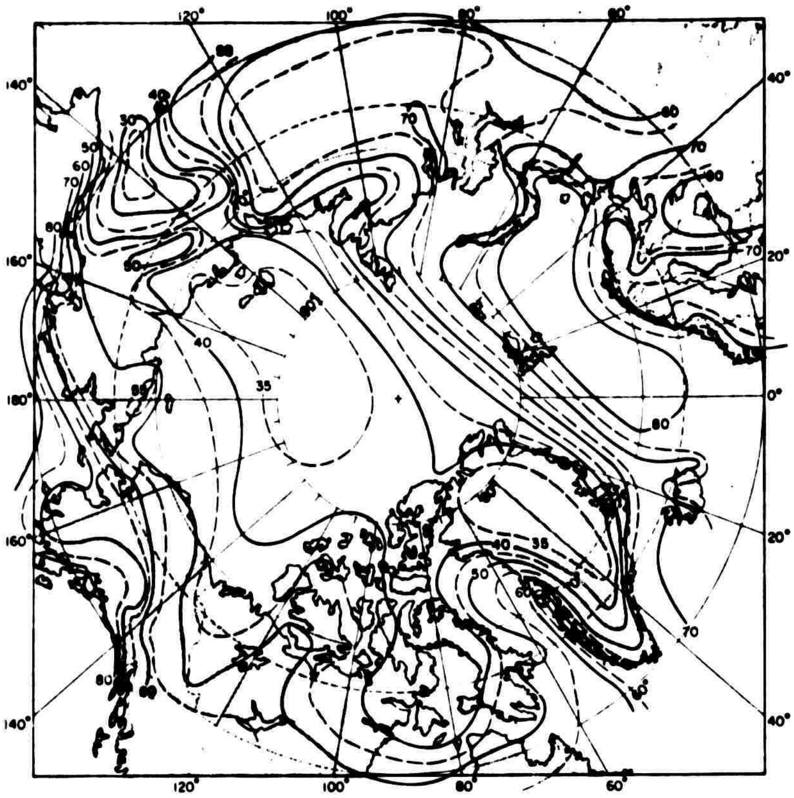


Figure V-13a JANUARY.

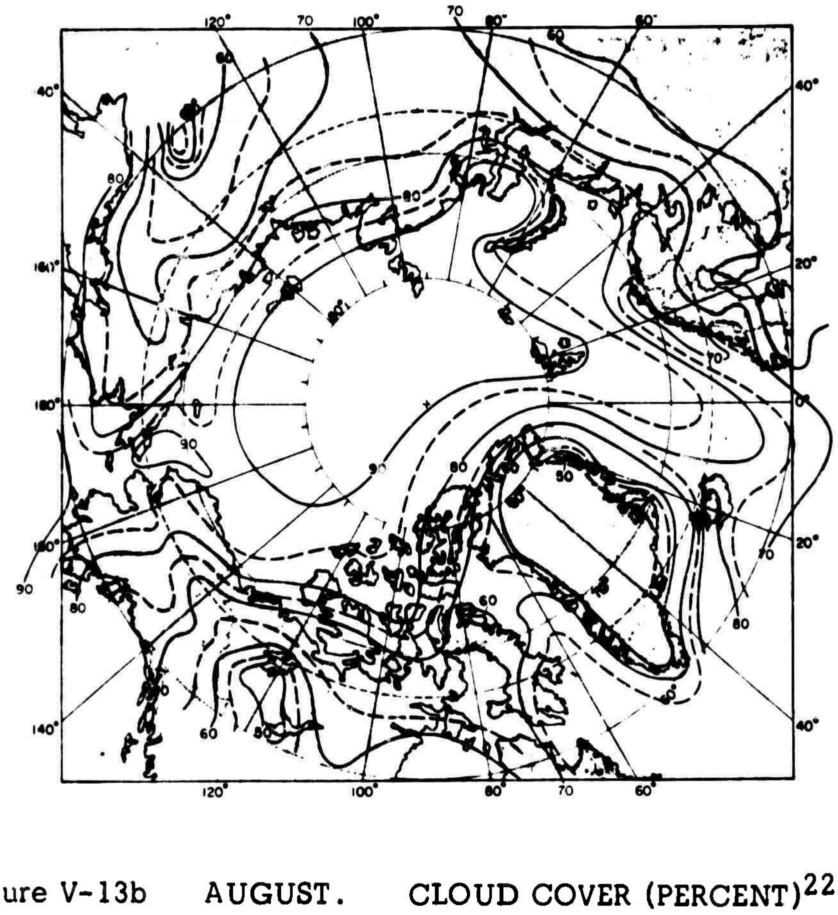


Figure V-13b AUGUST. CLOUD COVER (PERCENT)[229]

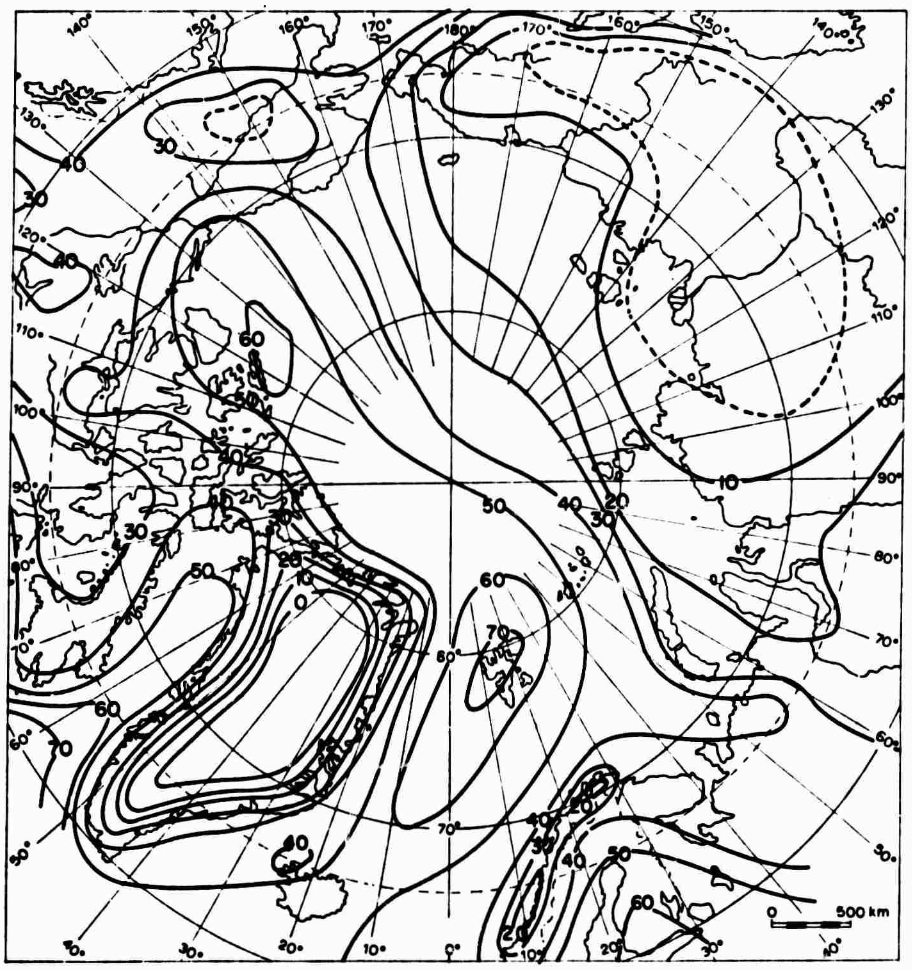


Figure V-14 FREQUENCY OF St AND Sc (%) IN WINTER

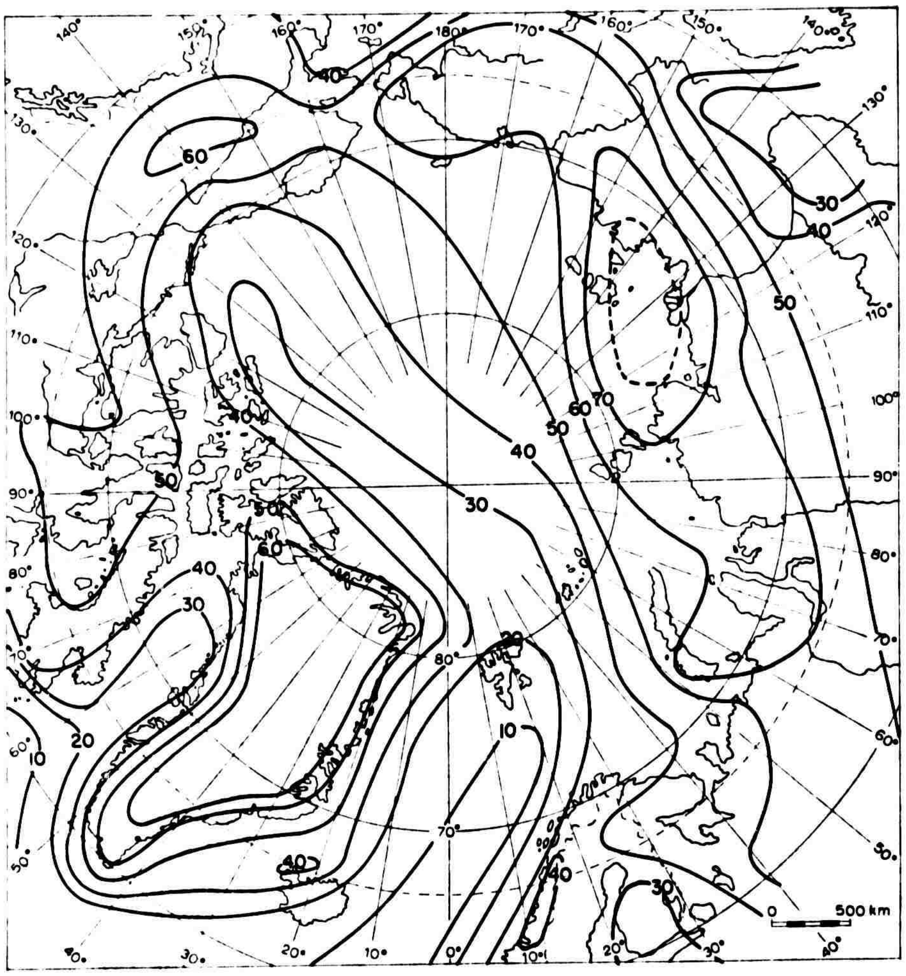


Figure V-15 FREQUENCY OF As AND Ac (%) IN WINTER

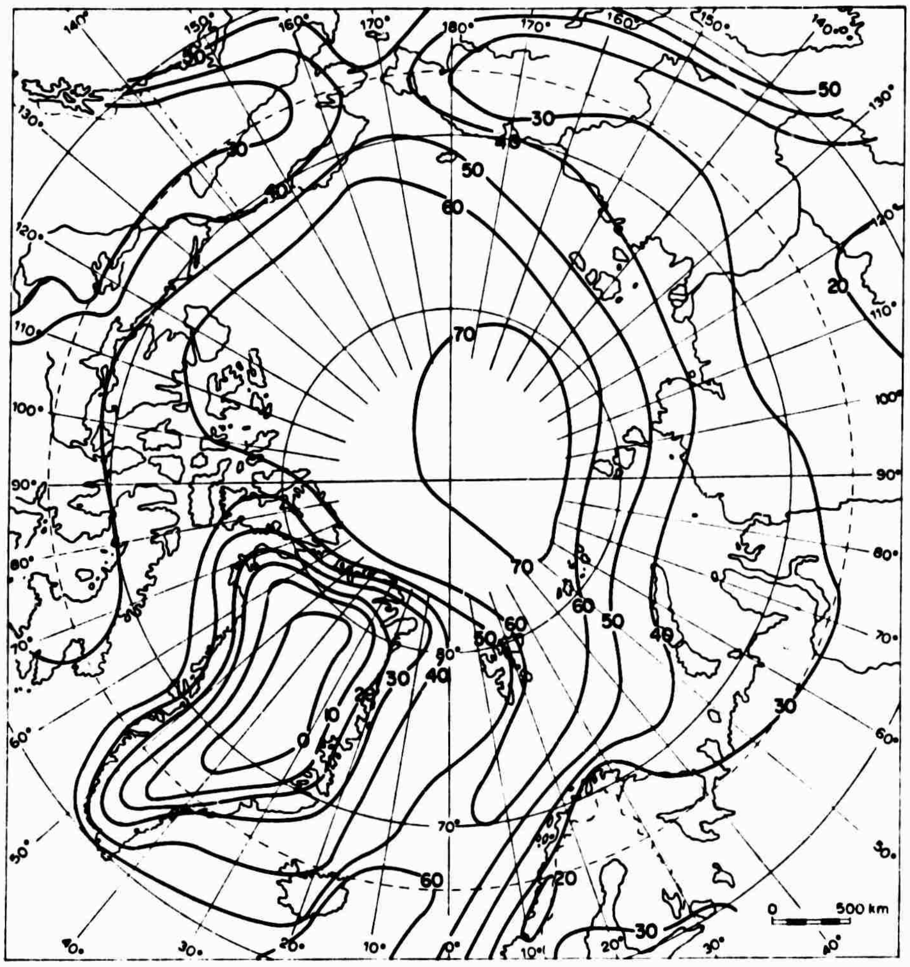


Figure V-16 FREQUENCY OF St AND Sc (%) IN SUMMER

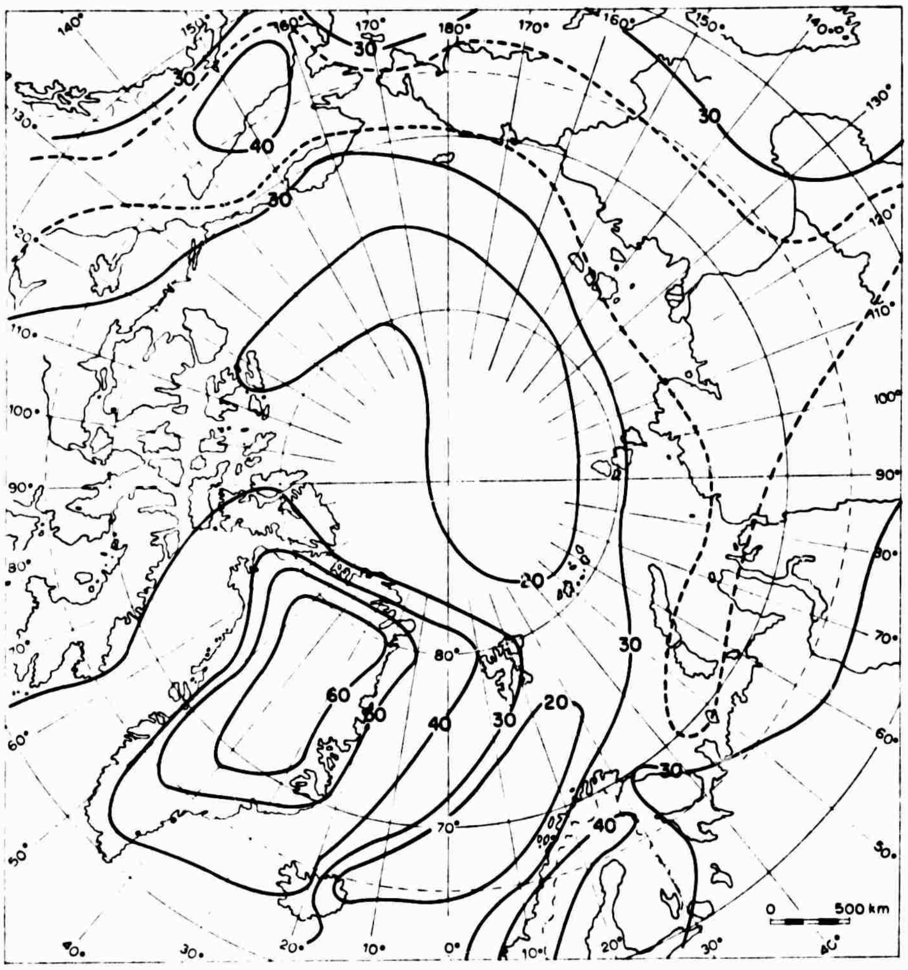


Figure V-17 FREQUENCY OF As AND Ac (%) IN SUMMER

In summer, with the water much colder in relation to its surroundings and with no extremely cold air available, the cumulus frequency and cloudiness approach the more normal patterns for the west wind zone - around 80% cloud cover. Due to reduced cyclonic activity, however, a relatively great proportion of the clouds is found in the medium cloud layers.

2. *East Siberian Type* - The whole of Siberia is dominated in winter by anticyclonic influences. The region of especially clear skies is found, not at 100°E, the zone of highest anticyclonic frequency, but nearer 140°E. The distance from sources of moisture must be responsible. Therefore, the characteristic stratus or stratocumulus formations under an inversion are conspicuously absent in this area. The clouds which exist are overwhelmingly cirrus (nearly 60%); even at the height of middle clouds the moisture content drops to low values, and only the infrequent cyclone is able to cause condensation.

The maximum cloudiness in the East Siberian type is reached in summer. Anticyclonic conditions are then much less frequent, and the Arctic Ocean has stretches of open water which serve as a moisture source. The frequency distribution of cloud types is rather uniform with a predominance of the convective types in the low and medium layers - a development characteristic of all continental arctic areas in summer.

The change from winter to summer conditions is gradual, while the decrease in autumn is abrupt. This difference between spring and autumn has little to do with the pressure distribution. The pressure curve (and also the temperature curve) is symmetrical in spring and autumn. The reason must be the availability of moisture. Because the snow cover is very slight in this area, the main moisture source during the spring and summer is the Arctic Ocean. Melting of the ice along the coast and the formation of puddles on the remaining ice is a gradual process, and accordingly the cloud amount increases gradually in eastern Siberia.

In autumn the freeze-over is quick, and a sharp decrease in moisture supply occurs rapidly.

It is noteworthy that this type is not as clearly developed in Canada. In spite of the large land area and the sheltering effect of the Rocky Mountains, the winter minimum is not as pronounced as in eastern Siberia. Contrary to the Siberian conditions, however, the increase during spring is quite rapid. This is not caused by extensive stratus layers as over the Arctic Ocean but by clouds of the medium types, which would indicate a dynamic reason for this development.

3. *Polar Ocean Type* - Over the Arctic Ocean the cloud amount is least in winter and spring and greatest in summer. Dynamic factors have least influence in this area. The polar frontal zone is, in fact, best developed in winter when the cloud amount in this area is low. During winter, the water content of the very cold air is too low for cloud formation, irrespective of the dynamic conditions. The most outstanding feature of the Arctic Ocean zone is the high frequency of stratus and stratocumulus in summer. It rises to 80% and stratus predominates in the majority of cases. The reason is the continuous cooling of air to the freezing point over the pack ice. These summer clouds are extremely uniform, extending as vast sheets over much wider areas than other clouds. The mean thickness of these cloud layers is 1,150 to 1,700 ft, a high value when the manner of their formation is considered. The water content of these clouds shows a pronounced decrease from the coast toward the pole. A further characteristic of this area is that the seasonal change in cloudiness is restricted to a very short transitional period, while rather constant cloudiness prevails during the rest of the year. Similar abrupt changes are found only in monsoon areas with their complete changeover in circulation pattern.

B.5.9.2 Occurrence

Winter is the season with the greatest mean deviation from latitudinal average. An outstanding winter feature is the zone of high cloudiness in the Norwegian Sea. It extends along the Eurasian coast far into the Arctic Ocean, until it gradually loses its identity. From its center line - Thorshavn south of Bear Island - where the described phenomena of the zone are most conspicuous, it changes its characteristics toward the east in the Arctic Ocean. Cumulus clouds disappear in the Barents Sea, and stratus and stratocumulus become less frequent. From the Laptev Sea eastward, middle cloud layers are most important with altostratus being especially frequent. With practically no moisture supplied from the frozen ground, cyclones penetrating to the east are only able to cause condensation in the higher levels. From the Norwegian Sea northwestward to the sea ice margin cumulus is replaced by stratus and stratocumulus as the water becomes colder and the time available to create unstable air becomes shorter.

Scandinavia and the western RSFSR show very high cloud amounts, with low clouds dominant. Medium clouds become dominant in western Siberia, much farther west than over the Arctic Ocean. They are replaced by cirrus from central Siberia eastward. The reason for this rapid thinning of clouds over the continent is the position of the Siberian anticyclones which divert the disturbances toward the Arctic Ocean.

Across the mountains of northeast Siberia toward the Pacific, the clear skies are replaced by very great cloud amounts with a dominance of low-cloud types as the Bering and Okhotsk seas are reached. No such clear pattern develops over the American continent where, apart from minor irregularities, the main feature is the gradual decrease in cloudiness to the north. There is a zone with few low clouds along the 65°N parallel, with a minimum in the interior Yukon valley. Mountain ranges there hinder free flow of air in the meridional direction. Another minimum occurs over the Canadian Arctic Archipelago where the lowest cloud amounts are observed. This same zone is characterized by a maximum of the medium-type clouds and has a much higher frequency of altocumulus than in the eastern hemisphere.

The influence of the Aleutian low and the Pacific frontal zone is hardly noticed in the Arctic. The mountain ranges form an effective barrier; and over the frozen surfaces of northern Canada the air does not pick up sufficient moisture, after the descent to the plains, to show another increase in cloud amount.

With the approach of summer, the Atlantic zone loses much of its individuality. It appears merely as an extension of the large area of cloud now covering the whole Arctic Ocean. It can be distinguished from the latter region only by the higher frequency of cumulus-type clouds. Although the reasons for the formation of these zones are different, in appearance there is little on which to base a distinction between the two, both having extremely dull and monotonous skies. At the continental shores both influences come to a rapid end. The farther south the location, the less cloudy are the skies and the more are the stratus and stratocumulus replaced by cumulus. As over the Arctic Ocean, the maximum cloudiness over the continent is reached in summer; but mean cloud amount remains lower, and a much greater diversity of cloud types exists.

The cloud maximum in summer is a peculiarity of the Arctic and contrary to that found in mid-latitudes. The reason does not lie so much in the different dynamic conditions as in the ground influences. In the Arctic the winter ground is frozen everywhere; and, because of the absence of sufficient incoming radiation for evaporation, the supply of moisture to the atmosphere is limited. In summer the incoming energy and the extensive moist surfaces permit high evaporation and cloud formation. In middle latitudes solar energy is received also in winter and so evaporation is not suppressed to the same extent. In summer the incoming energy is high, and the evaporation will dry the soil causing the relative humidity of the air to be lower than in winter and the cloud amount to be less.

The seasonal variability of cloudiness increases from south to north. This is a result of the increasing influences of the peculiar cloud regime of the Arctic Ocean. Latitudinal means of cloud-type frequency show that the North also has the highest seasonable variability of cloud type. The farther north, the greater the cloudiness in summer and the more frequent the low and dense cloud types. In winter, on the other hand, the cloudiness diminishes northward and the frequency distribution of types is more uniform.

A complete changeover of the north-south gradient in cloudiness is produced by this high variability in the north and the rather stable conditions in the south. During the winter there is a continuous decrease to the north, while an increase is found in summer. The changeover takes place very rapidly with the result that only during the short transition periods in spring (May) and in autumn (October) is there but little difference in the average cloud amount over the whole arctic sector.

