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# ARCTIC FORECAST GUIDE

U. S. NAVY WEATHER RESEARCH FACILITY
BUILDING R-48, U. S. NAVAL AIR STATION
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# FOREWORD

This manual has been prepared under Task 16, "Polar Analysis and Forecasting Techniques," and is intended to fulfill the need for a basic arctic forecast guide for the navy meteorologist.

This publication is a comprehensive reference which describes the physical features and the meteorology (climatology, structure and behavior of weather systems, and the analysis and forecasting of meteorological parameters) of the Arctic. It provides the necessary general information which enables the forecaster to bridge the gap between middle-latitude and arctic meteorology. In addition, much of the information included in this report will prove useful for the duration of the forecaster's tour of duty in this region.

Dr. Richard J. Reed, Associate Professor of Meteorology at the University of Washington and Technical Consultant to the Navy Weather Research Facility, has drawn from past and recent work of his own and others and integrated this information into this complete "Arctic Forecast Guide." Mr. John M. Mercer performed the final edit of this manual for the Navy Weather Research Facility.

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# INTRODUCTION

To the forecaster trained in middle latitude meteorology the Arctic is a somewhat strange and forbidding place located near or beyond the boundaries of his map. He is aware that there is no sharp line that divides the regions of his common interest from the less familiar areas to the north. Moreover, he is aware that basic physical principles are the same everywhere, that differences between high and middle latitude meteorology are more a matter of degree than of kind. Nevertheless, when first faced with the problem of making operational forecasts in the Arctic, he is apt to feel like the proverbial fish out of water as he strives to put his middle latitude experience to work in a new and unfamiliar environment.

The main purpose of the present work is to provide the forecaster who is newly assigned to the Arctic with background material that will ease his period of adjustment, though it is hoped that many of the ideas expressed herein, as well as some of the techniques, will prove useful to him beyond the period of his apprenticeship. It is conceivable that some readers may find this manual useful for purposes other than that for which it was intended. However, the limited objectives of the author should be kept in mind at all times.

It is necessary from the outset to establish boundaries for the region which is to be included under the term "Arctic". Much fruitless effort could be spent in trying to frame a "correct" definition of the term, but a little reflection will show that any definition must of necessity be somewhat arbitrary and suited to the purposes at hand. Because our main concern here is with a fluid medium, the polar atmosphere, it has seemed advisable to make the boundaries as broad and simple as possible. With this in mind we have set the 60th parallel as the outer limit of our area of interest.

It has not seemed appropriate in a work of this sort to burden the reader with an extensive bibliography. The writer would therefore like to acknowledge that the information contained herein comes from three main sources: (1) "The Dynamic North" (1956), especially the article "Meteorology of the Arctic" by Petterssen, Jacobs, and Haynes (March, 1956), (2) "The Arctic Circulation" by Hare and Orvig (1958), and "Arctic Weather Analysis and Forecasting" by Reed (1959). Considerable pains have been taken in the latter work to reference the contributions of other authors.

In addition to the basic sources mentioned above, a number of other papers are listed in the bibliography. These are mainly recent contributions which were not reviewed in the writer's previous work. Their listing here is not meant to imply that they are necessarily more important than other articles which are not directly mentioned.

## 1. PHYSICAL FEATURES OF THE ARCTIC

Although the Arctic is popularly thought of as a region of eternal cold and snow, it is in reality an area of considerable climatic diversity, exhibiting a variety of interesting and changing weather phenomena. Much of this diversity is a consequence of the varying shape and character of the underlying surface. It is appropriate, therefore, to begin a discussion of the meteorology of the Arctic with a brief review of the geography of the region, laying principal stress on the nature and configuration of the surface and on its physical properties. A unique facet of the polar environment is the extended periods of daylight and darkness. The chapter concludes with a brief treatment of this topic.

### 1.1 Surface Types

Because of the relatively broad limits which we have chosen to set on the term "Arctic", a number of diverse surface types will be enclosed within our sphere of interest. Over the Arctic Ocean and the adjoining seas is the perpetual and nearly unbroken pack ice. Greenland and other elevated regions of lesser size are covered or spotted with shields of ice known as glaciers. Along the fringe of the arctic seas is the tundra, a treeless region of marshy soil covered with small shrubs and thick growths of mosses and lichens. Further inland, where a more pronounced moderation of climate occurs in summer, the tundra gives way to forests of varying extent and density. And in the area between Greenland and Norway is a large expanse of open water maintained by the mild ocean and air currents which invade that area.

In the following paragraphs we will consider in greater detail the distributions of the various surface types enumerated above, pointing out various facts which may be of interest and use to the forecaster. The reader is advised to make frequent reference to figures 1.1 and 1.2, which contain names and locations of pertinent geographical features and boundaries of the different surface types.

#### 1.1.1 Pack Ice

By definition the term pack ice refers to any area of sea ice, no matter what form it takes or how disposed.

From figure 1.2 it is seen that the Arctic Ocean is affected by ice throughout the year. The ice covers generally more than 95 percent of the area, the extent of open water being greater in summer than in winter. In winter thermal stresses within the ice and wind stresses on the surface cause continual fracturing and the formation of narrow lanes or leads of open water. At the frigid winter temperatures the leads freeze over rapidly. In summer the newly frozen leads melt open, allowing the ice to move more freely and bringing a temporary halt to the fracturing except for minor breaks along the rough edges of the leads.