B.5.10 Precipitation

Snowfall is light in the Arctic, and stations in continental areas have maximum precipitation in late summer, i.e. a large part of the annual total falls as rain. The main precipitation over the Arctic Ocean is frontal in nature. Partly for this reason the annual amounts decrease northward (Fig. V-18). Along the margins of the Arctic Ocean the winter accumulation is, for the most part, less than 10 in; the Siberian and Canadian arctic coasts receive as little as 5.5 in per year (including rain and the water equivalent of snow). The annual precipitation over the Arctic Ocean is meagre - about 5 in. It is mainly in the form of snow and falls during autumn and late spring. The minimum precipitation occurs in winter. The cycle, and the paucity of precipitation, is

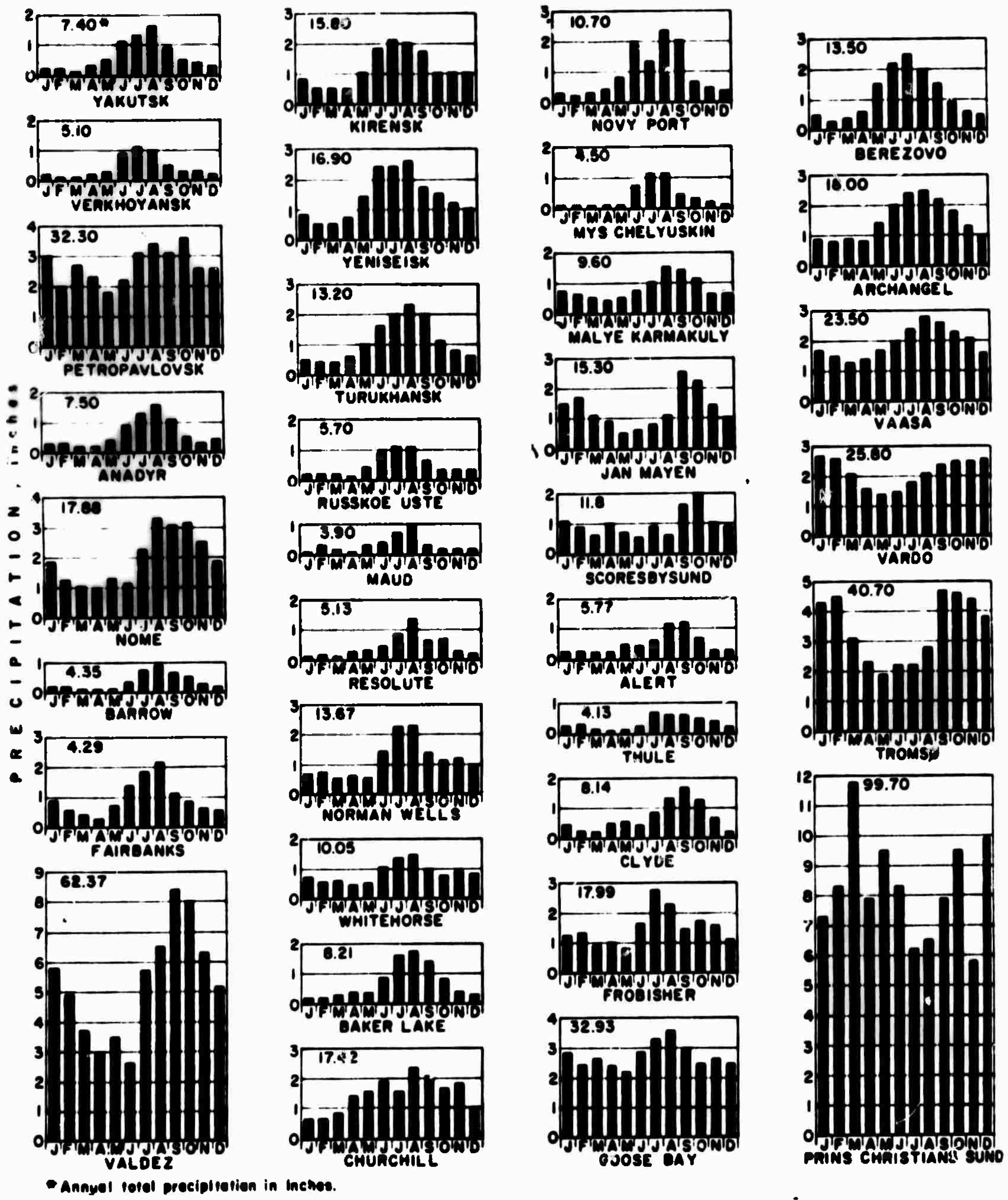


Figure V-18 MEAN MONTHLY AND ANNUAL PRECIPITATION AT SELECTED STATIONS. (SNOWFALL HAS BEEN REDUCED TO WATER EQUIVALENT, 10 IN. SNOW = 1 IN. WATER.

caused partly by the low moisture-holding capacity of cold air.

Although the amount of precipitation along the arctic coasts is small, the ground remains saturated for a long period in summer. Underground drainage is prevented by the permanently frozen subsoil. Evan though somewhat larger amounts of precipitation are received farther south at inland stations, the summer temperatures there are so much higher that drought conditions are manifested.

The characteristics of the snow cover, such as thickness and duration, have important climatic effects on the heat and moisture exchange at the surface. Over the central Arctic Ocean the snow cover becomes established in late August. The thickness may be about 14 to 16 in by late spring. Steady snow melt - caused by solar radiation - usually begins by the middle of June; and, while there are marked differences from year to year, the ice is usually snow-free by the middle of July.

B.5.11 Regional Climates

B.5.11.1 Northeast Atlantic

The climate of the northeast Atlantic is dominated by the massive northeastward transport of air winds from lower latitudes. In the southern part of the Norwegian Sea, winds are predominantly from the west. Along the Polar and Arctic fronts, which form the contact between warmer air from the south and the colder air in the north, there is an almost continuous stream of cyclones which move eastward and northeastward in consonance with the flow of the jet stream. The latter typically passes from the east coast of North America across Iceland to northern Scandinavia. In winter the air is unusually warm for the latitude while in summer its effect, at least in the more southerly positions, may be cooling. In spite of its great strength, it is not unusual for this characteristic meridional circulation to shift to zonal flow. When this occurs there are periods of a north-south interchange of air masses, which gives striking temporary anomalies. This was particularly conspicuous during winters in the early 1940's when Scandinavia experienced extreme cold conditions while those in Iceland at the same time were far milder than average.

East Greenland and Svalbard lie on the north side of this circulation and, especially in winter, experience cold conditions. The maritime climatic element of the open Atlantic fails to penetrate the coastal seas of East Greenland and Svalbard in winter when the pack ice is virtually continuous. The differences around the north Atlantic not only are regional but also are more local between the coasts, commonly with much fog and low cloud and very much clearer and calmer conditions in both winter and summer than prevail inland.

The greatest climatic amelioration in the Arctic is in northern Scandinavia as a result of the warm North Atlantic Drift. Prevailing westerly winds bring relatively mild humid air into the region. At open coastal localities, on islands, and over the southeast section of the Norwegian Sea, winter monthly mean temperatures are rarely below the freezing point and frost is likely to occur only on about one day in two in the middle of winter. Near Nordkapp temperatures are marginally lower. At the heads of the fiords, the daily mean temperature is 14°F. Inland temperatures are far more extreme; and at Karasjok, Norway, on the Finnish border, the lowest temperature ever recorded was -60°F. Summer temperatures are similar if taken as contrasts. At that season the coasts are cool (mean 50°F) while in the interior it is 5°F higher and extreme temperatures may reach 84°F. Annual precipitation, which on the coast may be in excess of 40 in and in the mountains of the Lofoten Islands reaches 100 in, falls to less than 20 in in the interior.

The climate of Iceland resembles that of northern Scandinavia, with normally mild winters and cool summers. All coasts have winter temperatures about freezing point and July monthly means of 50°F. It may be slightly cooler in the north and east while the interior exhibits a significant temperature range. Precipitation in most coastal areas exceeds 40 in and reaches a maximum in the southeast, where near the sea it exceeds 80 in and on the higher slopes of the glaciers is probably more than 150 in. In the interior and on the northeast coast, drier conditions prevail. Near sea level, in the south and west of Iceland, snow rarely remains for any extended period, although in the center and northeast sectors it is normal in winter. In coastal areas, especially in the south, relative humidity is high throughout the year. Fogs are not frequent on most coasts, although they may provide some hindrance to navigation, especially in the northeast and east. Off the east coast there are about 55 fog days a year while in the vicinity of Reykjavik there are about 10.

In several ways the climates of Svalbard and Björnöya are far harsher than those described for the Scandinavian sector of the northeast Atlantic. However, the effect of the high latitude is somewhat ameliorated by relatively warm Atlantic water that flows north along the west coast and turns eastward along the north coast of Spitsbergen. Mean annual temperatures near the west coast are about 48°F. During summer the temperature rises above the freezing point and the mean July temperature is about 40°F. Occasional days with temperatures up to 68°F may occur at the heads of valleys, particularly when warm fohn winds blow from the interior. The coldest months are February and March when the mean temperature is approximately 0°F. Temperatures above freezing point, however, may occur in any month; and midwinter thaws are not uncommon. In contrast, temperature variations in summer are small.

The east and northeast of Svalbard have a climate that is more continental than the west coast. Winter cold is believed to be greater and the summers are cooler. Throughout the archipelago weather is highly variable because of the passage to the south of cyclonic disturbances. Precipitation is low, but in general the west coasts receive more than the interior and the east coasts. The proximity of warm and cold seas and cold land result in frequent fogs along the north and west coasts. Björnöya, with 82 days of recorded fog in the year, has long periods of continuous fog during the summer months - there are only 6 clear days a year. The lowest temperatures are in late winter (the March average is 14°F), the highest (38°F) in July.

In East Greenland there is a gradual change from maritime climatic conditions near Kap Farvel to extreme continental conditions in the northeast. These changes are a consequence of the position of the land in relation to the normal track of cyclonic disturbances. In the extreme south the coldest month is normally January with a mean temperature of 25°F. The coldest period is progressively later in winter farther north, with mean temperatures falling to -22°F in Peary Land. Summer temperatures are more uniform along the coast although in the far north the presence of ice offshore and the ice sheet to the west hold temperatures to a maximum of about 41°F during the middle of July. Precipitation similarly varies from south to north. It is exceptionally heavy in the south of Greenland where it exceeds 100 in; northward along the coast it diminishes rapidly to less than 40 in on the south-central coast, less than 15 around Scoresby Sund; and true arid conditions of less than 6 in occur in the far north. In the south most of the precipitation falls as snow during the early part of the winter followed by a second maximum at the end of the winter.

B.5.11.2 Eastern North America

Eastern arctic America extends through more than 20° of latitude, and consequently both the duration of daylight and the climate show important differences between the northern and southern parts. In winter, south Baffin Island and Greenland are under an extension of the Icelandic low, and this has the effect on the Canadian side of bringing down vast quantities of cold arctic air from the north and northwest. The passage of these disturbances makes for variable wind direction. Temperatures at this season are uniformly cold on the Canadian side of Baffin Bay with average temperatures in January far below freezing at all points and ranging from 2°F near Resolution Island to -22°F on northern Baffin and Bylot islands. In contrast, the Greenland coastal temperatures are appreciably warmer from about freezing at Kap Farvel to 0°F in Disko Bugt. Winter in the whole of the southern sector tends to be a stormy period with many days with snow. Abnormally high temperatures may be found on the Canadian side when a low becomes stationary in the Labrador Sea in winter and large quantities of warm moist air are forced over southern Baffin Island. In consequence it is not unusual for rain to fall in every month on the coast.

Farther north the arctic continental influences become more pronounced and the differences between the Greenland and Canadian sides tend to disappear. From the Parry Channel northward incursions of warm air in winter are rare and there are long periods of comparative calm and extremely low temperatures. Mean temperatures in this period are about -35°F and minima below -58°F. The excessively cold period is broken only by short warmer windy periods when the temperature may rise to 5°F as warmer air is advected into the region. Snowfall is remarkably low, and at stations such as Eureka and Alert total snowfall may be less than 2 inches a month.

In summer the weather is far more uniform over the whole Baffin Bay region than in winter. At coastal sites the dominant influence is the cold sea; fog and cloud are normal and temperatures rarely rise above 59°F. The mean temperature in July on the coasts is normally about 40°F. Away from the sea temperatures rise quickly. At the heads of many of the fiords, in the calm conditions provided by the shelter of the mountains and with continuous sunlight, warm dry days may occur throughout the summer, in pleasant contrast to the cold damp of the coastal sites. At this season precipitation at sea level is almost entirely in the form of rain. In the south of the area 10 inches is not uncommon on the

Canadian side while on the Greenland coast a total precipitation of over 40 inches may occur. However, at both the Canadian and Greenland sites summer and annual precipitation decreases rapidly from south to north. Beyond 80°N total precipitation may be less than 3 inches and the climate is truly that of a desert. In many coastal areas where fiords divide the mountains and lead from ice caps, strong local winds are a common element in the complexities of the climate. These winds may be hot, and they bring a delightful change to the damp summer weather.

The weather of the eastern Canadian Arctic results from the complex interaction of continental air masses with oceanic influences originating over the North Atlantic. Although the coastal climates generally grade smoothly from one part of the region to another it is possible to recognize three climate subregions: the northern continental interior, the Labrador coasts, and a southeastern sector.

From January until June the northern interior may be considered part of the continent. At this season there are long periods of constant northwesterly winds which bring low temperatures (-25°F) and intense wind-chill to the whole area, Snowfall is not heavy but blowing snow is a frequent occurrence. Occasionally in winter warmer air penetrates from the Atlantic but this is not common. The summer weather in coastal localities is dominated by the proximity of cold seas. Only in the Hudson Bay lowland and near James Bay do mean temperatures rise above 40°F. Along the south side of Hudson Bay there may be long spells of warm summer weather with offshore winds; but elsewhere fog, low cloud, and drizzle characterize the weather of much of the summer.

The main climatic differences between the Labrador coast and the interior area result from the closer proximity to the low-pressure systems that develop or deepen in the Labrador Sea and Davis Strait. The effect for part of the winter is to carry Atlantic air over these coasts; mean temperatures are consequently about 20°F warmer than on the Hudson Bay and other interior coasts. Snowfall is significantly higher and the weather is much stormier. The differences in summer are not as great and the climates are essentially the same except for greater precipitation and somewhat stormier conditions on the Labrador coast.

The increase in winter temperatures and in precipitation at all times of the year, which was noted on the Labrador coast when contrasted with Hudson Bay, becomes even more pronounced in the southeast sector. However, long periods of northwesterly air flow in winter continue to keep the temperature abnormally low (20°F) for coastal localities in mid-latitudes. Everywhere mean January and February temperatures are below freezing point. Precipitation in the form of either rain or snow is much higher (5 in) than farther north at this time of the year. The arrival of spring is retarded by the cold water of the open Atlantic where fog and low cloud are common.

The central and western parts of the Queen Elizabeth Islands show a remarkable uniformity of climate due to the absence, except along the eastern margin, of any considerable relief. Also, for the greatest part of the year the Archipelago can be considered as a single land mass because of the frozen condition of the straits. During the summer months local conditions vary because of the presence of small quantities of open water. In winter the continuous darkness and loss of heat in the lower layers of the atmosphere lead to extremely low mean temperatures which average -35°F during the months of January, February, and early March. Extremely low temperatures, such as are found in some of the sheltered valleys of northern Siberia and northwestern North America, have not been recorded in the Canadian arctic islands where the absolute minimum at most stations has been about -60°F. Unlike many other areas in the Arctic, in winter there is no source of relatively warm air within many thousands of miles; and consequently, the temperature never rises above freezing point in the period from October to April. The winter is therefore a period of continuous cold broken only by brief warmer periods when winds upset the inversion in the lower layers of the atmosphere. It is essentially a period of clear skies, with as much as a third of the time cloudless, although ice crystals and low visibility may occur. Fogs are rare except in the vicinity of settlements and at one or two points at which open water occurs, particularly in late winter, such as the vicinity of Barrow Strait. The winds during this period are predominantly from the north and northwest, but they generally are not strong and calm periods are common. The absence of any major source of moisture and the extremely low absolute humidity preclude heavy precipitation during the winter months; and it is unusual to find more than a few inches of snow on the ground, although drifting into hollows is naturally considerable.

Summer is a period of cloud and fog as it is with so many other maritime arctic areas. Temperatures rise above freezing point in June, and during July the mean temperature is about 40°F. Extremely high temperatures are unknown, and absolute annual maxima of 55°F are probably typical figures for all coastal areas, although it may be assumed that in the interior of some of the larger islands higher temperatures are achieved. Despite the fact that temperatures rise above freezing point, there is a 50% chance of frost occurring on any day even in the middle of the warmest period. As the channels between the islands open up, fogs and low cloud become widespread. Precipitation is normally in the form of rain although snow may occur in any month of the year. The total precipitation throughout the islands is remarkably low - between 3 and 5 in of equivalent water. The islands are one of the most arid areas of the world; their moist appearance during the brief summer period is due to the permafrost. Even in this respect, however, soil moisture is significantly lower than in southern arctic areas. The maximum precipitation occurs in the middle of summer, occasionally falling as extremely heavy showers. Snowfall is normally heaviest in the early fall, when there are still considerable amounts of open water in the area. It decreases in October; and there is then a period of extremely light, often negligible, snowfall throughout the winter, until in many localities a second maxima occurs in the month of May.

B.5.11.3 Western North America

The winter climate of the southwestern archipelago is also typical of a continental arctic area as all the channels are frozen over by late November or December. In winter the area is situated between anticyclonic highs over the Mackenzie and Yukon and the low-pressure systems of Baffin Bay and the Icelandic low. In all except the extreme west of the region, winds are dominantly from the northwest and blow with remarkable constancy. The frequency of northwesterly winds and the wind velocity in winter increase toward the east. Temperatures are uniformly low with mean temperatures remaining roughly constant in January, February, and March - between 18°F in the southwest and -25°F in the northeast. Cloud amounts and snowfall are low, although blowing snow is often a problem. Such cloud and precipitation as there is probably forms from shallow low-pressure systems moving east across the islands from the Beaufort Sea.

Summer conditions show great contrasts in the area. At this time of the year the mainland coast is periodically under the influence of continental air masses from the south; and long periods of warm clear weather may be expected, particularly in the west and in the inner part of Bathurst Inlet. Mean July temperatures are 50°F or higher and occasional temperatures in excess of 75°F are not uncommon. North of the mainland, however, the channels have opened and temperatures close to the freezing point are common in areas adjacent to the sea except for the shallowest and most southerly waters. Consequently, in summer there is often great contrast between the pleasant warm southern sectors with the occasional summer showers and the north where temperatures remain about

40°F and there is fog and low cloud. At this time of the year, disturbances pass directly through the area bringing long periods of low cloud and cold weather, especially to the northern part. Everywhere summer is the period of maximum precipitation (4 inches). In the southwest of the area it may fall in comparatively short periods from rather violent local storms while in the north it often occurs as drizzle and light rain spread over several days.

Spring arrives abruptly on the western mainland coast, commonly helped by incursions of warm southwesterly air from the uplands of the interior which is warmed in the process. Fall is a stormy period with strong northwesterly winds, squalls, and snow flurries. With the freezing-up of the channels these quickly abate and winter sets in again.

The whole coastal zone of the Beaufort Sea exhibits an arctic climate characterized by long dark cold winters and short cool summers. During the winter there is remarkable uniformity of temperature throughout the area, with mean daily temperatures of about -20°F from late December to early March. At most localities February is the coldest month; and generally the northeast part of the area, Banks Island, is about 5°F cooler than elsewhere. Extreme winter minima may fall below -10°F on the coast and -60°F inland. Snowfall is light and average snow depths along the coast in late winter may be less than 10 inches although there are deep snow drifts. At Sachs Harbour on Banks Island, close to the tree line, snow depth is about 25 inches. The first snow cover usually settles on northern Banks Island by 20 September, and by the end of the month all the area has a 1-inch snow cover at least. The median number of days with snow on the ground varies from 275 on northern Banks Island to 235 in the Mackenzie Delta. Winter winds tend to be either from the northwest or the east and southeast, although near the mountains there is considerable topographic control. On the coast the average wind speed is about twice that of inland sites.

In summer, climatic contrasts between the coast and the mainland interior, especially the Mackenzie Delta, can be very great. The Delta and adjacent coast have the warmest summer temperatures on the North American continent for the latitude. The average July temperature for the central Delta is 57°F, and the close spacing of the isotherms along the coast emphasizes the difference between the warmer mainland and the cooler, still partly ice-filled sea to the north. The average July temperature along the mainland coast is 40°F, although freezing temperatures can occur at any time, Extreme summer maxima along the coast are 65°F. At inland sites 80°F is not uncommon and the highest temperature ever recorded at Aklavik is 93°F. The total annual precipitation is between 5 and 8 inches along the mainland coast and only 4 inches at Sachs Harbour. The region would, in fact, be a desert were it not for the permafrost that restricts drainage and low temperatures that limit evaporation. As it is, the warm summers of the Mackenzie valley draw the tree line far north and in the Delta spruce reaches practically to the sea.

The weather of the Chukchi Sea develops from the interaction between air masses developing over the northern Pacific, which are characteristically maritime, and the continental air masses that form in winter over the northern continents and the polar pack. In summer the latter air mass retreats over the polar basin and, at least in its lower layer, is moist.

In winter the frontal zone dividing the two air masses lies across the southern Bering Sea and the Aleutians; the main cyclonic storms consequently remain far to the south of the Chukchi Sea and winter is a period of extreme cold with mean temperatures in the coldest month of about 0°F along the coast of Seward Peninsula. Farther north monthly winter temperatures are significantly colder, descending below -30°F in the northwest (Mys Shmidta) and -20°F in the northeast (Barrow). These extreme northern conditions are associated with lower cloud amounts than farther south. The winter cold is physiologically intensified by unusually strong winds, particularly in the south of the Chukchi Sea.

In summer a much-weakened polar front retains its average position south of the Chukchi Sea while a second frontal zone lies along or north of the shores of the continent. Both are associated with frontal disturbances, but in summer they are weaker and their movements are more irregular. Local weather in the summer is closely related to distance from the sea. Open coastal and marine areas experience cloudy foggy and cool conditions with mean July temperatures in the range of 40 to 45°F; inland mean temperatures are higher (50°F). Absolute maxima are considerably higher and there is less low cloud and fog.

Precipitation is apparently higher on the Alaskan side than it is in Siberia. On the North American coasts annual totals are from 8 to 12 inches while in northeastern Siberia they are 6 to 8 inches. Maximum rainfall in all areas is in late summer and heaviest snowfall is in late fall.

B.5.11.4 North Pacific

The Bering and Okhotsk seas have climates that may be characterized as maritime polar. There are, however, great differences between various sectors and especially between the shores of the largely land-locked Sea of Okhotsk and more open Bering Sea. Climatic conditions are most uniform in the summer. Practically everywhere on the continental mainlands, summer (July) temperatures average above 50°F except in the Anadyr-Chukchi areas where they only reach about 45°F. On almost all the islands similar low temperatures occur. A few miles inland, away from the direct cooling influence of the sea, summer temperatures are appreciably higher and commonly average above 55°F with extreme maxima for a summer often 15 to 18°F warmer inland than on the coast. At inland sites in in Alaska and everywhere in Siberia summer is the period of heaviest precipitation. At island localities, especially the Aleutians, summer is often a period of continuous fog and cloud.

The summer uniformity of the coastal climates breaks down after September with the onset of cold weather. Winter is marked by the appearance of two contrasting climatic elements - intense cold and essential stability of weather in the continental interiors and the continuing relative warmth of the open and partially open sea. As a result of the prevailing and often strong northwesterly circulation on the Siberian side, intensely cold winds reach the coast; and mean January and February temperatures are below -5°F, except in southern Kamchatka, southern Sakhalin, and the Kuril Islands. The Kamchatka Peninsula and Kuril Islands are particularly stormy in winter and snowfall may be heavy. On the Alaskan side of the Bering Sea temperatures are 10 to 20°F higher. The greatest difference, however, is in the Aleutians where winter temperatures are barely below freezing point and are accompanied by heavy snow and rain.

On both sides of Bering Sea annual precipitation decreases rapidly from south to north and from the coast inland. The maximum annual precipitation in the Aleutians exceeds 80 inches, most of which falls as rain; but in the Bering Strait area the maximum is 8 inches. Precipitation falls for approximately 210 days a year in the Pribilov Islands and about 80 days at Bering Strait. Similarly on the Asiatic coast, precipitation increases from less than 5 inches in the north to more than 40 inches in the Kuril Islands.

B 5.11.5 Northern Siberia

The climates of the Laptev and the East Siberian seas are similar because of their remoteness from the warming effects of the Atlantic and Pacific oceans and also because of the influence of the northern mainland of Siberia, the coldest region in the northern hemisphere in winter. Precipitation is low, averaging from 6 to 9 inches annually. A few areas, including the Lena Delta, receive only 3.5 inches. In spite of these small quantities, the summer weather gives an impression of permanent dampness with high humidity and much cloud and fog. During winter, long periods of cold clear and practically clam weather prevail. Occasional storms travel eastward through the area bringing strong winds for 3 to 5 days. They are associated with light snowfall and reduced visibility resulting from the fine snow being picked up by the wind. Average temperatures in January and February are about -20°F and are marginally warmer over the sea ice.

The Novosibirskiye Ostrova are deserts, receiving only about 3 inches of precipitation annually. However, the relative humidity is high and a thin snow cover lasts for 9 months of the year. The mean temperature of the coldest and warmest months are -22 and 37°F respectively. Ostrov Vrangelya experiences stormier conditions than the mainland, particularly in the early winter. Easterly winds prevail in summer which is a period of continuous fog and low cloud.

The Kara Sea and the West Siberian Lowland have the same climate although there are often considerable differences in weather from west to east because of the occasional incursion of Barents Sea conditions into the western part and the colder Laptev Sea air masses to the east. In summer, the islands and the seas are dominated by the low surface temperatures of the sea. Consequently temperatures are low (32 to 39°F), there are long periods of continuous low cloud, and fog is common. This is the season of maximum precipitation but, while light rain may be recorded on numerous occasions, amounts are low. Inland, summer conditions improve with less cloud and higher temperatures although the effect of the long wide estuaries is to carry marine influences far inland.

Winter lasts from November to May. Temperatures are coldest in January and

February when at some localities thaws have never been recorded. At Ostrov Belyy temperatures in January and February average -11°F and at Ostrov Kikson -13°F. Lowest absolute temperatures do not go below -58°F and are 9 to 27°F higher than in the subarctic interior of Siberia. About 2 weeks after the temperatures sink below freezing there is a continuous snow cover that later in the season reaches a maximum depth of 10 to 30 inches depending on the situation. The coast is covered with snow in the first part of October and this lasts for 270 days. Precipitation decreases from west to east. The average per month is 0.1 to 0.3 inches. Large snow storms are accompanied by a rise in temperatures, and blizzards (purga) of snow blown along the surface are common with high winds.

The climate of Severnaya Zemlya is even more severe than on the mainland; this is so in spite of the fact that the winters are warmer. Summer temperatures rarely rise more than a few degrees above freezing point. Annual precipitation is about 4 inches. Winds are moderate and generally from the northeast, with maximum velocity in September and minimum in April and December. There are nearly a hundred days with snow storms in the winter, and snow cover lasts from the end of September to the beginning of July.

The Barents Sea has a typical polar maritime climate. Storms are frequent the year around and visibility is usually poor. Cloud cover averages 75% but is 95% at 75°N which is one of the cloudiest regions in the northern hemisphere Extreme maritime conditions become modified away from the Norwegian Sea and temperature and precipitation decrease to the east and north, while both show increasingly continental characteristics near the mainland.

The first signs of winter appear on the mainland in the form of snow and frost between late September and mid-October as the days shorten rapidly. This is the stormiest and cloudiest time of the year. Winter lasts from November to March. On the western mainland snow cover lasts 200 days - in Novaya Zemlya more than 240 days. Over the sea, winter temperatures are below freezing point although extreme cold is rare. In midwinter the Barents Sea area is 21 to 36°F warmer than the average for the latitude. On the land, mean January temperatures decrease steadily from west to east and average close to, or below, 0°F for the 3 months from January to March along the eastern margin of the Barents Sea.

The change to summer conditions begins in April and is marked during May and early June, which is a second period of much cloudiness and fog. Mean July temperatures are everywhere below 50°F except on the mainland away from the coast. In the east and northeast of the Barents Sea average temperatures barely reach 40°F and the highest temperature in the year rarely reaches 43°F.

Extreme storminess prevails on Zemlya Frantsa Iosifa which has a high frequency of cloudiness, fog, and wind. Snow and freezing temperatures must be expected at all times of the year. Conditions are slightly easier in Novaya Zemlya, although just as changeable especially in the south. Unusually strong, squally winds burst out from the valleys and fiords of Novaya Zemlya especially in the south and west. The winds may last from 1 to 5 days and reach several miles from the coast.

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B.6 TERRAIN

B.6.1 Geology

B.6.1.1 Landforms

The geology of the Arctic is not uniquely arctic since the earth's crust has migrated through geologic time. The major topographical elements of northern lands can only be understood by a reference to their geology. Their structure is roughly symmetrical about the Arctic Basin and has been instrumental in modelling the major elements of the surface form.

Three ancient shields formed predominently of Precambrian granites and gneisses are the main structural features of the land surfaces(Fig. 5-1). They are the Canadian-Greenland Shield in the western hemisphere, the Scandinavian Shield in northern Europe, and the Angara Shield in north-central Siberia. In many sectors the shields extend beyond their exposed margins but are buried under younger flat-lying sedimentary rocks. A large part of the North European Plain in Russia and smaller areas in the Mackenzie Lowland of northwest Canada developed in this way. The Angara Shield differs from the other shields in that only a small area is exposed; the remainder is buried under deep sediments which have experienced much faulting and in places have been linked with igneous activity. Between the stable shields and associated platforms are belts of strongly folded sedimentary rocks which are still today, in many areas, mountain ranges. Where the folding has been unusually intense and is of great age, as in Svalbard and northern Scandinavia, the physiography often differs little from the shield areas. In contrast, the younger fold belts, including those in the northwestern Queen Elizabeth Islands, the mountain ranges of Alaska, and the Kamchatka Peninsula, have developed scenery that is typically alpine in many places. One orogenic

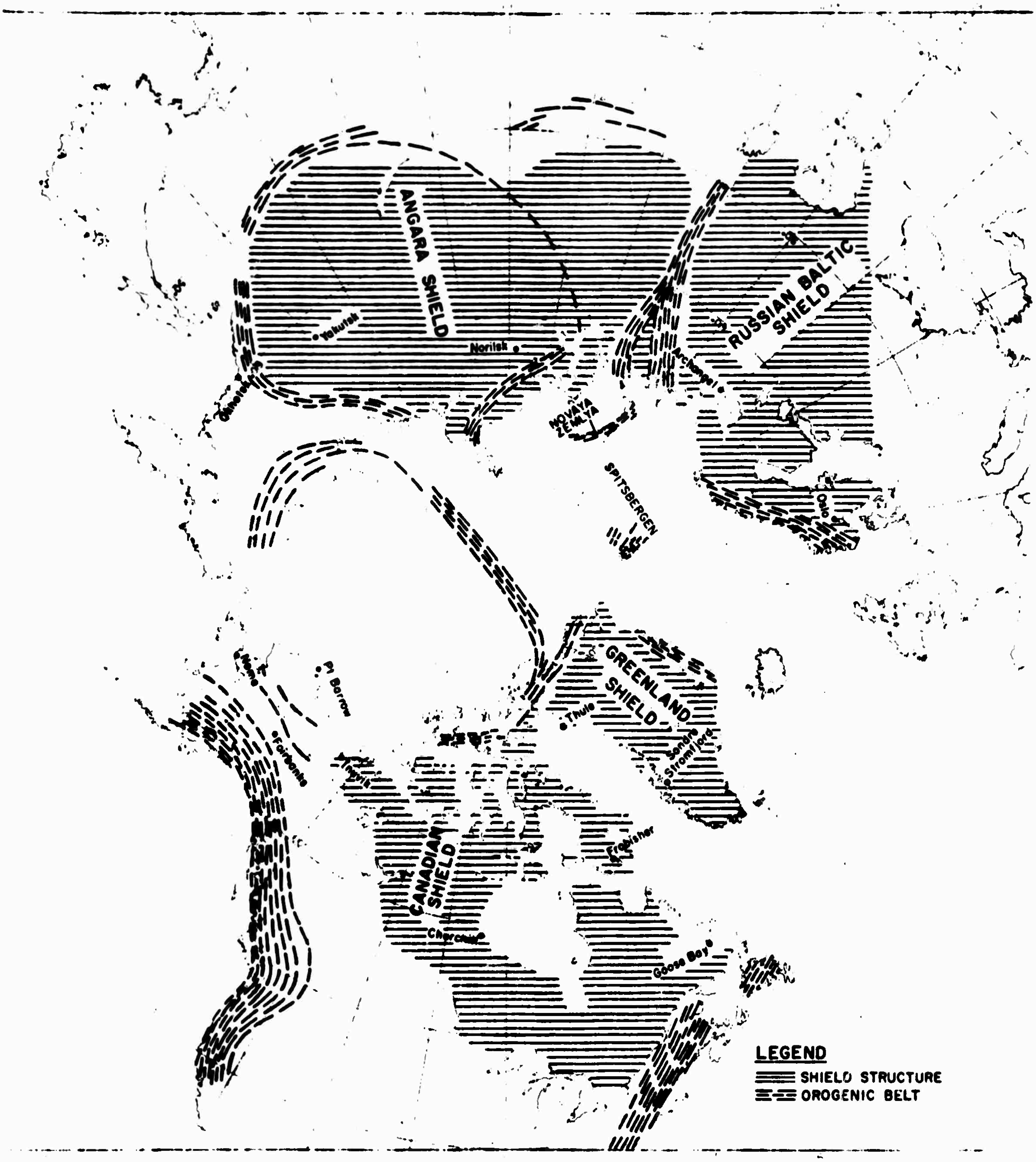


Figure VI-1 NORTH COLD REGIONS: MAJOR UNDERLYING STRUCTURES belt that includes Kamchatka is strongly seismic and includes numerous active and recently extinct volcanoes. The underlying geologic structure is little known where unconsolidated sediments of Quaternary age bury the older rocks in the north Alaska-Beaufort Sea-east Siberian coastal plains and continental shelves and in the West Siberian Lowlands.

The physical landscapes of the circumpolar regions contain major landforms, the origins of which may be traced to the distant geologic past, and minor landforms which have evolved in the last million years. The major elements, particularly in the shields, were already in existence before the onset of the Pleistocene glaciations. Practically universal in this category are the upland blocks which developed on ancient crystalline rocks. Today, with some later modifications, these uplands dominate the Arctic in castern Canada, East and West Greenland, Svalbard, Scandinavia, and northern Novaya Zemlya. Other widely distributed physiographic provinces are geologically younger. Rimming the north Pacific Ocean is a volcanic mountain belt that is part of the zone that encircles the Pacific Ocean; a second, less conspicuous volcanic sector is developed within the Mid-Atlantic Ridge, principally in Iceland. Regions of the late Tertiary and Quaternary deposition form another group. These include the coastal plains of northern Alaska and adjacent to the Beaufort Sea; the deltas of the Mackenzie, Yukon, and Siberian rivers; and the West Siberian Lowland.

Superimposed on the major elements are the modifications of the Quaternary. The most important of these resulted from the Pleistocene glaciations. Not all the circumpolar lands were glacierized as aridity was so intense in some areas that glacier ice failed to accumulate. Eventually all lands around the northern Atlantic and the Norwegian and Barents seas were at some period covered with ice sheets. The moisture of the north Pacific was less effective in penetrating into the western Arctic; and extensive unglacierized areas existed throughout the Pleistocene in northeastern Siberia, central and northern Alaska, and interior Yukon. The most conspicuous modification by the glacier ice was in uplands near the sea where outlet glaciers flowed down the river valleys to the ocean, resulting in the formation of the fiords that are such a striking feature of the upland coastal regions. Fiords developed at the same time elsewhere although their characteristics are usually rather different. This is particularly true of the south and southeastern coast of Alaska, Peary Land, and Axel Heiberg and western Ellesmere islands. In all cases the fiords are curved rather than angular and are commonly longer and wider than the classic fiord. Channelled ice also excavated deep valleys on the continental shelves including the Amundsen Gulf Sea Valley and the troughs between many of the Queen Elizabeth Islands, all now submerged. During the Pleistocene glaciation the highest uplands near the sea began to develop alpine scenery - vertical-sided valleys and knife-edge mountain ridges. Typically found on the west coast of Spitsbergen, narrow zones of similar scenery are found at many places in northern Scandinavia, Greenland, and the eastern Canadian Arctic. By contrast, in the interior of the northern continents glacierization led to a reduction in relief, either by ice scouring as found in many shield areas (e.g. interior Finland) or by redeposition of debris in glacial landforms.

The Quaternary glaciations and changes of sea level brought significant changes to the landscapes of the Arctic. In the period since, geomorphic processes unique to northern areas have produced further changes. These result partly from low temperatures and the freezing and survival of ice in the ground, while other special processes occur in the arid high arctic because of the absence of a continous vegetative cover.

A structural map of the circumpolar world suggests that there are close geologic links between the various continents. For physiographic purposes it is better to retain the basic continental divisions of North America, Eurasia, and the North Atlantic islands (Fig. 5-2). In the North American northlands six physiographic provinces are recognized. The largest is the Precambrian Canadian-Greenland Shield that occupies the Canadian arctic mainland, most of the islands south of Parry Channel, and a large part of Greenland. Across the vast area nearly 2,200 miles in diameter, crystalline granites and gneisses are the dominant rocks. Except where ice buries the land, the scenery is characteristically rocky and rolling with innumerable lakes and ungraded, fast-flowing streams. In North America, the Shield superficially resembles a saucer. It is highest along the eastern margin from Ellesmere Island to northern Labrador where glaciated peaks reach 6,000 to 7,000 ft above sea level and fiords penetrate the coast. West of the highlands there is a broad upland zone, commonly between 1,000 and 2,000 ft above sea level. Within the shield are basins, of considerable geologic age, that are floored by younger rocks, mainly limestones. They have been partly submerged to form Foxe Basin, Hudson Bay, and the channels of the southern part of the Canadian Arctic Archipelago. The western margin is generally less elevated although it produces a conspicuous scarp where it overlooks the Mackenzie Lowland. A belt of plains and low plateaus including the Mackenzie Lowland and the plains south of Parry Channel surrounds the Shield. The lowlands are developed on sedimentary rocks and are underlain by the Shield in this area. The Greenland sector of the Shield resembles an elongated saucer with a high mountainous rim. No young Paleozoic sediments are found within the Greenland Shield, but the island is apparently crossed by volcanic rocks that appear on the coasts in Scoresby Sund and Disko Bugt.

The Canadian-Greenland Shield, and associated platform areas, are partly

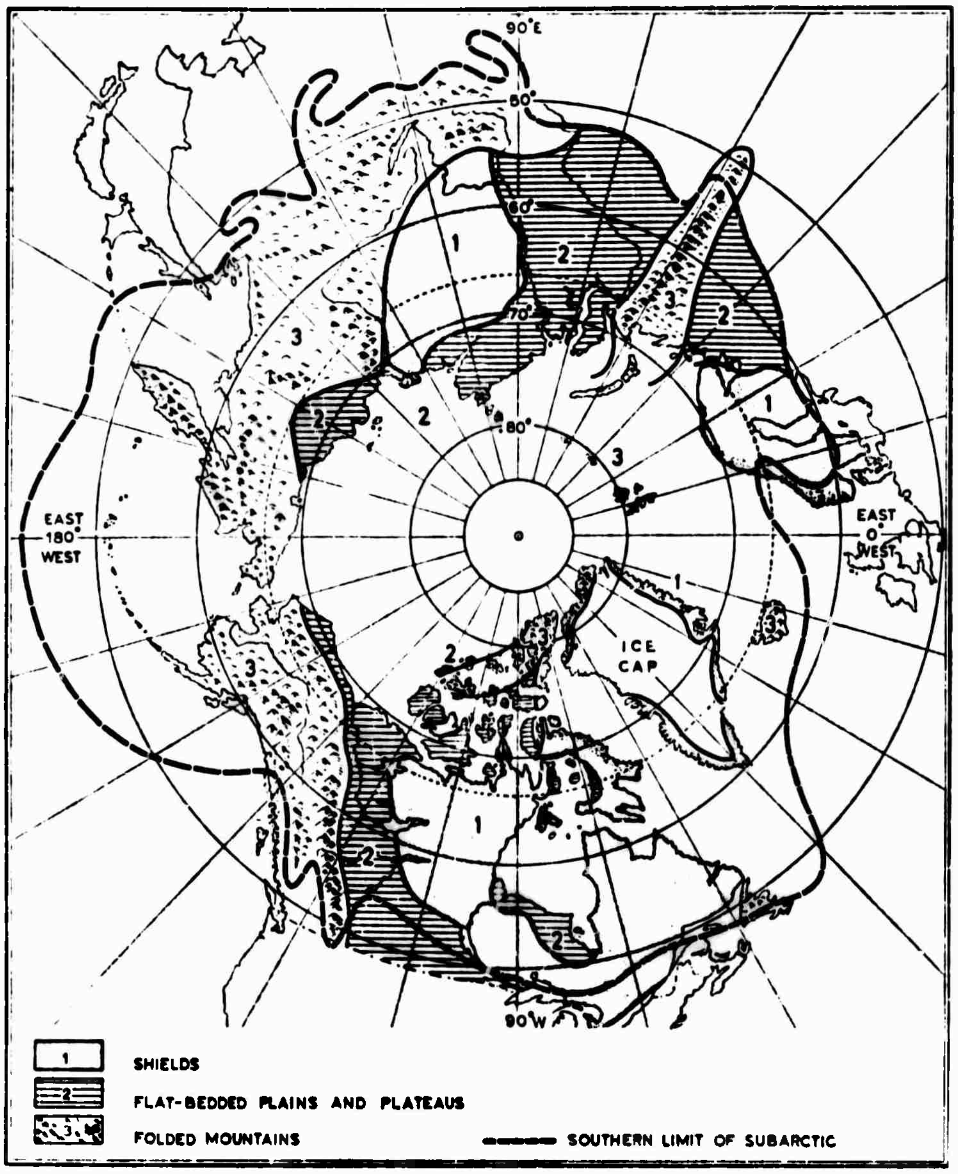


Figure VI-2 PHYSIOGRAPHIC REGIONS

surrounded by folded-rock zones. The Shield and this platform zone are regions of folded rocks of which three main groups may be recognized. The largest is the western Cordillera that includes the Cascades and Coast Ranges of western North America. The northern Cordillera in the Yukon Territory and Alaska may be divided into three provinces. The area adjacent to the Pacific Ocean is the Pacific mountain system which includes the Aleutian Ranges, the Alaskan Range, and the Chugach and other mountains; on the north and eastern side of the Cordillera are the Brooks Range of northern Alaska and the Richardson Range of the Yukon and Mackinzie district; and between the two mountain belts is a complex area of plains and highlands drained by the Yukon and Kuskokwim rivers. The second folded rock province is in the Queen Elizabeth Islands. It includes the parallel hill ridges on Melville, Bathurst, and Cornwallis islands and the alpine peaks of Ellesmere, Axel Heiberg, and Peary Land. The third folded province forms the mountains of northeast Greenland. Finally, along the edges of the Arctic Basin from Point Hope in northwest Alaska eastward to the Mackenzie and then along the outer northwestern islands of the Canadian Archipelago, is a coastal plain of sediments of late Tertiary and Quaternary age.

Several islands are found in the North Atlantic Ocean and the Norwegian Sea. All have structural links with the adjacent continents although physiographically they may be treated separately. The largest island, Iceland, is wholly of volcanic origin and lies at the intersection of the Greenland-northwest Scotland and the Mid-Atlantic ridges. Three hundred miles to the north-northeast is the small volcanic island of Jan Mayen. Farther north still is the archipelago of Svalbard. The terrain has evolved primarily on folded rocks although there is a great variety. The topography is typically mountainous in the west and generally plateau-like and often ice-covered in the center and east. Everywhere the uplands have been deeply dissected and the lower valleys drowned to produce fiords.

The Eurasian northlands contain several large upland regions separated by wide depressions. Two of the uplands, the Baltic Shield and the associated Russian platform in the west and the Siberian platform, are typical shield areas partly blanketed by younger sedimentary rocks. The Baltic Shield includes the Poluostrov Kol'skiy and the northern periphery of Scandinavia; it is an undulating upland with an average elevation of 1,500 ft that reproduces in many ways the landscape of the Canadian Shield. The surface of the Shield declines in the vicinity of the White Sea, and east of Arkhangel'sk the Precambrian rocks disappear beneath sediments. Along the west side of the Shield the Precambrian rocks are replaced by geologically younger, highly folded rocks; and there is a deeply dissected highland zone with numerous fiords and alpine peaks exceeding 6,000 ft. The central Siberian platform is conspicuously different. In the northwest is a high section developed primarily on flat rocks through which the rivers flow in deep valleys and where mountains such as the Putorana exceed 6,000 ft. The extreme northeast part is generally lower.

Between the two shield and platform divisions, a belt of folded rocks along the Ural-Novaya Zemlya axis produces a broad flat-topped upland with a general height of about 3,000 ft. The Urals show little sign of glaciation, but Novaya Zemlya is dissected by glaciers, glaciated valleys, and fiords. A second folded zone produces a vast highland massif of complex mountains and plateaus in northeast Siberia. Many of the ranges are arcuate and two of the largest, the Verkhoyanskiy and Cherskiy khreket, both have peaks over 10,000 ft.

The West Siberian Lowland province, over 500 miles wide, extends southward from the Kara Sea between the Urals and the central Siberian platform. The lowlands are developed on horizontal sedimentary rocks covered with deep glacial, alluvial, and marine deposits. They form a remarkably flat, occasionally terraced, poorly-drained plain. The smaller, but in other ways similar, North Siberian Lowland province separates the hill ridges of Poluostrov Taimyr from the Central Siberian Uplands to the south. The eastern limit of these lowlands is the mouth of Khatanga River, but similar plains occur discontinously far east along the coast and include the New Siberian Islands.

B.6.1.2 Soils, Permafrost and Vegetation

(See pages 265-A through 265-K)

B.6.1.3 Shorelines

Arctic littoral processes in ice-free summer months do not differ greatly from those in temperate latitudes. The differences are more of degree than of kind, particularly as the annual period of wave action and the amplitude of the waves is reduced in the presence of ice. However, several distinctive processes are operative at break-up and freeze-up of sea ice.

Sea ice has, in general, a secondary role to waves as an agent in erosion, transportation ,and deposition. During the fall, when air temperatures drop below freezing, an ice shelf commonly forms along arctic beaches where there is a low tidal range as water from storm waves and spray freeze on the beach to leave layers of ice-cemented gravel and ice. Sand and gravel are often washed onto the developing ice shelf and become incorporated. This ice shelf protects the beach against subsequent wave action and ice-shove. If pack ice moves onshore early in the winter, shorefast ice with adhering debris may be shoved inland onto the beach, thus adding material to it. During the winter, shorefast ice and offshore ice stop all wave action. Pack ice at break-up, and also later in the summer, may push onto the beach or impinge against the sea cliff, if present. Ice moving onto a beach tends to act either as a plough or a raft. Where ploughing occurs, the ice planes and pushes debris into mounds and ridges, usually below the upper limit of storm waves. Where rafting occurs, debris frozen and

Soils, Permafrost and Vegetation

Most of the soils of the Arctic are quite young and poorly developed. In large part this is due to the presence of continental glaciers which covered the area and have retreated only in the recent, geologic, past. Other reasons are frost action, the retarded biochemical activity resulting from the low thermal regime, and a lack of adequate moisture due to the slight precipitation. Drainage is frequently very poor as a result of the presence of permafrost, hence many arctic soils are moist.

As a result of these conditions, the soils are usually shallow and have not developed genetic horizons. Most commonly they are alkaline or netural but occasionally they may be acidic. In the islands north of the mainland, polar desert soils are often found. They may show genetic horizons, not only on more level surfaces but on some of the steeper slopes. The surface is frequently as much as 50% coarse fragments, pebbles and rocks and can exhibit the characteristics of aridity: salt crusts, alkali flats, and carbonate accumulations. In profile they pass from the desert pavement of the surface to dark brown, sandy, gravelly loam to red-, gray-, and yellow-brown sands. Vegetation is always sparce and may cover less than 1% of the surface.

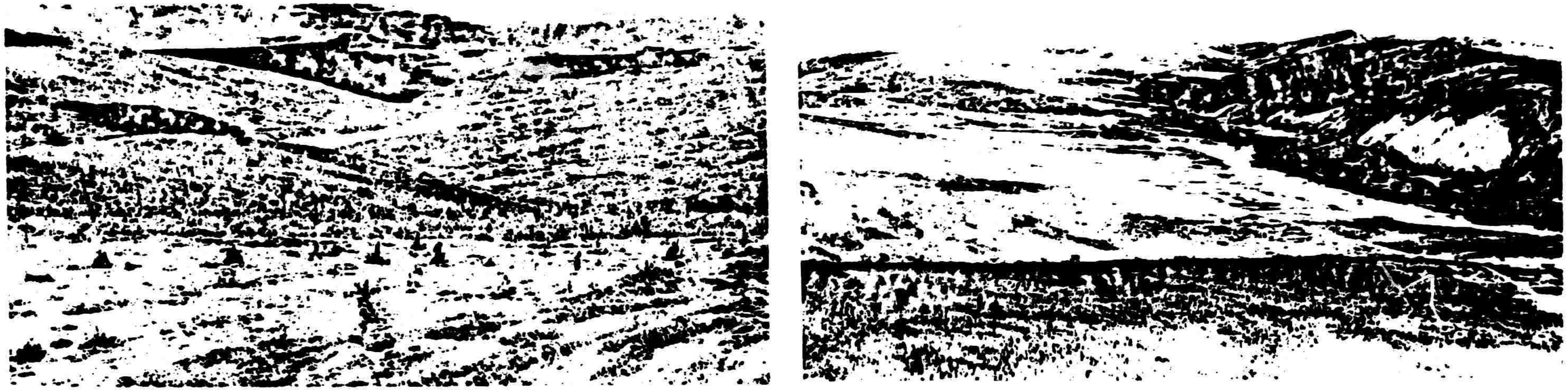
Elsewhere, the soils range from bog to arctic brown to glei to subarctic brown forest and wooded to podzol, depending upon parent material, slope, exposure, vegetative cover and the length of time since the area was last covered by ice. The occurrence of the various types has been mapped on a small scale for the entire region but detailed maps exist for only a few areas which have been of concern for any of numerous reasons.

The vegetative cover of most of the arctic is best described by the term tundra. Originally the Lapp word for the treeless areas of northern Scandinavia, it has come to mean the entire plant community of the treeless areas. It includes lichens, mosses, grasses, sedges and woody shrubs. Some tree species, such as willow, birch and alder may occur in sheltered arctic valleys and may attain a height of six to eight feet. The arctic deserts are the least vegetated regions with their lichen-covered rocks and scattered clumps of hardy perennials. The next most developed community is the lichen-moss tundra which occurs on the coarse materials of tills, terraces and strands. Further south this dry tundra contains increasing amounts of heath and berry-producing plants. When precipitation and the nutrient supply are adequate, these communities will completely cover the surface.

The wet tundras contrast markedly with the dry as they occur on finer-grained soils with poor drainage. The plant cover is a dense mat of grasses, sedges and moss. As a result of the drainage, bogs are frequent in the low areas. Frost action leads to a hummocky surface in which the mounds are moss covered and the swales support the moisture-loving species. Finally, as the treeline is approached, the tundras begin to support increasing numbers of shrubs.

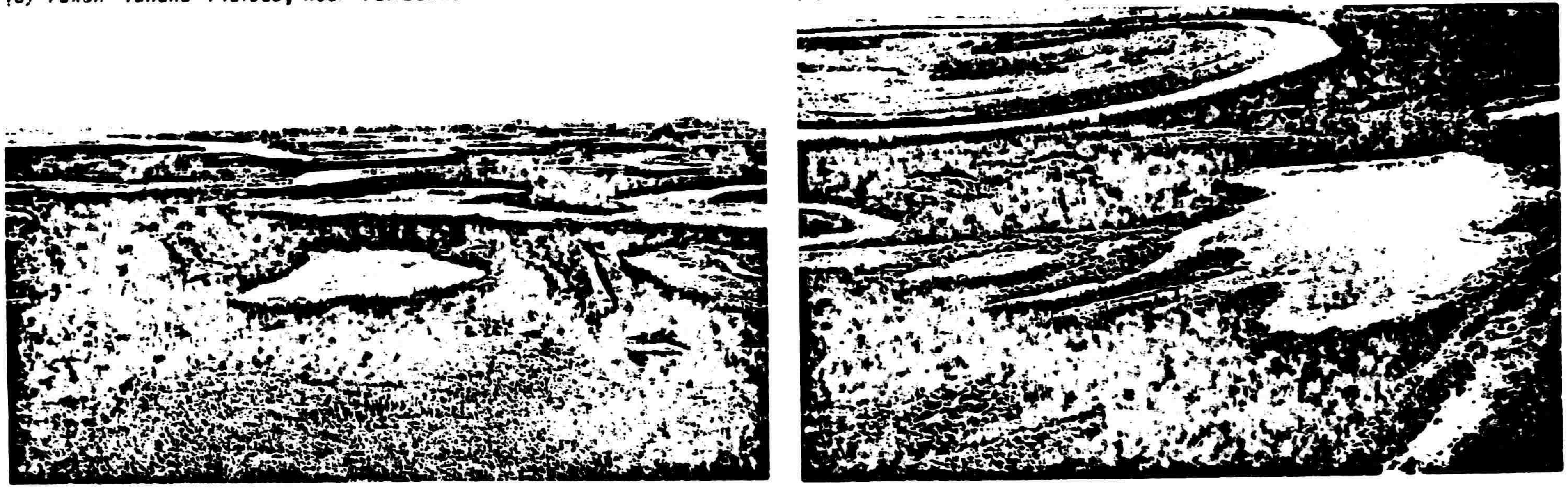
Despite the problems in defining it and its state of geographic flux, treeline is the most obvious of the arctic boundaries. It is taken to be the poleward or altitudinal limit of standing trees and in consequence there is a gradual transition from tundra, without trees, to small stands of spruce and larch to larger stands of greater numbers of species and finally the thick boreal forests of the subarctic VI-3).

One of the most unique phenomena of the Arctic is the presence of permafrost. This perennially frozen state of the subsurface soil and rock leads to many of the surface conditions which distinguish the region, strongly influences the surface condition and vegetative cover and creates many of the engineering problems which occur there. When permafrost occurs in bedrock or areas of coarse, well drained materials it does not significantly alter conditions from those found elsewhere on the globe. However, in areas with poor drainage and fine-grained soil materials it can be the controlling element. Throughout the area in which perma-



(a) Yukon-Tanana Plateau, near Fairbanks

(b) Yukon-Tanana Plateau, near Circle



(c) Yukon Flats, near Beaver Creek

(d) Yukon Flats, lower Chandalar River

Figure VI-3 TOPOGRAPHIC TYPES IN THE CENTRAL PLATEAUS frost occurs the surface freezes and thaws annually with the passing of the seasons. The depth of the active layer varies with latitude and soil type from as little as a foot to six feet or more. Since the permafrost is impervious to surface runoff and roots, it may severely restrict plant growth and prevent water from draining as it would in other situations. The frost-induced swelling of the active layer causes a variety of irregular surface features, such as those illustrated in Fig.VI-4, and further disrupts the root systems of the plant cover. Most prominent of the surface manifestations of permafrost is the patterned ground (Figs.VI-5 and 6) associated with the presence of subsurface ice wedges (Fig.VI-7c). On slopes, permafrost can cause massive slumping of the surface materials as they glide over the water-lubricated permafrost table or it can cause finer, less disruptive, stripes and fingers of sloughed material to work their way downslope. Such conditions mix the soil layers, may expose buried layers of organic materials, and result in differing vegetative covers within very short distances.

Since it is a phenomena of temperature any change in the thermal regime of an area can lead to thawing of the permafrost. Naturally this results in thermokarst topography and slumped banks and cliffs (Fig.VI-7); and when induced by human activity, it can significantly accelerate erosive processes (Fig. VI-8) or wreck havoc on structures which were not properly designed. The mechanical problem associated with the degradation of permafrost is that, upon thawing, the portion of it which was ice melts and the water runs off if a slope is present or the material loses its capacity to support an overburden. The amount of ice present in permafrost may vary from an insignificant number of minute crystals in porous materials to as much as 90% in some sedimentary deposits and has necessitated the extended classification of soils, as shown in Table 6-I.

The ability to travel on surfaces underlain by permafrost is clearly related to temperature. During winter, when the active layer may be frozen through and

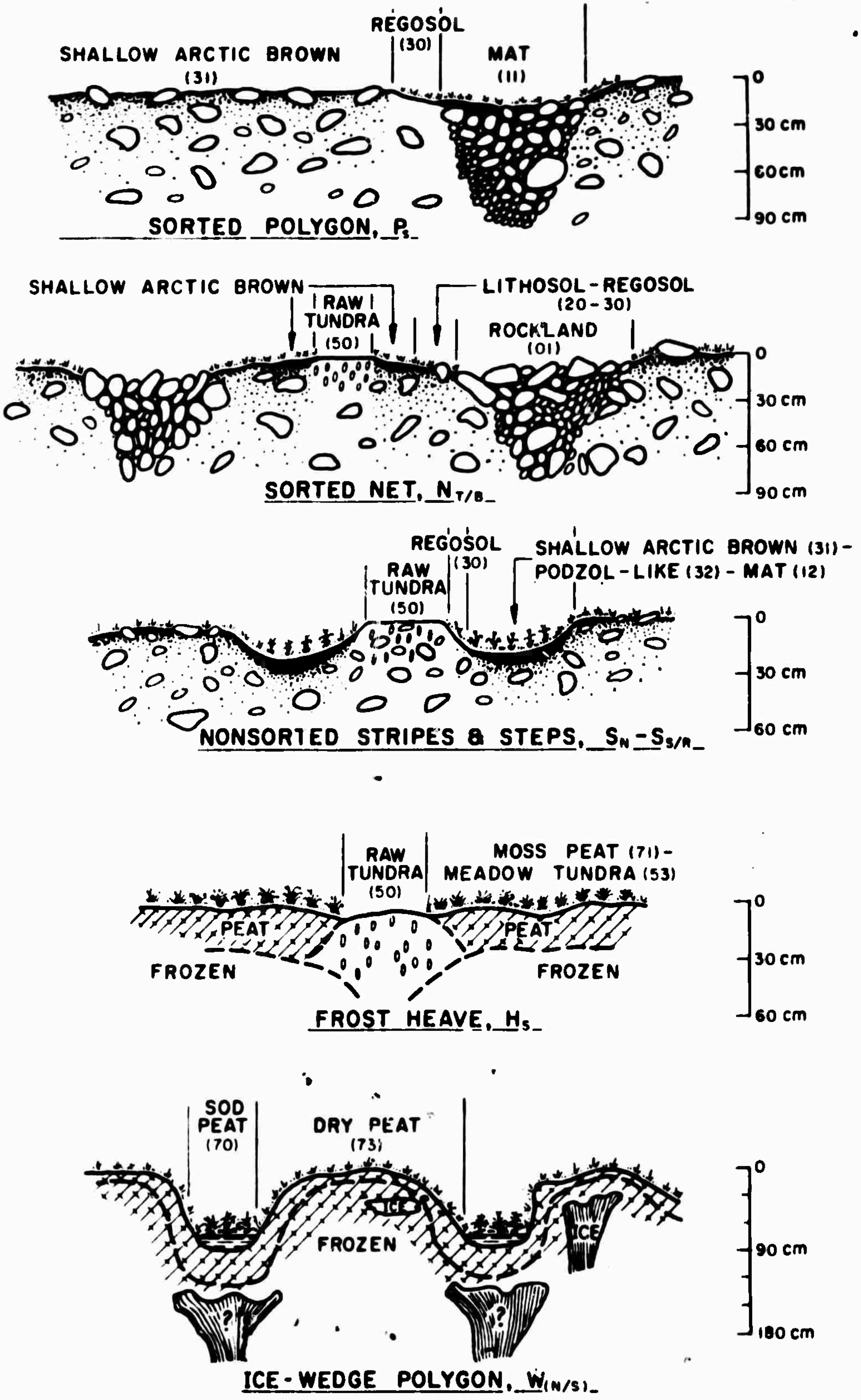
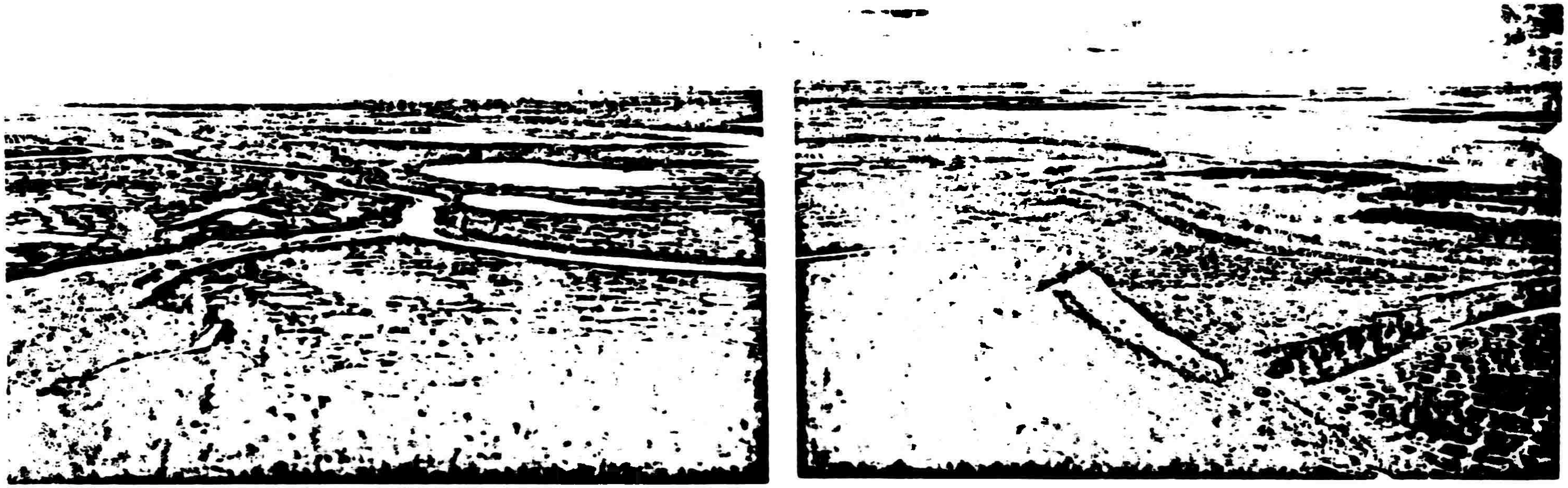


Figure VI-4 VARIATIONS IN SURFACE FEATURES AND SOIL MORPHOLOGY AS A RESULT OF FROST ACTION



(a) Outer Coastal Plain, Ikpikpuk River

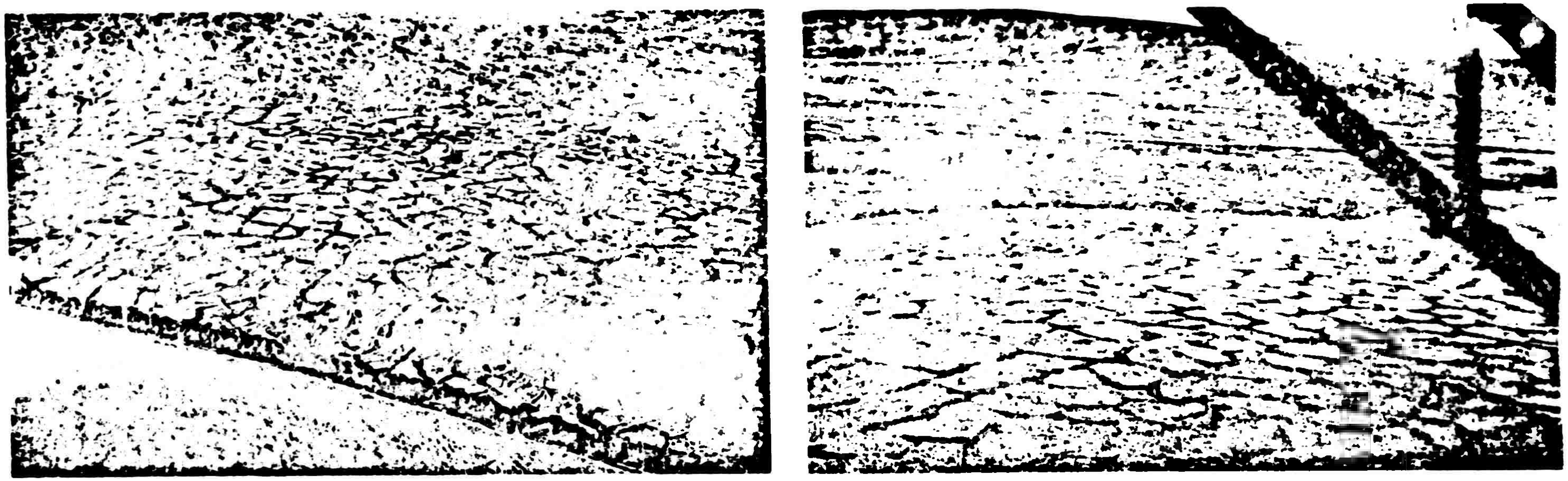
(b) Outer Coastal Plain, Ikpikpuk Delta area

NOT REPRODUCIBLE

(c) Inner and outer coastal plain transition, Ikpikpuk River

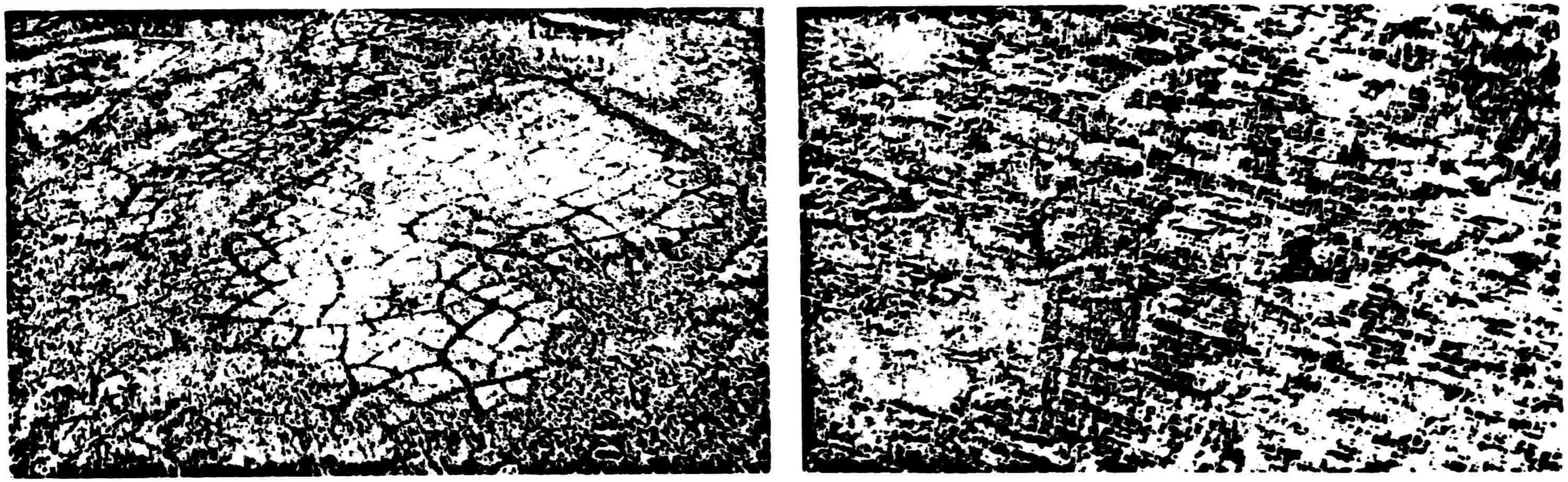
(d) Inner Coastal Plain near the Topogorak River

Figure VI-5 TOPOGRAPHIC FEATURES OF THE ARCTIC COASTAL PLAIN



(a) Arctic Coastal Plain

(b) Seward Peninsula



(c) Yukon Flats

(d) Tanana Valley

Figure VI-6 LOW ALTITUDE OBLIQUE PHOTOS OF RAISED - CENTER TYPE POLYGONS IN ALASKA

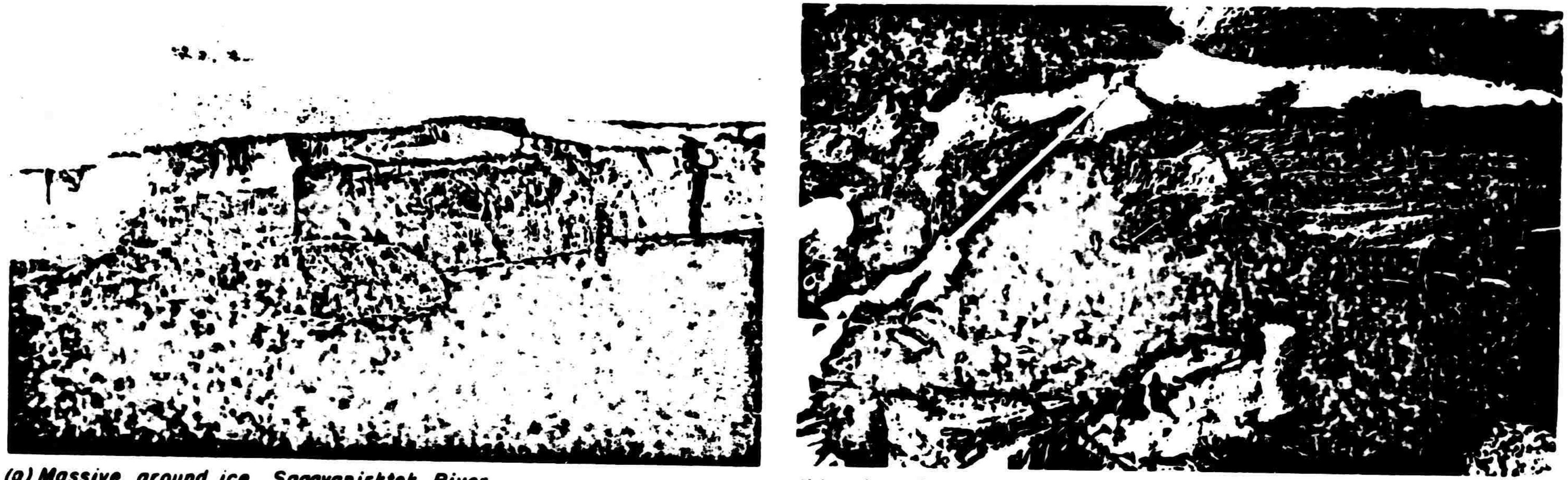
NOT REPRODUCIBLE

(a) Unchecked highway culvert effluent

(c) Drainage ditch thaw near Umiat

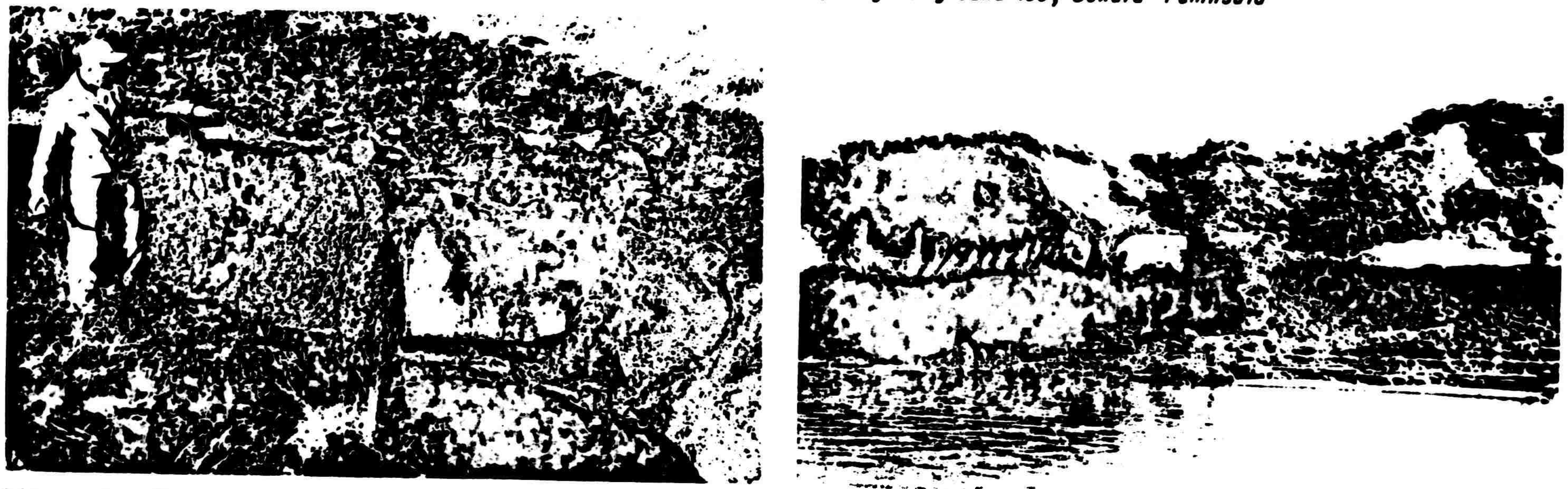
(b) Seven foot cavity in the "cat tracks" in 24 hours of thaw, Barter Island

Figure VI-7 ACCELERATED EROSION RESULTING FROM RUNNING WATER PASSING OVER FROZEN SOILS



(a) Massive ground ice, Sagavanirktok River

(b) Ledge of ground ice, Seward Peninsula



(c) Ice wedge, Nome

(d) Massive ground ice, Kobuk River

Figure VI-8 TYPES OF ICE AND ICE-SOIL FORMS

Table 6-I CLASSIFICATION SYSTEM OF FROZEN SOILS

PART I DESCRIPTION OF SOIL PHASE (a) (Independent of Frozen State)	Classify Soil Phase by the Unified Soil Classification System							
(1)	Major Group		Sub-Group		Field Identification (6)	Pertinent Properties of Frozen Materials Which may be Measured by Physical Tests to Supplement Field Identification (7)	Guide for Construction on Soils Subject to Freezing and Thawing	
	Description (2)	Designation (3)	Description (4)	Designation (5)			Thaw Characteristics (8)	Criteria (9)
PART II DESCRIPTION OF FROZEN SOIL	Segregated ice is not visible by eye (b)	N	Poorly bonded or friable	Nf	Identify by visual examination. To determine presence of excess ice, use procedure under note (c) below and hand magnifying lens as necessary. For soils not fully saturated, estimate degree of ice saturation: Medium, Low. Note presence of crystals, or of ice coatings around larger particles.	In-Place Temperature Density and Void Ratio a. In Frozen State b. After Thawing in Place Water Content (total H_2O, including ice) a. Average b. Distribution Strength a. Compressive b. Tensile c. Shear d. Adfreeze Elastic Properties Plastic Properties Thermal Properties Ice Crystal Structure (using optical instruments) a. Orientation of Axes b. Crystal Size c. Crystal Shape d. Pattern of Arrangement	Usually thaw-stable	The potential intensity of ice segregation in a soil is dependent to a large degree on its void sizes and for pavement design purposes may be expressed as an empirical function of grain size as follows: Most inorganic soils containing 3 percent or more of grains finer than 0.02 mm in diameter by weight are frost-susceptible for pavement design purposes. Gravels, well-graded sands and silty sands, especially those approaching the theoretical maximum density curve, which contain 1-1/2 to 3 percent finer by weight than 0.02 mm size should be considered as possibly frost-susceptible and should be subjected to a standard laboratory frost susceptibility test to evaluate actual behavior during freezing. Uniform sandy soils may have as high as 10 percent of grains finer than 0.02 mm by weight without being frost-susceptible. However, their tendency to occur interbedded with other soils usually makes it impractical to consider them separately. Soils classed as frost-susceptible under the above pavement design criteria are likely to develop significant ice segregation and frost heave if frozen at normal rates with free water readily available. Soils so frozen will fall into the thaw-unstable category. However, they may also be classed as thaw-stable if frozen with insufficient water to permit ice segregation. Soils classed as non-frost-susceptible under the above criteria usually occur without significant ice segregation and are usually thaw-stable for pavement applications. However, the criteria are not exact and may be inadequate for some structure applications; exceptions may also result from minor soil variations. In permafrost areas, ice wedges, pockets, veins, or other ice bodies may be found whose mode of origin is different from that described above. Such ice may be the result of long-time surface expansion and contraction phenomena or may be glacial or other ice which has been buried under a protective earth cover.
			Well bonded: No excess ice	Nb: n				
			Well bonded: Excess ice	Nb: e				
	Segregated ice is visible by eye. (Ice 1 inch or less in thickness) (b)	V	Individual ice crystals or inclusions	Vx	For ice phase, record the following as applicable: Location Orientation Thickness Length Spacing Hardness) per Part III below Structure) per Part III below Color) per Part III below Size Shape Pattern of arrangement Estimate volume of visible segregated ice present as percent of total sample volume.		Usually thaw-unstable	
			Ice coatings on particles	Vc				
			Random or irregularly oriented ice formations	Vr				
			Stratified or distinctly oriented ice formations	Vs				
PART III DESCRIPTION OF SUBSTANTIAL ICE STRATA	Ice (Greater than 1 inch in thickness)	ICE	Ice with soil inclusions	Ice + soil type	Designate material as ICE (d) and use descriptive terms as follows, usually one item from each group, as applicable: Hardness: HARD, SOFT (of mass, not individual crystals) Structure: CLEAR, CLOUDY, POROUS, CANDLED, GRANULAR, STRATIFIED Color (Examples): COLORLESS, GRAY, BLUE Admixtures (Example): CONTAINS FEW THIN SILT INCLUSIONS	Same as Part II above, as applicable, with special emphasis on Ice Crystal Structure.		
			Ice without soil inclusions	Ice				

DEFINITIONS:

Ice Coatings on Particles are discernible layers of ice found on or below the larger soil particles in a frozen soil mass. They are sometimes associated with hoarfrost crystals, which have grown into voids produced by the freezing action.

Ice Crystal is a very small individual ice particle visible in the face of a soil mass. Crystals may be present alone or in a combination with other ice formations.

Clear Ice is transparent and contains only a moderate number of air bubbles. (e)

Cloudy Ice is relatively opaque due to entrained air bubbles or other reasons, but which is essentially sound and non-pervious. (e)

Porous Ice contains numerous voids, usually interconnected and usually resulting from melting at air bubbles or along crystal interfaces from presence of salt or other materials in the water, or from the freezing of saturated snow. Though porous, the mass retains its structural unity.

Candled Ice is ice which has rotted or otherwise formed into long columnar crystals, very loosely bonded together.

Granular Ice is composed of coarse, more or less equidimensional, ice crystals weakly bonded together.

Ice Lenses are lenticular ice formations in soil occurring essentially parallel to each other, generally normal to the direction of heat loss and commonly in repeated layers.

Ice Segregation is the growth of ice as distinct lenses, layers, veins, and masses in soils, commonly but not always oriented normal to direction of heat loss.

Well-bonded signifies that the soil particles are strongly held together by the ice and that the frozen soil possesses relatively high resistance to chipping or breaking.

Poorly-bonded signifies that the soil particles are weakly held together by the ice and that the frozen soil consequently has poor resistance to chipping or breaking.

Friable denotes extremely weak bond between soil particles. Material is easily broken up.

Thaw-Stable frozen soils do not, on thawing, show loss of strength below normal, long-time thawed values nor produce detrimental settlement.

Thaw-Unstable frozen soils show, on thawing, significant loss of strength below normal, long-time thawed values and/or significant settlement, as a direct result of the melting of the excess ice in the soil.

NOTES:

(a) When rock is encountered, standard rock classification terminology should be used.

(b) Frozen soils in the N group may, on close examination, indicate presence of ice within the voids of the material by crystalline reflections or by a sheen on fractured or trimmed surfaces. However, the impression to the unaided eye is that none of the frozen water occupies space in excess of the original voids in the soil. The opposite is true of frozen soils in the V group.

(c) When visual methods may be inadequate, a simple field test to aid evaluation of volume of excess ice can be made by placing some frozen soil in a small jar, allowing it to melt and observing the quantity of supernatant water as a percent of total volume.

(d) Where special forms of ice, such as hoarfrost, can be distinguished, more explicit description should be given.

(e) Observer should be careful to avoid being misled by surface scratches or frost coating on the ice.

NOTES:

The letter symbols shown are to be affixed to the Unified Soil Classification letter designations, or may be used in conjunction with graphic symbols, in exploration logs or geological profiles. Example - a lean clay with essentially horizontal ice lenses.

The descriptive name of the frozen soil type and a complete description of the frozen material are the fundamental elements of this classification scheme. Additional descriptive data should be added where necessary. The letter symbols are secondary and are intended only for convenience in preparing graphical presentations. Since it is frequently impractical to describe ice formations in frozen soils by means of words alone, sketches and photographs should be used where appropriate, to supplement descriptions.

The abbreviation nfs is commonly used to designate non-frost-susceptible materials on exploration logs and drawings.

Arctic Construction and Frost Effect Laboratory, U.S. Army Engineer Division, New England, Waltham, Mass. January 1961 bearing on as much as a thousand or more feet of permafrost, the surface will sustain extremely heavy loads and movement is governed more by the surface roughness and slope then by bearing strength. In summer, the load-bearing capacity of arctic terrains varies from the pavements of polar deserts, which can sustain landings by large cargo aircraft, to the boggy mucks of the muskeg which may fail at pressures as low as 1 or 2 psi. Studies of the nature and extent of arctic lands and soils are and have been conducted by many organizations, of which the Cold Regions Research and Engineering Laboratory is prominent in this country, while the mapping of it lies within the purview of the branches of the Geological Survey.

Construction on and travel over permafrost can be accomplished but requires that engineers and planners know and respect it and be prepared to face larger costs than would otherwise be necessitated. One of two courses must be followed: either the thermal regime must be left undisturbed or all permafrost-susceptible materials must be removed from the area of concern. Where a single structure or reasonably permanent facility such as an airfield is involved, it may be practical to consider evacuating the frost-susceptible materials and replacing them with coarser material which is not subject to frost action. The only other recourse is to leave the permafrost intact and cover it with a sufficient amount of gravelly fill to completely insulate the surface from frost action (Fig.VI-11b). It is usually advisable to follow this course even if a building is to be set on piles embedded into the permafrost. Such a plan may well necessitate vast amounts of borrow and may require hauling it long distances. The alternatives are generally unwarranted or unacceptable.

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incorporated in the ice may be transported onto the beach, locally above the reach of storm waves. Usually storm waves soon rework most of the transported material and by freeze-up little evidence of it is visible; however, in high arctic seas that are permanently choked with ice, ice-push beach ramparts are a normal feature at the rear of beaches.

On coasts where the tidal range is considerable, boulder barricades are the most conspicuous sign of the action of sea ice. Typically there is a narrow string of boulders parallel to the shore and several hundred feet offshore. At low tide they are clearly visible above the water surface; at high tide, when they are submerged, they represent a navigational danger on the approach to many open beaches. Generally the depth of water increases rapidly beyond the boulder limit. The pushing and rafting of boulders by the drifting sea ice that forms the boulder barricades is also responsible for the development of boulder-covered flats which are exposed at low tide near the heads of arctic bays. Beach morphology is affected by grounded sea ice and permafrost ice. Oval ponds are produced by the melting of lenses of ground ice that have developed in the back-shore zone of beaches. Hollows and soft areas on sand beaches are more likely to result from melting of grounded ice floes.

There are strong regional differences between Pacific or western arctic shores and those in the Atlantic, or eastern, Arctic. The former may be defined as extending from Cape Chelyuskin in Siberia to the Beaufort Sea and Amundsen Gulf. This sector is characterized by long stretches of unconsolidated frozen sediments, shallow offshore depths, and small tidal range. The Atlantic arctic coasts include those of the eastern Canadian Arctic, Greenland, the Atlantic coasts proper, the Barents Sea, and to a lesser degree the Kara Sea. Here the coasts are commonly bedrock. Coarse-grained glacial sediments are widespread, and the tidal range is moderate to high.

Offshore bars, baymouth bars, spits, tombolos, and other bar deposits are found along much of the Pacific arctic coast. This is to be expected where offshore depths are modest and the supply of debris from coastal recession and rivers is large. Offshore bars parallel hundreds of miles of coast, separated from it by a lagoon perhaps several miles wide. Spits build out where there are directional changes in the coast in response to winds and currents. Driftwood and large logs become stranded on the bars and spits near a source of driftwood, such as east and west of the Mackenzie Delta. The higher parts of bars and spits rise well above sea level where they have been built by storm waves, ice shove, and wind action.

Large deltas are common in the Pacific and include the Lena, Indigirka, Kolyma, Colville, Mackenzie, and others. They terminate either in the open sea or, more often, in estuaries and bays fringed with bars. Deltas which build out into the sea have shoal water offshore from the distal islands. The rivers contribute sediment, principally fine-grained, in the early summer at break-up and after severe storms.

The main characteristics of the Atlantic shores are the spectacular cliffs which can be found in many sectors. In the Precambrian Shield areas the majority of cliffs are preglacial or, in the case of fiords, of glacial origin. These cliffs show little evidence of strong contemporary erosion subaerially or by waves. In contrast, many of the cliffs in younger sedimentary and igneous rocks, particularly in Svalbard and Iceland, have developed massive postglacial scree slopes that are active today. In some parts of the Atlantic Arctic, notably Scandinavia, a rock platform that varies in width up to several miles separates the cliffs from the sea. This feature, known as the strandflat, is locally drowned forming a skerry coast.

The main origin of beach material on the Atlantic coasts is glacial drift.

In the western Arctic, sea cliffs which may rise to 100 ft or more front along hundreds of miles of coast. The cliffs are usually eroded into unconsolidated frozen sediments with an ice content varying inversely with the grain seze. In sands and gravels, only a bonding cement occurs; but in cohesive, fine-grained silty clays, clays, and peats, the ice content (weight of ice to dry soil) may reach hundreds of percent. The frozen sea cliffs retreat by undercutting, thermal erosion, and slumping. The greatest erosion occurs during storm surges when the water level may rise 5 to 10 ft. When there is a combination of a high water level and strong waves, the slumped debris at the foot of the sea cliff may be removed and a notch 5 to 15 ft deep eroded into the cliff to produce an overhang in the frozen sediments; this is followed by sediment failure along vertical fault planes and retreat of the cliff face. Thermal erosion and slumping along steep cliffs poses a hazard below, because of a constant downward rain of debris. Thermal erosion may also produce glacier-like mud streams. Differential thermal erosion is particularly marked along coasts where the large ice-wedges of tundra polygons are present. Gullies selectively etch out the ice-wedges to produce indentations or hanging valleys along the cliffs. In winter, the sea cliff and beach zone is an accumulation site for large snow drifts which persist and protect the cliff foot in early summer. Rapid slumping from above may bury the snowbanks completely and some survive over the summer, thus adding additional protection to the cliff foot for a second summer season. The type of sediment in the sea cliff tends to be reflected in the number and size of adjacent beaches and bars. As a generalization, beach and bar formation diminishes directly with reduction of grain size - ice muds may have no associated beach and bar deposits.

In the offshore zone, where ice pressure ridges may ground in depths reaching down to 100 ft, irregular depth fluctuations of 5 or more ft in horizontal distances of 25 to 50 ft have been found superimposed upon the gently sloping bottom. These irregularities have been attributed to the grounding of pressure-ridge ice.

The drift contained particles ranging in size from microscopic grains to boulders. Some were deposited as a heterogeneous mass (till); large quantities were laid down along valley walls by turbulent meltwater streams beside disintegrating ice masses as bottom or shore deposits of short-lived ice-blocked lakes or as outwash plains in front of glacial termini. Undoubtedly, many of these features still lie beneath the sea.

Blue clays of marine origin are an important source of beach materials. These sandy, fine-grained sediments are frequently exposed along the coast by stream or wave erosion. Their color is caused by the reduced state of iron in secondary minerals. They were deposited in low-energy sedimentary environments toward the end of the glaciation when a rising sea level drowned the coastal area appearing from beneath the ice. Subsequently many of the submerged coasts were uplifted, so that these, and older deposits, have emerged from the sea. In the process, coarse sands and gravels have been spread on top of them by waves and currents.

The beaches can be a very complex mixture of fluvial, glacial, wind, and marine sediments. The backshore is often sand and gravel, sometimes with boulders breaking the surface and often with poorly developed dunes. The foreshore and offshore zones have their fines washed away; and if the source of beach material contains coarse particles like gravel and boulders, these become concentrated in the zone of most intense wave action. This is strikingly apparent where boulder beaches are developing in glacial till. Elsewhere shingle beaches and bars are widely distributed. Except in shallow bays and troughs, the bottom tends to drop away quickly. During storms, eroded beach sediment is easily carried into deep water, from which it cannot be returned during calm periods. This condition, as well as a frequent paucity of sediment, contributes to the formation of relatively narrow beaches. If it were not for the cover of ice which protects them from destruction by winter storms, the attrition of beaches would be even greater.

Glacial and marine deposits which emerged from the sea during uplift of the land have been reworked by waves, and a record of this has been left along many coasts in the form of terrraces and drift-free bedrock up to levels of 100 to 750 ft.

More generalized descriptions of the shoreline characteristics are given below in the section dealing with Regional Characteristics. Detailed information concerning most of the arctic shoreline is lacking or of uncertain accuracy. The geological surveys of the respective circumpolar nations map their arctic coasts with the greatest accuracy possible and the hydrographic services chart the near-shore waters and coasts and issue Pilots and sailing directions. Information of this nature is, in general, available from the governments of the various nations.

B.6.2 Regional Characteristics

B.6.2.1 North Atlantic

There is considerable uniformity in the topography around the northeast Atlantic. Characteristically, the coasts are developed on crystalline or intensely folded rocks. Flat-topped uplands dominate the landscape, with altitudes varying in different areas from 2,500 to 5,000 ft. The uplands are often intensely dissected into isolated pyramidal peaks, notably in Spitsbergen. Straight-sided, linear valleys of great length are common; the coastal sections are drowned forming fiord coasts. Surficial sediments are typically coarse; clastic materials (sands, gravels, boulders) and fine sediments are relatively rare. This is significant because, although permafrost is found in the colder coastal area (Svalbard, East Greenland), engineering problems due to its presence are minimal. The main exception to these generalizations of the topography is along the active volcanic zone that coincides with the Mid-Atlantic Ridge and on which Iceland and Jan Mayen are located.

The environment of Greenland is everywhere dominated by the presence of the inland ice sheet which, together with secondary glaciers, covers five-sixths of the island. Ice-free areas are restricted to coasts; they have a maximum width of about 120 miles, and even in these areas there are numerous small ice caps and glaciers. On the east coast of Greenland the margin of the inland ice stops short of the sea except south of Angmagssalik and in the large outlet glaciers of the northeast at Nioghalvfjerdsfjorden, Jøkelbugten, and Dove Bugt. At other points many smaller valley glaciers calve icebergs into the Greenland Sea.

Greenland has the shape of an elongated basin with its major axis lying north-south. In the interior of the island the earth's crust is depressed by the ice sheet and in the north it lies 800 ft below sea level. The rim of the basin is penetrated by deep valleys through which local and inland ice glaciers move toward the sea. The rim is highest in the east where it reaches 12,000 ft in the Gunnbjørns Mountains. Two major ice-free areas are present in North and East Greenland - on the east coast between Scoresby Sund and Germania Land and in Peary Land in the North. The widest belt of ice-free land is around Scoresby Sund; but because of the deep indentation of the coastline, the inland ice is never more than 44 miles from the sea.

Peary Land and the east coast north of 68°N has evolved on rocks that are geologically younger than the Shield; some are folded and others are horizontal sandstones and basaltic lavas. The folded rocks have produced a mountain topography not unlike the Shield mountains except that the lowlands are commonly broader, particularly in Peary Land. The horizontal rocks, especially around Scoresby Sund, form plateau landscapes although the table mountains are often so dissected that only peaks and ridges survive.

A third topographic element characteristic of East Greenland, in addition to the mountains and ice, are the fiords which are among the largest and the deepest in the world. The fiords in Peary Land resembled those of the northwestern Canadian arctic islands with wide swinging curves and great width that in many ways are more typical of drowned river valleys than glaciated fiords. Elsewhere in the area they are usually straight with sharp, angular breaks. All the east and north coasts are underlain by permafrost. Unstable soil conditions and patterned ground are found throughout the whole of the ice-free east coast although they may be somewhat less common in the far north due to great soil aridity.

The west coast of Greenland from Robeson Channel to Kap Farvel is a deeply indented, fiord coast over 1,000 miles long. There are three distinct topographic sectors. In the north, beyond Kap York is a small ice-free area of sedimentary rocks that have developed table-top plateau typical of Peary Land and northeastern Greenland. For 390 miles between Kap York and Upernavik, the inland ice approaches close to Melville Bay except for a narrow, intermittent coastal strip of land and for the offshore islands.

The third sector includes the remainder of the coast. It contains the greatest amount of ice-free land and virtually the whole population of Greenland. This central and southern strip of West Greenland has developed primarily on granite and gneiss, Pre-cambrian Shield rocks. An exception is between Upernavik and the north side of Disko Bugt where there are Tertiary basalts similar in age and characteristics to those south of Scoreby Sund and in Iceland. They produce similar flat-topped tablelands. The landscape of the Shield areas is characteristically deeply glaciated with ice-smoothed rock knobs and hills. Along the coast are skerries associated with an arctic strandflat. The fiords penetrate the full width of the ice-free zone, effectively isolating the land into blocks. The intervening uplands exceed 7,000 ft at several points and these are glacier-covered. Southeast of Egedesminde the coast is especially divided by inlets and is much lower than elsewhere.

Topographically, Iceland is a young country, directly or indirectly entirely igneous in origin. It is characterized by straight lines and rough-hewn slopes, a lack of the softly rounded outlines characteristic of a mature landscape. Basalt-lava plateaus, which occupy nearly half the island, consist of a mosaic of tilted blocks. They are separated by fiords that have resulted from glacial deepening and widening of river valleys originally guided by fractures. The highest mountain is 5,046 ft and many summits exceed 3,000 ft.

The landscape is very different in a median zone that crosses the island from southwest to northeast and is filled with the Quaternary Palagonite formation. The landscape is generally that of a tableland with an inland elevation of 1,000 to 2,500 ft above which rise ice caps and volcanoes. Around the margins are lowland plains in which lava and ash are buried to a great extent beneath debris of glacier rivers.

Today only 1% of Iceland is covered with woods, although at the beginning of settlement about 1,100 years ago, about one-fifth of the country was wooded. Destruction of the trees was followed by intense soil erosion. More than half of Iceland is now a desert or semidesert, and the nakedness and lack of woods is one of the most striking features of the country.

The coasts of Iceland are of two distinct types. Rocky coasts characterize the plateau basalt areas of the northwest, north, and east. These coasts are highly irregular in outline, incised with numerous fiords and inlets, and have spectacular seaward cliffs facing the open sea; in places the cliffs exceed 1,600 ft in height. The fiords are shallower than the classic Norwegian type. Shallow deltas have formed at the heads of the fiords and curved shingle spits project from fiord sides where the tributary streams reach the sea. On the inner sides of the spits are good natural anchorages which are quite deep, and many of the villages and towns in the fiords are built on them.

The west coast is lower than the fiord areas, and the coasts of Faxafloi and Breidafjordur in particular are typical skerry. The southeast and south coast, from Hamarsfjordur in the southeast to Thorlakshofn, are sandy and smooth in ouline. Lagoons and long barrier spits and islands are common. These coasts have no natural anchorage. Most of the material building up the coast comes from debris-laden glacier rivers. The same type of coast on a smaller scale occurs where glacier rivers discharge at the heads of broad inlets in other parts of the country such as Heradsfloi in the northeast and Axarfjordur, Skagafjordur, and Hunafloi in the north. The bar along the south coast of the Snaefellsnes Peninsula consists mainly of shell sand.

Jan Mayen is wholly volcanic rocks, lavas, tuffs , and sands derived from them. The main landform is the volcanic cone of Beerenberg, the highest point on which is 7,500 ft above sea level. The southwest of the island is a lower hilly area that barely exceeds 2,600 ft. The connecting land between Beerenberg and the southwest hills is low and contains lagoons, separated from the sea by ridges of sand and gravel. The northernmost lagoon ,Nordlagunen ,is 130 ft deep. Sorlagunen, on the southeast coast, on the other hand, is so shallow that it dries out in the autumn. Large quantities of driftwood occur on the beaches. When not hidden in fog, Beerenberg may serve as an important landmark, visible from afar. There is no shelf around the island. The coast is steep and inaccessible over long distances, and skerries and pinacles occur in the sea close to those parts of the coast. Jan Mayen lacks sheltered anchorage and landing is exceptionally difficult.

From the sea Svalbard presents striking arctic scenery with rows of dark mountain peaks separated by bluish-white glaciers. The archipelago is composed of a major island, Spitsbergen, as well as Nordaustlandet on the northeast and many smaller islands. In the southern part of Spitsbergen many glaciers reach the sea, particularly on the east coast where they coalesce to form almost continuous glacier fronts with steep ice faces. Icebergs formed from these glaciers are of limited dimensions compared with Greenland or Baffin Bay icebergs. More than half of Spitsbergen and three quarters of Nordaustlandet are covered with glaciers. In addition to the two large ice caps in Nordaustlandet there are others in the northern part of Spitsbergen , north of Isfjorden, and in the southern and southeastern part of the island. In contrast, in the area between Isfjorden and Bellsund there are ice-free plateau mountains and ice-free valleys.

Spitsbergen is deeply incised by fiords of which Hornsund, Bellsund, Isfjorden and Kongsfjord-Krossfjorden are the most conspicuous on the west coast. Isfjorden is 55 miles long. On the north coast Woodfjorden and Wijdefjorden are the larger ones. Wijdefjorden is 60 miles long. The direction of the fiords is principally derived from the geologic structure.

A striking coastal feature is a low foreland in front of the steep mountain sides. At its greatest development it rises from approximately 100 ft below sea level to about 160 ft above it. The surface is remarkably even, sometimes with outcropping bedrock and sometimes covered with silt and gravel. The foreland is often swampy with numerous ponds and lakes. The width of the emerged foreland varies. In the southern part of Prins Karls Forland the strandflat extends across the island and is almost 5 miles wide. At Daudmannsyra to the north of the entrance to Isfjorden it is 6 miles wide. It is well developed between Bellsund and Isfjorden and from Isfjorden to north of Kongsfjorden. It has great extension also on the north coast of Spitsbergen. At some points, the foreland comes gently down to sea level, whereas at other places it faces the sea in steep cliffs 10 to 30 ft high. At Bjornoya, where the strandflat is well developed on the northern part of the island, its sea cliffs are about 100 ft high.

Along the fiords of Spitsbergen, particularly in the Bellsund and the Isfjord area, wave-built plains and terraces are common. These are depositional features formed by longshore drifting created by wave currents. Beach material is derived from streams which carry great quantities of debris seaward during the summer and from neighboring cliffs. Large cuspate forelands of sand and gravel are in this way formed and prograded by the addition of successive beach ridges. The beach plains slope in the direction in which the forelands prograde. Many extensive ridged beach plains have surfaces sloping from 130 ft above sea level down to the present-day beach, some of them even from 200 ft.

Bjornoya is geologically related to Spitsbergen but in many respects is dissimilar and there are no fiords, harbors, or glaciers. The island is triangular in shape with the apex at the south and is about 11 miles across. The north and northwestern sector is a lake-dotted plateau developed mainly on horizontal sandstones. It is cliffed although the highest part does not exceed 300 ft. In the southeast, partly on Hecla-Hoek rocks, are hills, the highest of which, not inappropriately, is Miseryfjellet (1,760 ft).

The greater part of northern Scandinavia geologically and topographically belongs to the Scandinavian Shield. The Precambrian rocks that underlie it are of great age (Archaean) and are mainly crystalline granites and gneisses; an exception is in the extreme northeast where somewhat younger (Proterozoic) sedimentary rocks are found between the head of Porsangen and Varangerenfjord.

Southwest from Nordkapp along the Atlantic coast is a belt of younger strongly metamorphosed sediments and intrusive rocks that were incorporated in the Caledonian folding and contain a greater variety of rocks than the Shield.

The land mass is highest along the zone of Caledonian folding and heights between 3,200 to 5,000 ft are common and reach 6,500 ft at a few points close to the Norwegian-Swedish border. From there the land surface slopes gradually southeast, towards the Gulf of Bothnia. Numerous rivers flow across this slope in wide valleys often containing long, narrow lakes to the Gulf of Bothnia. The rivers flowing from the drainage divide to the Atlantic are in contrast short and turbulent. Near the coast the land mass is deeply entrenched by steep-sided fiords and in places separated by sounds from the mainland notably in the Lofoten Islands. The combinations of highest mountains and deepest valleys (fiords) near the coast produces spectacular, rugged scenery in this area. There are numerous giant, cirque hollows on the mountain sides, formed by glaciers in the past. When excavated from many sides of a mountain massif, the cirques leave only a skeleton of ridges and summits. A good example is the island of Moskenesoy, the westernmost large island in the Lofoten group. During the Ice Age the mountains were stripped of deposits and appear rounded and ice-smoothed or as roughly weathered peaks and escarpments. At many places along the fiords raised marine clay and sand deposits, of late glacial or post glacial age, form ledges and plains and also constitute the fill in the lower part of many valleys. Such terraces are found at 65 to 230 ft and even higher, above present day sea level.

A second characteristic of the coastal topography is the strandflat, a belt of islands, skerries, and rocks and a rim of low foreland in front of steep mountain sides. The low foreland rises to a height of approximately 165 ft and may extend below sea level to a depth of about 100 ft. Generally the surface is neither flat nor continuous as the islands and peninsulas are separated by sounds and fiords. The summits rise to a uniform low height. Bedrock outcrops at many places on the strandflat, but often there is a thin cover of loose deposits ranging from clay to gravel.

The topography of the eastern part of northern Scandinavia, from Hammerfest to the Russian border, is of a different character. The interior is less mountainous and the relief lower. Most of the rolling surface lies between 1,000 and 1,500 ft above sea level. The hills drop precipitously to the sea and leave, in most places, no room for settlement.

B.6.2.2 Eurasian Arctic

The Barents Sea leads directly out of the Norwegian Sea; it is flanked by the European mainland, Novaya Zemlya, Zemlya Frantsa Iosifa and Svalbard, and is roughly square with sides about 400 miles. All the coasts have an arctic environment, but permanent land ice is restricted to islands on the north and northeast sides.

Several distinct physiographic provinces occupy the mainland coast and immediate hinterland on the south side of the Barents Sea. In the west, granites, gneiss and other crystalline rocks of the Fennoscandian Shield underlie Kolskiy Poluostrov and Karelia. The Shield is predominently a rocky, rolling upland with generaal elevations of 500 to 650 ft near the coast and rising to about 1,600 ft inland near the treeline. It is deeply dissected, especially near the Norwegian border where it is penetrated by deep, broad fiords that trend northeast-southwest. The land is covered with countless lakes and small, poorly organized streams; moraine ridges, eskers, and peat bogs are scattered over the surface. Along the coast there are often clearly marked terraces resulting from higher sea levels in the past. The surface of the Shield descends towards the east and disappears beneath younger sedimentary rocks of the Russian Platform on the east side of the White Sea. These rocks dip gently to the southeast; on them are developed broad plains crossed by escarpments that are highest near the White Sea coast. Karst topography is well-developed at several localities and the lowlands particularly are marshy.

Lowlands extend from the edge of the Shield to the Urals except where they are interrupted by the uplands of Poluostrov Kanin and Timanskiy Kryazh. The former is generally low-lying except in the north, where there is a ridge of highly dissected hills. The summits are low, averaging 500 to 850 ft; at Mys Kanin Nos, where the hills reach the sea, they are under 500 ft but even so, the headland is in striking contrast to the alluvial lowland in other parts of the peninsula. The only other rocky headlands are where the Timanskiy Kryazh reach the sea at Cheshkaya Guba. This range is formed of folded strata, and elevations in it reach 1,000 ft.

The lowlands of the Malozemel'skaya and Bol'sheze-mel'skaya tundras separate the Timanskiy Kryazh and the Urals. The terrain is marshy, with muskeg bogs and deep peat soils developed on thick postglacial marine sediments and glacial deposits. Inland, sand dunes up to 300 ft high, mark the outer margin of the glaciated area. The major feature of the coast are two large bays; Cheshkaya Guba and Pechorskay Guba. The latter is the estuary of the Pechora River, which has formed a complex delta with many small islands. Throughout, the coast is lined with spits and bars.

The Urals are a hilly upland composed of foldes rocks. The highest peak, Gora'Narodnaya, (6,200 ft) is 210 miles inland, and closer to the sea, few points exceed 5,000 ft. The mountains strike northeast until 68°N where they turn sharply to the northwest. The western slopes descend steeply to a low plateau, and then to the lowlands; the eastern slope is more gentle. The hills become much lower after the change in direction and 40 miles of lowland tundra separates them from the Khrebet Pay Kohoy, the most northerly part of the range on the mainland. The Pay Khoy and northern Urals have distinct glacial landforms including hanging valleys, tarns and ice-abraded rock zones.

The treeline extends from Varanger Peninsula in Norway, wast through the middle of the Kol'skiy Poluostrov, and then follows roughly the 67°N parallel to the Urals: consequently all the Barents Sea coasts have arctic vegetation, but the White Sea penetrates deeply into the forests. South of the tundra proper, forest-tundra merges slowly into taiga forests.

A single large island, Otrov Kolguyev is present in the southern Barents Sea. It is roughly circular, with a diameter of 50 miles. Altitudes are everywhere low and only reach 500 ft in the center of the island. Spits and sandbars line the coasts and separate lagoons from the open sea. Surface drainage is poor and low tundra mounds are widely distributed in marshes. In the north of the island the dominant vegetation is lichen and moss tundra; in the south it is shrub tundra, with dwarf birch and willow. Otrov Kolguyev and the adjacent mainland are underlain by deep, continuous permafrost.

In the southeast corner of Barents Sea, Ostrov Vaygach continues the Ural structure. It is separated from Yugorskiy Poluostrov by Yugorskiy Shar, a narrow strait 3 to 4 miles wide. The island is crossed by two northwest-southeast trending hill ridges which attain an elevation of 600 ft. Where they reach the sea, the hills produce low cliffs but elsewhere there are sandy beaches.

Novaya Zemlya is a massive land barrier that separates the Barents and Kara seas. It consists of two islands divided by the narrow, winding, steep-sided Proliv Matochkin Shar. The islands are a structural continuation of the folded sediments of the Urals that curves in a southwest-northeast arc for about 550 miles. Long, linear valleys, fiords and straight coastlines are characteristic of the topography and result from intense faulting. The eastern side of the islands is somewhat lower and less rugged than the west.

The southern part of Novaya Zemlya, for about 80 miles north of Proliv Karskiye Vorota is a plain that rarely exceeds 330 ft in the highest parts and is covered with wet, hummocky tundra. At about 72°N the interior is higher and rises gradually northwards for 120 miles at which point the island has become a steep-sided, flat-topped block with an upland surface reaching 3,300 ft. This sector is dissected by straight valleys, many of which have been drowned to form fiords, and one, Matochkin Shar, cuts completely across the upland.

Zemlya Frantsa-Iosifa is an archipelago of several hundred islands on the northeast margin of the Barents Sea between 80° and 82°N and 45° and 65°E. The largest island, Georga, is about 70 miles long, but the majority are much smaller. The cessation of igneous activity was followed by widespread faulting that disrupted the land mass and the surrounding continental shelf into the archipelago and intervening deep channels. The resulting plateau islands have steep, often cliffed shores and interior uplands that reach 2,300 ft, although 1,000 ft is more typical. The margins of the uplands have been eroded by ice to form cirques and glacier troughs, but over 80% of the land is ice-covered. Outlet glaciers reach the sea and produce icebergs, and in some localities form small ice shelves.

The West Siberian Lowland is a vast, flat plain, nearly 400 miles wide in the arctic sector, where it extends from the Urals to the Yenisey River. The lowlands have developed on horizontal sedimentary rocks that are buried beneath Quaternary glacial, marine and fluvial deposits. The monotony of the marsh-covered, ill-drained, tundra landscape is broken by occasional terraces, and low moraines. The highest elevations rarely exceed 500 ft and are found where bedrock approaches the surface between the Ob' and Yenisey estuaries.

The shores of the Kara Sea are deeply indented by estuaries to form peninsulas. In the west, the largest, Poluostrov Yamal, projects 240 miles northwards into the Kara Sea. It is a plain of sands and clays, the highest elevation being in the south-central part of the peninsula where a moraine reaches 330 ft. The west coast is occasionally terraced and has occasional low cliffs, but in general the shores are low, flat and swampy. The surface of the land is covered with lakes and swamps through which rivers appear to wander aimlessly. Ostrova Belyy, north of the peninsula is similar topographically. Along the east shore, Obskaya Guba is the drowned lower course of the Ob' River. It is over 320 miles long and has an average width of 40 miles. At its head the mouth of the Ob' has a limiting depth of only 8 ft, and seagoing ships cannot enter. A branch of the main estuary, Tazovskaya Guba extends southeastwards. South of it is Trazovskiy Poluostrov, and north of it, the very similar Gydanskiy Poluostrov. The latter is a wide peninsula between the Ob' and Yenisey rivers that is divided in the north by Gydanskaya Guba into two subsidiary peninsulas. The interior is slightly hillier than the remainder of the arctic lowlands, and ridges reach 600 ft. Along the coast the land is low and swampy, with maximum elevations less than 100 ft.

The estuary of the Yenisey is 200 miles long. It marks the edge of the lowlands as the land rises immediately east to the hills of Poluostrov Taimyr. The mouth of the river itself has a least depth of 33 ft and so is accessible to some sea-going vessels.

The treeless coast and hinterland of Poluostrov Taimyr are dominated by Gory Byrranga, a mountain range parallel to the coast and about 45 miles inland, with altitudes up to 4,000 ft. The uplands descend gently to the north forming rolling terrain. The west and central coasts of the peninsula are low-lying and highly irregular with innumerable offshore islands. In the east, near Mys Chelyuskin, the coast is generally hillier.

The Kara Sea contains hundreds of islands, the majority of which are close to the Taimyr coast. Ostrova Belyy, north of Poluostrov Yamal, is the westernmost large island. Other islands include Ostrova Uyedineniya in the central Kara Sea and Ostrov Wize and Ostrov Ushakova between Zemlya Frantsa Iosifa and Severnaya Zemlya. These islands are composed of recent deposits and have undulating terrain inland with low cliffs along their coast. There is little vegetation.

Severnaya Zemlya lies north of Mys Chelyuskin between 78°N and 81°N. There are four large islands in the archipelago and many small ones. Two-fifths of the archipelago is ice-covered. The most southerly island, Bol'shevik, is the second largest and is separated from the mainland by the 32 miles wide Proliv Vil'kitskogo. The western side is low while the east coast is high and often cliffed. The interior has low, rounded hills rising to about 3,000 ft. The remainder of the archipelago is separated from Ostrov Bol'shevik by Proliv Shokal'skogo which has depths of 600 to 900 ft and is suitable for navigation as it is deep, wide and often ice-free in the summer. Oktyabr'skoy Revolyutsii is the largest island. It resembles Ostrov Bol'shevik except that it has four small ice-caps in the corners of the island that are the remnants of a larger ice-cap that once covered the whole island; the highest elevation, 3,200 ft., is on the northern ice-cap.

Soviet coastal waters are effectively divided into western and eastern sectors by Polvostro Taimyr and Severnaya Zemlya. The eastern waters are further sub-divided by the Novosibirskiye Ostrova into a western sector, the Laptev Sea, and an eastern section, the East Siberian Sea, extending as far as Ostrova Vrangelya.

The coast and hinterland bordering the Laptev and East Siberian seas are composed of extensive lowlands backed by a highland region of mountain ranges and high plateau, several of which extend to the coast. There are numerous, north-flowing rivers; the largest, from west to east, are the Khatanga, Anabar, Olenek, Lena, Indigirka and Kolyma. The northern Siberian Lowlands extend eastwards from the lower Yenisey River, where they merge with the West Siberian Lowlands, to the Kolyma River; they form a lowland strip approximately 1,300 miles long that in places is 600 miles wide, but narrows to 6 miles where uplands approach the sea. In the west the lowlands occupy a basin that separates the hills of Poluostrov Taimyr from the Siberian Plateau. This section is occupied by the Khatanga River which is drowned in its lower part to produce an estuary more than 140 miles long. The mouth of the estuary is practically blocked by the large island, Ostrova Begicheva.

Between the Khatanga and Lena rivers two ridges break the monotony of the lowland. The Kryazh Pronchishcheva running parallel to the shore between the Anabar and the Olenek rivers is a folded range with elevations up to 1,000 ft. Kryazh Chekanovskogo, a ridge that rises to 1,650 ft ,runs northwest-southeast and separates the basin from the Lena River. Another hilly sector occurs east of the Lena where the Verkhoyanskiy Khrebet approaches the sea, but this is replaced east of Guba Buorkhaya by an alluvial lowland. The lowlands have developed on unconsolidated Quaternary deposits (sands and silts) that are underlain by continuous permafrost and which contain large quantities of ground ice. Pingos, thermokarst landforms, polygonal ground, thaw lakes and intricate periglacial drainage patterns are widespread. In some places the area of the lakes is equal to that of dry land, and in the summer, much of the lowland is impassable to surface travel. The majority of lakes are oval ,shallow (3 to 13 ft) and are rarely longer than 2 miles. The lakes and the rivers are interconnected and when the rivers are in flood they flow into the lakes, while at other times the lakes feed the rivers. The relief is monotonously flat with gently rolling tundra sloping imperceptibly to the north except where it is interrupted by low morainic ridges. In striking contrast to western Siberia, east of the Anabar, deltas rather than estuaries are characteristic of river mouths.

Tundra, marked by conspicuous pologinal patterned ground and composed of mosses and lichens with some herbaceous plants, dwarf birch and willow, occupies the coastal lowland. Taiga vegetation ,with larch and white birch, penetrates north along the valleys of large rivers and the treeline nearly reaches the sea in the vicinity of the Lena.

The Lena River is more than 2,000 miles long. The delta, the third largest in the world, is a complex of Quaternary sediments which begins 80 miles inland and has an area of 12,350 sq miles. It is made up of over a thousand islands and sandbars. The depth and position of the intervening channels are modified by tides and strong winds and the general fall of water level in the summer. The shape and number of islands is altered constantly by melting of ground ice, deposition and erosion. Frost mounds sometimes more than 100 ft high, frozen silt, and ice cliffs are widely distributed and the whole delta is a mass of thaw lakes.

Northeast Siberia, east of the Lena Delta is a combination of coastal lowlands, arch-shaped mountain ranges, isolated mountain groups and plateau. The uplands have developed on folded sedimentary rocks, with granite intrusions usually underlying the highest peaks. The mountains are deeply dissected but may have accordant tops. Summits above 6,000 ft are common; all were modified by glaciers during the Pleistocene, which produced hanging valleys, cirques, tarns, and moraines that descend to about 1,200 ft. The coastal lowlands (the Kolyma Plains) lack glacial features except for a small area south of the Ostrova Novosibirskiye.

The Peninsula of northeast Siberia, when considered as a whole, is dominated topographically by mountains, but only three ranges approach close to the Arctic Ocean. In the west, the Verkoyanskiy Khrebet, drop sharply to the Lena River, but slopes are gentler on the east side. Summits in the northern part of the range exceed 6,500 ft. East of the Kolyma valley, a second group of ranges descend towards the coast in steep escarpments crossed by deep canyons on the west side of Chaunskaya Guba. A third mountain group, the Hnadyrski Khrebet lies between the Anadyr basin and

(Paragraph continues on following page. The above inadvertently omitted by typist.)

the coast of the Chukchi Sea. The massif is divided at Zaliv Kresta. The western section is highly dissected by deep glacial valleys, and has alpine mountains with elevations in the center up to 5,900 ft. The eastern section is lower with rounded uplands and is crossed by broad northwest trending valleys. The northern part of the range backs the coast a short distance inland from east to Chaunskaya Guba to west of Mys Shmidta. Spurs from this range reach the coast and form cliffs and rocky points. The Yukagirskoye Ploskogor'ya; south of the lower Kolyma, has a much dissected surface and broad, shallow valleys. There are some isolated mountains with elevations up to 3,600 ft.

The Yana, Indigirka, and Kolyma plains, occupy most of the coastal region of the East Siberian Sea. They extend from Guba Buorkhaya east to the mouth of the Kolyma River and penetrate inland along the middle and lower courses of the Indigirka and Kolyma. They are all close to sea level and are broken only by lakes and the meandering distributaries of the three rivers after which they are named. The coastal strip is a low, swampy tundra, with scattered areas of higher elevations, occasional low cliffs, and some hills inland. The three large rivers have extensive deltas. There are many lagoons and bays on the coast detached from the sea by spits and bars and mostly with narrow entrances. North of the Kolyma are six small rocky islands, the Ostrova Medvezh'i, which are the only islands in the East Siberian Sea located an appreciable distance from the shore.

Separating the Laptev and East Siberian Seas are numerous low islands known collectively as Ostrova Novosibirskiye. They are the much eroded extension onto the continental shelf of the folded fountain ranges of northeastern Siberia. The sediments on the Ostrova De-Longa were buried by basalt lavas which today form the flat upper surface of the islands and the precipitous black cliffs. Ostrov Benneta, the largest of the group, is approximately 7 miles wide and the highest point, which is glaciated, exceeds 1,300 ft.

The islands of the main and southern groups are generally low except for the western third of Ostrov Kotel'nyy where hills reach 1,225 ft. The uplands in this sector are developed on limestone, slate and basalt each of which has a different scenery. The valleys are straight and deep, a characteristic that has been attributed to faulting ,and the coast has several deep embayments. The central part of Ostrova Koteln'nyy, known as Kemlya Bunge is a sandy, low, flat plain, which is only slightly above the level of the ocean, and is occasionally flooded by sea water.

In the southern group, the interior of Ostrov Bol'shoy Lyakhovskiy is also hilly, but elsewhere in the archipelago the terrain is flat and gently undulating and does not rise above 300 feet. The plains are developed on sands and silts which contain massive ground ice. The ice is concentrated in shallow valleys. Where it is found along the coast it forms cliffs that in some places are 165 ft high. The ice has melted rapidly in the 20th century leading to important coastal changes and the disappearance of two small islands. The coastal waters adjacent to the plain are exceptionally shallow. The islands have little vegetation and only the better watered valleys have closed tundra.

Ostrov Vrangelya is 70 miles off the Poluostrov Chukokskiy. It is crossed by two mountain ranges that separate the northern and southern lowland. The summits are either flat or gently domed, the remains of an ancient peneplain. The southern range is shorter than the northern but contains the highest peak, Gora Sovetskaya (3,600 ft). In general the relief is complex and highly dissected, with lattice drainage and some large, flat intermontane valleys. West of the ranges are flat, low hills under 1,000 ft high; on the east is the Eastern Plateau , an undulating upland with isolated flat-topped peaks with a maximum elevation of 650 ft. The northern range is a gently rolling area, little dissected, with relic mountains less than 2,000 ft high. It slopes gently to the northern coastal plain, called Tundra Akademii.

Tundra Akademii is a flat plain of Quaternary marine sediments covered with fluvial sands, gravels and loams up to 100 ft thick. There are also raised terraces left from marine transgressions in the interglacial periods. The surface is cut by shallow river valleys; thermokarst depressions and lakes are common. There is a coastal bar on the northeast coast, with a string of small lakes inside it. The Southern Coastal Lowlands, divided in the center by Zaliv Krasina are relatively narrow and taper to the east until they disappear. The western half is similar to Tundra Akademii, except that the rivers are more entrenched and have steeper gradients. On the east this tendency is more pronounced.

The easternmost Siberian mainland is dominated by the confused mountain mass of the Chukostskiy Khrebet. The mountains are complex fold structures, and in general, the central and eastern parts of the range near the coast have rounded outlines with elevations from 2,500 to 3,300 ft. In the interior the highest summits exceed 6,500 ft. Although the mountains are lowest near the Chukchi Sea, broad coastal lowlands are restricted to a single plain that extends southeast from Amguyema River for 140 miles. The eastern part of this plain is drowned to form Kolyuchinskaya Guba. Elsewhere along the coast hills frequently reach the sea, isolating small plains.

The coast shows evidence of submergence and has an irregular outline where the sea has entered valleys. The shallow inshore waters favor the development of barrier bars and long sectors have straight sandy bars and spits which separate lagoons from the open sea. The entire Siberian coast lies beyond the treeline, and characteristically in the lowlands, damp marsh tundra has developed on a terrain underlain by permafrost and with inumerable small lakes.

East of Kolyuchinskaya Guba the mountains come closer to the coast and at Mys Serdtse-Kamen elevations near the coast reach 2,000 to 3,000 feet. Mys Dezhneva is a limestone mountain reaching 2,540 ft connected to the mainland by a low isthmus.

B.6.2.3 North Pacific

At the west end of the Aleutian arc are the Komandorskiye Ostrova, a group of four islands, 90 miles from the Kamchatka coast. The terrain of the two largest islands, Ostrova Beringa and Ostrova Mednyy, is rugged, but although there is frequent seismic activity, there are not active volcanoes and the maximum elevation is only 2,100 ft. The coasts are rocky except at the north end of Ostrova Beringa.

The Asiatic land mass on the west side of the Bering Sea generally experiences a harsher environment than Alaska. This is reflected in the vegetation which is restricted to tundra and arctic desert in Siberia as far south as 60°N, and by the boundary of continuous permafrost which is more than 80 miles farther south on the west side of the Bering Sea than the east. Although the geological history of northeast Siberia has many points in common with Alaska the main topographic elements are not as clear-cut or symmetrically distributed. The major physiographic regions are the Poluostrov Kamchatka, the Koryakskiy Khrebet, Poluostrov Chukotskiy, the Anadyr' Lowlands and the north and west coasts of the Sea of Okhotsk.

Kamchatka is a peninsula, more than 600 miles long, that separates the Sea of Okhotsk from the Bering Sea and the northwest Pacific Ocean. Two main mountain ranges, divided by the broad valley of the Kamchatka River dominate the Peninsula. Forty volcanoes are known, almost all on the eastern side of the Peninsula; fourteen are active. The highest almost reaches 16,000 ft and many exceed 10,000 ft. In spite of the rigors of the land the summers are sufficiently warm in the lowlands to support forests, and permafrost occurs only sporadically near sea level. The west coast lowland with maximum widths of 40 miles is considerably broader than the east; the western marshy coastal plains are fringed in the central and southern sectors by lagoons and offshore sand and gravel bars. Ostrov Kariginskiy is a continuation of the more easterly interior mountain range. It resembles the nearby east Kamchatka coast, being rugged and with maximum elevation of about 3,000 ft.

The Koryakskiy Khrebet extends the rugged topography of Kamchatka northeastwards in a series of mountain chains that run roughly parallel to the coast. Summit heights commonly exceed 5,200 ft in the interior ,and the coastline has occasional peaks with elevations reaching 4,000 ft. The coast resembles the Pacific side of Kamchatka with large promontories where the main ranges meet the sea.

The Poluostrov Chukotskiy high land block resembles the Alaskan Seward Peninsula and consists of fold mountains. The ranges have no dominant trend, although there is a tendency for the larger valleys to be aligned in a northwest direction. The mountains have rounded contours with summit elevations between 2,800 and 3,300 ft. There is strong evidence in the landscape of former intense ice sheet erosion, particularly in the deep U-shaped valleys, and ice smoothed hill tops. The southeast coast where the mountains are cut off at Bering Strait has a deeply indented fiord coastline. The nearby Diomede Islands in Bering Strait have steep rocky shores and rounded relief.

The Sea of Okhotsk is produced where the continent of Asia is separated from the Pacific Ocean by Kurilskiye Ostrova and the Poluostrov Kamchatskaya. The western boundary south of 54°N is formed by the island of Sakhalin.

Ostrova Kurilskiye run in a narrow concave arc along the southern boundary of the sea. They are part of the Pacific volcanic arc which extends from the Aleutians through Kamchatka, the Kuril Islands and Japan; there are over 100 cones in the island chain, of which 38 are active today. The small islands in the chain are usually single volcanic cones, while the largest are formed by a complex of connected cones. All the islands are composed of ash and lava in alternate layers except for the northernmost, Ostrov Shumshu, and the southermost, Kunashir Island which are geologically related to the adjacent mainland and are of sedimentary origin.

The island of Sakhalin is over 500 miles long and is separated from the mainland by the Strait of Tartary. Two mountain ranges, an eastern and a western dominate the relief of the island; both subside in the north leaving an extensive lowland with elevations up to 2,343 ft at Cape Yelizaveta. The mountain ranges are separated by a central lowland, 3 to 18 miles wide, which starts at Guba Terpentiya and strikes approximately northwards before turning east with the Tym River to form a coastal plain. The eastern range is higher than the west; altitudes in the center reach 6,600 ft. The shores of Sakhalin are relatively steep and rocky in the southeast, except where the central lowland meets the sea at Guba Terpentiya. The western range is formed of several parallel ridges, of which the westernmost is the highest and has the sharpest relief. In the north end the mountains become lower and more gentle, and the coastal plain broadens to a substantial lowland which finally merges into the northern plain. The eastern coast in the center and north of the island is lined with almost continuous sand bars.

The mainland coast of the Sea of Okhotsk is over 1,000 miles long and although it is generally high and even mountainous, there are conspicuous differences in several sectors. In the extreme west, the hill ranges meet the coast at right angles, and deep bays and several large islands are the result. From the Uda River northeastwards to Okhotsk more than 300 miles away, the interior ranges lie parallel to the coast which is exceptionally smooth and steep with no coastal plain. Between Okhotsk and Magadan the coast is more irregular and there are small plains, but they diminish again east of the latter town; and the northeastern coast is highly irregular with rugged peninsulas and only occasional small plains with lagoons and bars along the coast.

B.6.2.4 North America

On the North American side of the pole the Arctic Basin approaches close to the continent east of Point Barrow, and thus divides the coastal waters into two parts. The enclosed channels of the western Canadian Arctic Archipelago, which form a separate unit are also naturally divided into two by Parry Channel. Consequently there are four basic sectors in the western North America Arctic:

1) Chukchi Sea (Wrangel Island to Point Barrow)
2) Beaufort Sea (Point Barrow to Banks Island)
3) Southwestern Canadian Arctic islands and mainland, and
4) Western Queen Elizabeth Islands.

Alaska

With the exception of the Arctic Coastal Plain, Alaska is wholly within the Western Cordillera physiographic province of North America. The Cordillera in Alaska consists of parallel complex mountain ranges separated by plateau and lowlands. The west coast of Alaska cuts at right angles across the northwestern margin of the province forming large peninsulas that correspond roughly to mountain ranges. The two main peninsulas south of the Chukchi Sea are the Seward and the Alaska. Between them, a third blunt peninsula bounded by Norton Sound and Bristol Bay terminates in Nunivak Island and has superimposed on it the Yukon and Kuskokwim deltas.

Seward Peninsula projects nearly 160 miles towards Siberia between Kotzebue and Norton sounds. The geology is extremely complex and the interior is dominated by a mass of uplands and mountain ranges. Across the center of the Peninsula, the York Mountains and associated ranges reach nearly 3,000 ft. In the southern part the Kigluaik and Darby mountains include the highest mountains in the Peninsula. From the sea, the overall appearance of the south side of the Peninsula is rugged and the relief in the higher coastal areas is 1,600 to 2,300 ft. There is, however, considerable variation and there are coastal lowlands between Cape Douglas and Solomon (including the Nome area) and between Moses Point and Koyuk.

The upland coasts are sharply indented where valleys reach the sea. These are filled with alluvial deposits, mainly gravels, in contrast to the north coast where silts are dominant. Along part of the coast, gravel spits and bars are characteristic whilst elsewhere rock terraces are prominent features.

The coastal highlands between Norton Sound and Bristol Bay are bisected by the Yukon-Kuskokwim drainage complex. On the north side a highland sector encircles the eastern and southern coasts of Norton Sound from Norton Bay to the Yukon Delta. In this area, hill ridges strike in a northeasterly direction roughly parallel to the sound. The crests vary in height between 1,[illegible]00 and 2,000 ft and have been much rounded by frost action and solifluction. Only small valleys reach the sea. The southern highland sector extends southwards from the Kuskokwim Delta to the lowlands surrounding the Nushagak River at the head of Bristol Bay. The hill ridges have the same orientation as that south of Norton Sound, but here are at right angles to the coast; known as the Kuskokwim Highlands, the rocks are more resistant and the scenery more rugged than in the Norton Sound area. The hills rise 1,000 to 2,000 ft at the coast and increase inland to heights of 3,000 to 5,000 ft. At Togaik Bay the hill ridges project into the sea and form a number of small offshore islands with maximum elevations of more than 800 ft. The ridges are separated by wide, flat valleys filled with thick deposits of alluvium. Many of the landforms were modified by intense mountain glaciation ,particularly in the interior.

The twin delata of the Yukon and Kuskokwim rivers is the largest lowland in the area. It is characteristically arctic, treeless and underlain by permafrost, innumerable thaw lakes, deranged drainage, tundra polygons and pingos. The Nushagak-Kvichak lowlands at the head of Bristol Bay have a deep mantle of Quaternary deposits. The coastal relief is flat and monotonous; drainage is poor and continually shifting, resulting in countless lakes, marshes and sloughs.

The islands in the Bering Sea vary in size from the St. Lawrence Island, that is over 80 miles long , to rocky islets. The majority are of volcanic origin. The only appreciable sedimentary rock outcrops are on St. Lawrence Island, where they form rugged relief in the southwest of the island, that contrasts sharply with the remaining two-thirds of the island, which is low with many lakes and marshes. There has been volcanic activity in recent times, and a 2,070 ft volcanic cone dominates the center of the island. The south coast is lined with barrier beaches, cuspate forelands, spits and lagoons.

St. Matthew and the Pribilov islands are rugged, hilly, volcanic islands with elevations of 1,000 to 1,500 ft in the interior. Sea ice, nearly continuous fog and absence of harbors contribute to navigation hazards around them. St. Matthew is a series of low peaks rising from the sea and connected by low sand and gravel spits. Elevations are between 1,000 and 1,500 ft. Of the five islands in the Pribilov group only two, St. George and St. Paul, exceed 20 sq. miles in area, and are inhabited; mainly of basaltic lava outflows, their highest elevation is less than 1,000 ft.

Nelson Island lies in a re-entrant of the Yukon-Kuskokwim Delta. It is developed on sedimentary rocks and lavas which rise to about 100 ft and contrast conspicuously with the nearby deltaic plains. The coasts facing the sea are marked by precipitous headlands linked by beaches. Nearby Nunivak Island is of volcanic origin and is surrounded by bluffs between 50 and 400 ft high. Occasional dune formations occur along the coast.

The Aleutian Range extends for 1,700 miles from near Anchorage to the coast of Kamchatka. The range is continuous in the east where it forms the Alaska Peninsula but is largely drowned in the central and western sectors. It exhibits the characteristic pattern of the geologically young island arcs found around the Pacific with high relief, steep topography, strong earthquakes and active vulcanism; almost 50 volcanoes have been active in the last two centuries. Both the islands and the Peninsula are formed of lava outflows, volcanic cones and calderas, intermixed with sedimentary strata formed by erosion of local igneous deposits. For the most part the coasts of the Aleutian Range are lower on the north side, the topography rising sharply to the south. Poor drainage, dunes, sandbars and lagoons characterize the Bering coast of the Peninsula. The peaks reach their maximum altitude on Unimak Island (9,978 ft) and in the Alaska Peninsula (11,200 ft).

The only significant lowland area in northeast Siberia is a zone of subsidence between the Koryaskiy and Chukotskiy highlands that has been filled by deposits of the Anadyr River. Although the terrain is generally flat, the volcanic and igneous rocks that underlie the alluvium project through, producing low hill ranges at several points. Drainage in the lowland coastal areas is inhibited, and marshes, thaw lakes and lagoons are widespread. Bars and spits are characteristic of the coast which resembles closely the northern coast of Alaska and the Chukchi Peninsula.

The northern Alaskan coasts are also dominated by mountain ranges but at many points they are far from the sea, and the lowlands are correspondingly larger than on the Siberian side. The principal mountains are the Brooks Range. They are relatively subdued in the western area where they are known as the de Long Mountains. The range is about 80 miles wide and has many peaks over 3,000 ft; a few rugged, glaciated summits exceed 5,000 ft. The mountains approach close to the sea between Point Hope and Kotzebue Sound. On their north flank is a broad foothill zone which is close to the coast northeast of Cape Lisburne. North side of the mountains and the foothills is the flat Arctic Coastal Plain that attains its widest development near Barrow.

The Beaufort Sea coasts from Point Barrow to Cape Prince Alfred comprise three major regions: a western unit corresponding closely to Alaskan territory; the mainland plain of the northwestern Canadian mainland, including the Mackenzie

Delta; and,the coast of Banks Island. Each region may be further subdivided as discussed below.

Northwestern Alaska comprises a mountain sector, the Brooks Range; a Foothills Province; and a Coastal Plain. The same threefold pattern is found east of Point Barrow. The Brooks Range is 120 miles from the sea south of Point Barrow, but the distance lessens until along the U.S.-Canadian border it is only 25 miles away. The relief is often extremely rugged with east-west trending ridges. In the northern part accordant summits rise 7,000 to 8,000 ft. Extensive glaciation occurred in the Pleistocene, and there is evidence of ice tongues 35 miles beyond the mountain front. In the higher northeastern parts of the mountain group, small cirque and valley glaciers are common.

The southern portion of the Foothills Province consists of irregular, isolated hills, east-trending ridges, and mesa-like mountains. In the north the structure of the foothills is much simpler and consists of folds of Cretaceous sediments. The hills rise from 600 ft in the north to 1,200 ft in the south.

The Arctic Coastal Plain slopes very gently from an average of 600 ft near its southern border to sea level along the coast. The plain has developed on ice-indurated sand, silts and gravel of the Gubik formation. Ice volumes are often as high as 70% in the upper 20 ft of sediment; consequently if melting of the permafrost occurred, subsidence of the land surface might be as much as 12 ft in some areas, and could cause some low lying coastal areas to be inundated. Organic material in the sections consists of plant and animal remains, some very old, preserved in the frozen ground. At a few localities, notably in the White Hills, lignite and coal are exposed. The Plain is poorly drained, and is covered with thousands of lakes and swamps. Thaw lakes are one of the most striking features of the Coastal Plain Province. Covering over half the surface, the lake basins enlarge by thawing the frozen ground, and often coalesce. They range from a few feet to several miles long, and are up to 20 ft deep. Many of the lakes are elongated with their long axis having a pronounced N 15°W orientation.

Most of the rivers on the arctic slope originated in the Brooks Range. They cross the plain ~~the~~ in shallow braided channels and have built large deltas in the Arctic Ocean. Most are navigable during the short summer by small boat in their lower reaches. Large seasonal fluctuations are common, with bank-full conditions during the spring compared to virtually no flow in the braided channels during the summer and fall. On the Colville River nearly half of the annual flow takes place in a three week period following break-up. The rivers are generally frozen for eight months of the year.

The arctic coast is basically a barrier coast with sand spits, offshore bars, and islands alternating with low bluffs for virtually the whole distance. Between the bars and the mainland are shallow lagoons suitable for the passage of small boats. On the mainland, behind the lagoons is commonly a bluff 15 to 40 ft high formed in frozen sediments. Point Barrow is the most northerly point on the coast. It is a low spit of sand and gravel about 4 miles long and 400 yards wide. The coast is receding at most localities, the average loss reaching 10 ft per year. Wherever the coast forms a point or is otherwise well exposed to the sea, shore retreat can exceed 30 ft a year. Point Drew, Cape Simpson and Brownlow Point, east of Flaxman Island have exceptionally rapid retreat.

The mainland coast of northwest Canada, adjacent to the Beaufort Sea consists of three parts. In the west, the Yukon Coastal Plain, which is only 2 miles broad at Demarcation Point where the Cordillera is closest to the sea, widens to over 15 miles nearer the Mackenzie. It is a monotonous land of open, often wet tundra crossed by the numerous streams that drain the British Mountains. The Plain has developed on silts and sands which have been eroded by streams and waves, and deformed by glaciation into modcrate relief. Low cliffs are common along the coast; they are 20 to 40 ft high and are receding in many places at about 1 ft per year. The sand from the cliffs is deposited on the beach, which may be as much as 50 ft wide. The sand is slowly transported offshore.

The Mackenzie Delta occupies a former deep embayment in the coast that has been filled with alluvial deposits brought down by the river. The waters north of Shallow Bay are particularly shoal, with depths rarely exceeding 20 ft as far as 17 miles offshore. Between the Blow River and the inner parts of Shallow Bay recession reaches 10 ft per year. The Delta consists of thousands of islands, surrounded by an intricate network of channels and interconnecting lakes. At break-up, the water level rises 10 to 15 ft in the middle parts of the Delta, and less in the outer parts. During onshore stroms, and at flood stage in early spring, the Delta is water-covered for tens of miles inland and only branches of submerged willows outline the channels. The coastline is barely visible from the sea as the outermost alluvial islands are at sea level. Some of the higher islands, such as Kendall and Garry islands, which are outside the recent Delta, rise 200 ft above sea level. They are the eroded, detached portions of an older delta.

East of the modern Delta, the coast is formed by an old complex Pleistocene delta which includes all the Tuktoyaktuk Peninsula. The terrain is undulating, slopes seawards, and is disrupted only by low swampy areas and higher sandy ridges. East of Atkinson Point, half of the surface is covered by shallow rounded lakes, many of which are oriented with their long axis in a north-south direction. The coast is low, sandy spits and parabolic sand dunes are common, and in places recession reaches 10 ft a year. The most conspicuous features are the pingos which are concentrated in groups in the area around Kugmallit Bay. The coast from

Anderson River north to Baillie Islands is indented by estuaries, indicating submergence, and the estuarine part of Harrowby Bay is bordered by prominent terraces. Waters offshore are shallow along the whole coast from Demarcation Point to Cape Bathurst, partly because of fluvially transported silt and sand from the larger rivers. Sea level fluctuates with changes in wind and pressure conditions; fluctuations in Harrowby Bay reach 8 ft.

The Thesinger Bay coast of Banks Island resembles the Yukon coast; cliffs are up to 60 ft high in the east and 125 ft close to Cape Kellet. The fine sand and silt cliffs are cut rapidly back by marine erosion, and the sediments form a protective mudbank just below tide level. Occasional gaps are eroded by watercourses in the ridge-like coastline, and the lower reaches of some rivers include sandbars and lagoons. Except for two prominent ridges with bluffs immediately north of Cape Kellett, all of the west coast of Banks Island is low. The rivers, of which several rise near the east coast, enter the Beaufort Sea in flat open valleys and often have braided channels and shallow lakes. Between the valleys, the terrain is rolling and rises to a height of several hundred feet a few miles inland. Along this coast the unconsolidated sediments containing ice lenses are being rapidly cut back, helped by a slow sinking of the land. Along the Barnett Bay coast, offshore islands provide protection from wave and ice erosion, and crustal sinking has led to a drowned coastline.

The whole of the area is north of the treeline although in Coronation Gulf, northern extensions of the boreal forest project north along the rivers and in the case of Coppermine River, reach to within a few miles of the sea. Along the mainland coast west from Kent Peninsula, there is a southern variety of arctic tundra with thickets of willow and birch bushes in sheltered localities and a rich proliferation of species. East of Kent Peninsula and on the southern islands the vegetation is visibly poorer, but nonetheless forms closed tundra communities. In the northern sector of the islands where the summer temperatures are significantly lower and drought conditions develop on the deep shattered bedrocks, there is a striking impoverishment of the species and there are broad areas of rock deserts.

In a general sense the terrain of this part of the Canadian Arctic has developed on Precambrian Shield rocks on the mainland, and younger sedimentary, platform rocks, overlying the Shield on the islands. The two most conspicuous arches are in west-central Victoria Island, culminating in the Shaler Mountains, and along the eastern margin of the sector where a denuded arch produces the uplands of Somerset Island and Boothia Peninsula.

The Canadian Shield on the mainland west form Dease Strait is composed predominantly of sedimentary and volcanic rocks, and the topography and terrain are consequently unlike most shield areas. Everywhere the shores and coastal lowlands are overlain by quantities of silt and fine sand deposited in a high level sea that existed at the close of the glaciation. Several distinct coastal sections can be recognized resulting primarily from the type of bedrock.

Along the south side of Dolphin and Union Strait there is rolling hill country developed on sedimentary rocks. Although altitudes inland exceed 2,000 ft, the drift is generally so thick that rock outcrops are rare, except where they form cliffed headlands. Slopes are long and gentle and the impression from the coast is rather of rolling lowlands than an upland region. The mainland coasts of Coronation Gulf, Bathurst Inlet and Dease Strait form a second region that contrasts strongly with the coasts farther west. The igneous rocks commonly form spectacular cliffs, in many cases rising several hundred feet out of the sea. The lavas form strings of elongated islands in Coronation Gulf and Bathurst Inlet which have a cliffed side and a gently sloping backface. Typical of this part of the coast is the great variety of rock types and consequent landforms.

The shores include cliffs, volcanic shingle strandlines rising in steps from the water to more than 650 ft and flat, silt beaches.

South of Queen Maud Gulf the dominant rocks are granites of the Canadian Shield. The land slopes gently to the north and passes imperceptibly beneath the sea producing shoals and many islands. In summer the inshore water is extremely muddy, and the coast is one of the most difficult in the Canadian Arctic for ship operations. Similar crystalline rocks are exposed around Chantrey Inlet, but the relief is somewhat greater and low cliffs and bluffs rise out of the water to form hills several hundred feet high. Adelaide Peninsula, south of Simpson Strait, is similar to the large islands north of the mainland. The shores are shallow and contain numerous islands, the majority of which are submerged drumlins.

The southwestern islands have uplands in two areas: the first in central and eastern Banks Island, and western and west central Victoria Island; and a second on the eastern margin in western Prince of Wales Island, Somerset Island and Boothia Peninsula. Between the two regions is a broad lowland, which shows many minor differences but has an overall terrain uniformity.

The southern highland zone reaches Amundsen Gulf in Nelson Head and Cape Lampton; heights exceed 2,000 ft and cliffed coasts or cliffs fronted by a narrow coastal strip are typical. This area contrasts strongly with the northeast upland which is basically a tableland that is between 1,000 and 1,500 ft in the east; it drops to the west where it has been dissected by deep valleys, some of which are drowned forming Castel and Mercy bays. The central lowlands of Banks Island are rolling plains with few rock outcrops and a surface covered with glacial drift, sand and gravel.

The western Victoria Island uplands are separated from Banks Island by Prince of Wales Strait which is about 8 miles across at its narrowest and more generally about 12 miles wide. In the northwest corner of Victoria Island, corresponding roughly to Prince Albert Peninsula, there is rolling hill country in the interior, but the shores are usually low-lying and driftcovered, although at a few localities near the coast, altitudes may exceed 1,000 ft. Southeast of Prince Albert Peninsula, roughly as far as a line from the north side of Prince Albert Sound to Hadley Bay, is an area that resembles the Coronation Gulf sector with cliffed escarpment coasts, particularly in the north, and gently dipping back slopes. In the southwest, overlooking Prince Albert Sound, the relief is not as great. Characteristic features of the western upland of Victoria Island are the long inlets, some of which penetrate nearly 80 miles into the island.

The eastern upland includes a strip of hills, not more than 4 miles wide on the east side of Prince of Wales Island, that exceed 1,000 ft in height and reach the coast in steep slopes and cliffs. Across Peel Sound or Somerset Island a similar Shield coastline extends between Aston Bay and Wrottesley Inlet. Somerset Island is essentially a tableland. In Boothia Peninsula Precambrian rocks form the hilly backbone of the Peninsula with limestone hills and plains on either side.

Between the two uplands the lowlands of the Victoria Basin are nearly 250 miles across. The central, lowest part is drowned and forms M'Clintock Channel and Victoria Strait, which separate Prince of Wales and King William islands from Victoria Island. The coasts are remarkably uniform over considerable distances. Typically, the beaches are formed of shingle; behind them the coast rises in strandlines produced during the emergence of the land in postglacial times; the elevated beaches are separated by low, marshy hollows and shallow lagoons. At a few localities, glacial deposits modify the general effect because of the presence of drumlinoid islands and occasional low morainic hills. Glacial landforms are more prominent in land because even where the till is thick on the coast, the surface was washed over by the postglacial sea and strandlines were impressed on them. The most impressive of these hills is in the Rawlinson Peninsula of western Prince of Wales Island. In detail the land rises at varying rates away from the sea. In some places, as in the southern part of Victoria Island, the plains fail to rise above 200 ft for many tens of miles whereas in the northeast corner of Victoria Island they slope steeply up from the water.

The Queen Elizabeth Islands have the shape of a right-angled triangle with a 650 mile base along Parry Channel. East of the 90° meridian including Ellesmere and Devon islands, access is from Baffin Bay. Within the remainder of the area are five islands, each with an area greater than 2,000 sq. miles and many scores of smaller islands.

The three largest southern islands, Melville, Bathurst and Cornwallis, have a variety of topography as they cut across the three main geological regions of the archipelago. Melville Island, the largest of the three, is also the highest with points in excess of 3,000 ft. In the southwest is a plateau rising to over 2,000 ft and deeply cut by fiords and steep-sided valleys. Cliffed coasts with surfaces as high as 1,000 ft are comparable to the upland coasts of the central Arctic. A belt of folded rocks that has developed parallel hill ridges and valleys runs from northwest to southeast across the center of the island. The northwest coast of Melville Island consists of fiords, but elsewhere the coasts are much lower with deltaic river mouths and long, smooth coastlines. The Bathurst group of islands show, with Melville Island, similarities with a southern plateau and hill ridges traversing the remainder of the island. Cornwallis Island, the smallest of the three, is topographically more uniform. The southeast corner is a plateau that dips rather gently towards the north into low rolling country in which perhaps the most conspicuous element, as on all the other islands, is the absence of close vegetation associations. Grinnell Peninsula is somewhat higher; but in other ways is topographically comparable to the group.

The central islands, including Prince Patrick, Mackenzie King, Borden and the two Ringnes islands are low-lying with only occasional points above 1,000 ft. Close to the coast and particularly in Prince Patrick Island, gravel surface deposits are widespread; elsewhere the soil is generally derived from the underlying rocks and often contains much fine-grained material which is particularly susceptible to solifluction. Indeed, the solifluction in this area is probably more striking than in any other part of the Canadian North. Only occasionally, and notably on Ellef Ringnes is there higher ground, which reaches the sea in cliffs. Appearing at several points on the islands are the low hill domes developed around gypsum intrusions.

Axel Heiberg Island differs strikingly from the other islands. It is essentially a mountain mass broken into two parts through the center. Both the northern and southern uplands are covered with ice caps. The western side of the island contains deep valleys and fiords towards which valley glaciers flow from the main icecaps. Between the fiords are peninsulas with local ice caps and often extremely rugged scenery. In contrast the eastern and northeastern sides of the island are generally lower, consisting of low hills through which outlet rivers from the main ice cap flow towards Nansen and Eureka sounds. This coast is generally lower.

The east North American Arctic is centered around the waters of Baffin Bay and Davis Strait. Topographically the coasts may be divided into:

1) the Canadian-Greenland Shield highlands,
2) the horizontal, sedimentary rock plateau around Lancaster and Jones sounds, and
3) the folded mountain coasts of Axel Heiberg and Ellesmere islands, and northwest Greenland.

Ellesmere, Devon and Axel Heiberg islands form the most northerly and the highest part of the mountain rim of eastern Arctic Canada. Ice caps are widely distributed and movement by land and sea is restricted by ice and the terrain. The highland block is deeply penetrated by fiords and wide channels that split it into separate islands and topographical units. The main height of land is near the east coasts and consequently the major fiords open westwards into the Sverdrup Basin.

Complex mountain ranges developed principally on the folded sedimentary rocks occupy a large part of central and northern Ellesmere and Axel Heiberg islands. The higher peaks have an average elevation of 6,600 ft and in the United States Range, exceed 9,500 ft. Although many of the fiords have high cliffed sides, narrow coastal lowlands occur at many points. The mountains are generally ice-covered and outlet glaciers reach the sea at the head of several fiords. A lower, essentially glacier-free region of plateau and hills, roughly 30 miles wide , crosses the mountainous region from Robeson Channel to Fosheim Peninsula.

Between Buchanan Bay in eastern Ellesmere Island and Cape Sherard at the entrance to Lancaster Sound, the most northerly sector of the Canadian Shield forms the Canadian coast. Rugged alpine mountains with peaks reaching 6,600 ft are nearly covered by ice caps and the only ice-free areas are close to the sea and occasional nunataks.

Central Devon Island and the south coast of Ellesmere Island west of Starnes Fiord are backed by plateau of horizontal sedimentary rocks having upper surfaces at 1,500 to 2,000 ft. There are many small fiords often with cliffed sides. Elsewhere narrow coastal plains are typically covered with elevated shingle strandlines.

Prince Regent Inlet and its extension as the Gulf of Boothia is the most easterly of the large sounds that penetrate south from Parry Channel and in so doing, separate the islands of the southern half of the Canadian Archipelago. On the west side, a structural arch forms Boothia Peninsula and Somerset Island. South of Bellot Strait the coast is mainly developed on shield rocks that produce low hills with a maximum elevation of about 1,000 ft. Between Thom and Pelly bays the Shield is strongly fractured resulting in long, steep-sided inlets and rectangular islands. At two points, Simpson Peninsula and northwest Boothia Peninsula, Palaeozoic limestone forms coastal plains characterized by elevated shingle strandlines inland. In the south-east of the island, the coast rises slowly inland to low hills, but north of Cresswell Bay, the cliffs over 1,000 ft high are practically continuous.

The east side of the Gulf of Boothia is in several respects a mirror image of the west coast. In the southern half, the west side of Melville Peninsula is a Shield coast where hills reach the sea, except for a narrow limestone plain in Committee Bay, which includes Wales Island. North of Fury and Hecla Strait, hilly terrain is replaced by limestone plains from Agu Bay to north of Bernier Bay. Farther north both sides of Bordeur Peninsula have naked rock hills and plateau with cliffs and steep shingle slopes forming the coasts.

Borden Peninsula has greater topographic variety than the peninsula and islands to the west, but it is basically the same. It consists of complex, barren plateau and hill ranges. Over long distances the coasts are cliffed or have narrow coastal plains; only at the heads of Eclipse Sound and Admiralty Inlet are the coasts lower and provide access to the interior.

The eastern coasts of Bylot and Baffin islands are an area of scenic magnificance with high, ice-covered mountains rising abruptly from long fiords that penetrate far into the interior. The fiords attain their greatest development in the north between Pond Inlet and Cape Henry Kater where many are more than 60 miles long. The coast is flanked by a continental shelf 16 to 24 miles wide. A limited number of soundings are available from the fiords. In Inugsuin Fiord there is a sill at the entrance at a depth of 325 to 340 ft whereas the maximum inner depth is 2,050 ft and at the head of the fiord the water is still 330 ft deep. Eglinton and Sam Ford fiords also appear to possess sills at 330 ft and 475 ft respectively.

Topographically this northern sector is divisible into two zones:

1) extensive foreland composed to stratiffed glacial material extending on the seaward side of the east coast mountains from Cape Henry Kater northwards to Cape Hunter, and
2) fiords and mountains.

The latter face the open sea north of Cape Hunter and the coastal scenery is composed of sheer cliffs, mountain peaks, fiords and deep glaciated transection valleys. Coastal cliffs in the Buchan Gulf tower 5,000 ft above the sea. The highest mountains occur in the middle sectors of the fiords and elevations of over 5,000 ft are common. Farther west the mountains become more massive, the amount of glacial erosion decreases and the fiord heads lead steeply onto the rolling interior plateau at 1,500 to 2,500 ft. Between Royal Society Fiord and Coutts Inlet the alpine peaks of the middle fiord area are replaced by steep-sided, flat-topped mountains that are usually sites for ice-caps when the mountains exceed 3,000 ft. Numerous glaciers descend to sea level from these sources especially in the middle fiord areas.

South of Cape Henry Kater the coast recedes to form Home Bay. Although the mountains are massive and summit elevations range from 2,500 to 3,600 ft, the topography is more subdued that farther north and there are few glaciers. From the southeast side of Home Bay to Cape Mercy at the entrance to Cumberland

Sound, the coastal scenery is again spectacular with fiord walls rising sheer to 4,000 ft in places and mountains with summits exceeding 5,000 ft. Many peaks are alpine in character although flat-topped mountains characterize the land around Okoa Bay, north of the Penny Ice Cap. Fiords are numerous and penetrate up to 35 miles inland. Most of the high tops are covered by ice caps from which glaciers descend to and near sea level. North of Cape Dyer the coast trends southeast-northwest, but this changes to S 30°W at the Cape.

The southeast side of Baffin Island extends for 290 miles between Cape Dyer and Resolution Island. The coast is penetrated by two major inlets, Cumberland Sound and Frobisher Bay, while farther south an even larger break in the continental margin forms Hudson Strait. The north shore of Cumberland Sound is deeply indented with fiords, with mountains close to the sea that exceed 5,000 ft. The head of the sound is lower than the sides, and valleys through hilly terrain provide routes to Nettiliing Lake and the Foxe Basin Lowlands. The summits rise again on the south side towards Hall Peninsula, which has numerous fiords, large rocky islands and uplands close to the sea that are ice-covered. A similar coastal pattern occurs around Frobisher Bay. The north shore is broken by many fiords: at the head a lowland leads northwest towards Amadjuak Lake while the south side is an almost unbroken line of cliffs.

The east coast of Canada from Bylot Island to the Atlantic Provinces is essentially a highland region that blocks penetration into the interior of the continent. At two points the continental margin is breached by submerged depressions. In the north between Baffin Island and Labrador, Hudson Strait leads westwards to drowned basins in the center of the Canadian Shield which form Hudson Bay and Foxe Basin. The second gap between the southern edge of the Canadian Shield and the Appalachians constitutes the Gulf of St. Lawrence. Between Hudson Strait and the Strait of Belle Isle the continent faces the northwestern sector of the Atlantic Ocean known as the Labrador Sea. All the interior seas and the coastal belt of the Labrador Sea have arctic or subarctic environments and heavy sea ice late in winter and spring.

Foxe Basin is a circular depression of great geological age in the Canadian Shield. Almost all the islands in the basin are developed on limestone and in few places do they rise more than 100 ft above sea level. Occasionally bed rock is exposed on the islands but generally a mantle of frost-shattered limestone boulders forms the surface. The low relief and the effects of the postglacial marine transgression restrict drainage, and marshes and ponds are widespread. Old sea beach ridges are found almost everywhere on the islands recording the continuing slow emergence of the land. Extremely shallow water occurs around many of the islands and those parts of the mainland which are formed of similar rocks. Offshore bars, spits, mud flats and lagoons are a dominant element on the shores. The largest island, Prince Charles Island discovered in 1948, is in the northeastern part of the basin. Rowley, Bray, Kock, Jens Munk and Spicer islands, together with Baird Peninsula, have similar topography. Air Force Island resembles the others with the exception of a Precambrian rock outcrop at the northern end.

The east coast of Melville Peninsula from the vicinity of Hecla and Fury Strait to beyond Parry Bay is also a low-lying area of limestone, with very little variation in relief and a typically flat, monotonous topography. The limestones are replaced by Precambrian crystalline rocks 25 to 30 miles west of Hall Beach. On the granites and gneisses of this group, rugged hills and uplands with a high proportion of exposed bedrock, contrast sharply with the lowlands.

The largest plain around Foxe Basin is the vast flat or gently sloping area in west Baffin Island known as the Great Plain of the Koukdjuak. A striking characteristic of the plain is the profusion of small, near circular, coalescing lakes, ponds and marshy depressions that have resulted from the melting of ground ice combined with the effect of wind-directed wave action on the pond margins.

On the southwest side the plain passes imperceptibly from limestone onto granites. After many miles the bedrock ridges break the terrain and the remainder of Foxe Peninsula is mainly an area of crystalline granites and gneisses forming diverse hilly topography with elevations over 1,000 ft above sea level.

Southampton Island combines the two basic relief types of Foxe Basin (and Hudson Bay); while the north and northeast coast consists of highlands of Precambrian rocks with steep cliffs rising abruptly from Foxe Channel, the south and southwestern part of the island is made up of gently sloping limestone plains and plateau.

Hudson Bay is a larger version of Foxe Basin. A central depression in the Shield now forms the Bay. The result is a wide variety of landforms which may be classified into four major relief types:

1) areas of rugged topography with outcrops forming hilly sectors of widely varying relative relief;
2) subdued areas of flat aspect often at low level and sometimes covered by shattered rock debris or glacial material;
3) lowlands that have developed on Palaeozoic limestons; and
4) more restricted areas formed on metamorphic rocks that occur in folded belts or belts of steeply dipping strata and are mainly restricted to the east side of Hudson Bay where they produce prominent escarpments and ranges of hills.

The west coast of Hudson Bay between Roes Welcome Sound and Churchill is Shield country. In the north a narrow, rocky coastal plain gives way to an interior plauteau and a hilly area with elevations over 1,000 ft near Wager Bay. South of Chesterfield Inlet there is a striking change in physiography. The Hudson Bay Lowlands are an enormous area of low relief and poorly developed drainage. A great part of the area is composed of wet muskeg and swamp with islands of unconsolidated glacial material. This region which is over 180 miles wide has a very shallow gradient towards Hudson Bay. Extensive raised marine beaches and deposits are found great distances inland. The Hudson Bay Lowlands continue to the head of James Bay, but on the east side of the Bay, Precambrian granite-gneiss rocks again appear. Inland from the complex island-studded east shore of James Bay, the land rises over 500 ft within about 40 miles; the area is a complex of lakes, rivers and poorly drained muskeg.

On the east side of Hudson Bay, north of Cape Jones the Precambrian Shield rocks are fringed along the coast by a series of island escarpments which extend as far as Cap Dufferin near Port Harrison. These escarpments also occur on the mainland where they exceed 1,500 ft above sea level in places. They have steep east-facing slopes or cliffs and gentle western slopes towards Hudson Bay. The interior of the central part of the east coast is a rocky upland with numerous lakes and a complex drainage pattern. Elevations 40 miles inland from Richmond Gulf exceed 1,500 ft. Northwards, hieghts decrease as far as the latitude of Cape Smith, where a bold range of hills trends southwest-northeast with heights over 1,000 ft. The northwestern part of the Ungava Peninsula is marked by the impressive cliffs of Cape Wolstenholme while inland a relatively even plateau reaches an average height of 1,400 ft.

Seventy miles offshore on the 60th parallel are the Ottawa Islands. They are the northernmost of a complex chain of islands extending southwards almost as far as James.Bay. The largest group is the Belcher Islands.

Hudson Strait is essentially a parallel-sided basin, about 65 miles wide and nearly 400 miles long. The south side is broken by the deep embayment of Ungava Bay. With the exception of the limestone plateau that forms Akpatok Island in Ungava Bay, all the coasts are developed on Shield rocks. The north shore is everywhere rocky, cliffed and often very rugged. The coast rises slowly from the southeast corner of Baffin Island to about 1,000 ft in the center of the Strait around Big Island. Farther west there is an overall decrease in altitude and the coast becomes shallow and is deeply indented with many islands. The land is higher on the south side of Foxe Peninsula. The Quebec shore of Hudson Strait is high and cliffed between Cape Wolstenholme and Wakeham Bay, but farther east it decreases in height, and low hills and rock plains form the west side of Ungava Bay. The east coast of the Bay is low, but rises close to Cape Chidley to become hilly.

Northern Labrador, between Nain and Cape Chidley is mountainous; in the Kiglipait, Kaumajet and Torngat ranges, peaks in excess of 3,300 ft not uncommonly rise directly from the sea, and the highest points are over 5,000 ft. The coast is deeply indented by fiords and a few small glaciers manage to survive.

South of Nain the coast gradually becomes lower and hills, bare rock ridges, and intervening rocky lowlands characterize the coast as far as the Strait of Belle Isle. The coastline is low, irregular and fretted with numerous islands and bays. Everywhere it is barren even though trees are found a few miles inland. Sandwich Bay is a major indention and Hamilton Inlet is even larger. In the south the land rises to 1,500 ft within 8 miles of the coast although along the faulted coastline of the Strait the coast is quite bold. Several apparently drowned river valleys are notable features in the south. Lake Melville is contained in a steepsided inlet with elevations approaching 2,000 ft on the south side in the Mealy Mountains.

Physiographically, Newfoundland is several plateau that decrease in elevation in a northeasterly direction. The highest surfaces are in the west and southwest, where boulder-covered uplands reach 1,500 to 2,000 ft, while on the northeast side rocky plains and low hills disappear under the Atlantic.

Two topographic sectors may be recognized on the Atlantic Coast between Strait of Belle Isle and St. John's. In the north is the straight fault-guided coast of the east side of the northern peninsula. Inland the ground rises rapidly to the barren summits of the Long Range Mountains. Rivers have eroded deeply into the mountains and the drowned lower sections of the valleys break the straight coast with many small inlets. At the north end of the peninsula elevations are less and embayments are larger than farther south.

The remainder of the northeastern coast comprises rocky headlands and peninsulas formed where hill ranges reach the sea; they are separated by submerged lowlands that include Notre Dame, Bonavista, Trinity and Conception bays.

The coasts of the Gulf of St. Lawrence exhibit strong contrasts as a result of the presence of three very different structural regions in the area. The Quebec shore is formed by the southern edge of the Canadian Shield; this is usually a scarp face 1,000 to 1,500 ft high that drops gradually towards the Strait of Belle Isle. The precipitous scarp is separated from the sea at many points by a narrow, rocky coastal plain, a few miles wide. Rivers pouring out from the interior deposited large sand deltas when the sea was higher than today, and these deposits often mantle the coastal lowland.

The south and east coasts of the Gulf have developed on rocks of the Appalachian-Acadian Province. Some rocks, notably granites, have proved highly resistant to erosion and form upland coasts, while others, particularly sandstones and shales, are weak and form the lowland coasts that ring much of the southern part of the Gulf. In the first category are the high, often cliffed, coasts of Gaspe Peninsula, Cape Breton Island and much of the west coast of Newfoundland. Lowland coasts are present the eastern shores of New Brunswick, Prince Edward Island and northwestern Nova Scotia; they are generally sandy with highly developed spits and beach bars which virtually close the long estuaries that characterize the coasts. In a few areas the lowlands end abruptly in sandstone cliffs, 10 to 100 ft high.

The Magdalen Islands, which are part of the lowland group, consist of isolated, sandstone and glacial till islands joined by sand bars. Anticosti Island in the north of the Gulf differs from other parts. The island is developed on Palaeozoic limestone and has low cliffs and rocky ledges along the coasts.

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