In modern times the pack ice has averaged about 8 to 10 feet in thickness. Broad areas of more or less flat ice of uniform thickness are interspersed with ice ridges or hummocks which may extend several feet above the general level of the ice and ten's of feet below. The ridges are formed by the jamming of large pieces of ice or floes one against the other during periods of fracturing and may occasionally rise to 30 feet.

The earliest measurements of the ice thickness, taken before the turn of the century, gave values of about 12 feet, indicating a decrease in thickness in recent decades. This trend parallels a warming trend at arctic weather stations which has been remarked on by numerous authors.

The upper layers of the pack ice attain their coldest temperatures in February, but the lowest layers continue to cool until June so that the ice attains its maximum thickness just before the summer melt period commences. In June and July the snow which has accumulated since the previous summer melts, and in July and August an ablation of the ice itself takes place. Generally, the thickness of the pack is diminished about 2 to 3 feet by the melting. Most of the melt water drains into the ocean through cracks and holes but a significant amount of it collects in pools on the slightly undulating surface of the pack, forming small lakes. Personnel stationed on the ice find the summer melt period a time of considerable annoyance and discomfort.

In late August the melt season comes to an end, and snow begins to spread its insulating blanket over the surface. Once the snow cover is complete the surface of the Arctic Ocean behaves, as far as energy transfer is concerned, just as adjacent land surfaces.

Ice conditions in the adjoining seas are for the most part similar to those in the Arctic Ocean. Except near the shore, where a piling

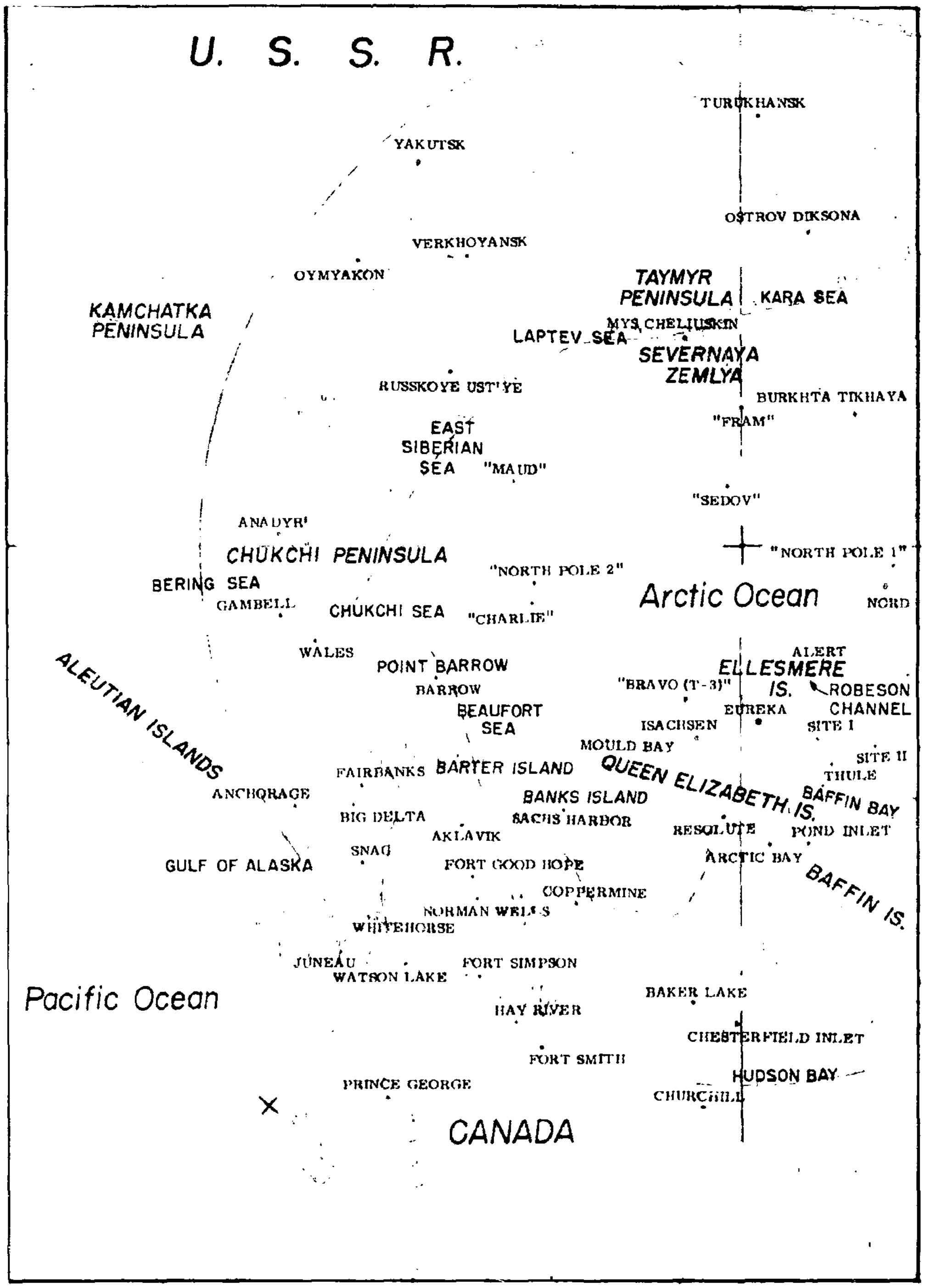



**Figure 1.1. Geographical Location Map of the Arctic.**

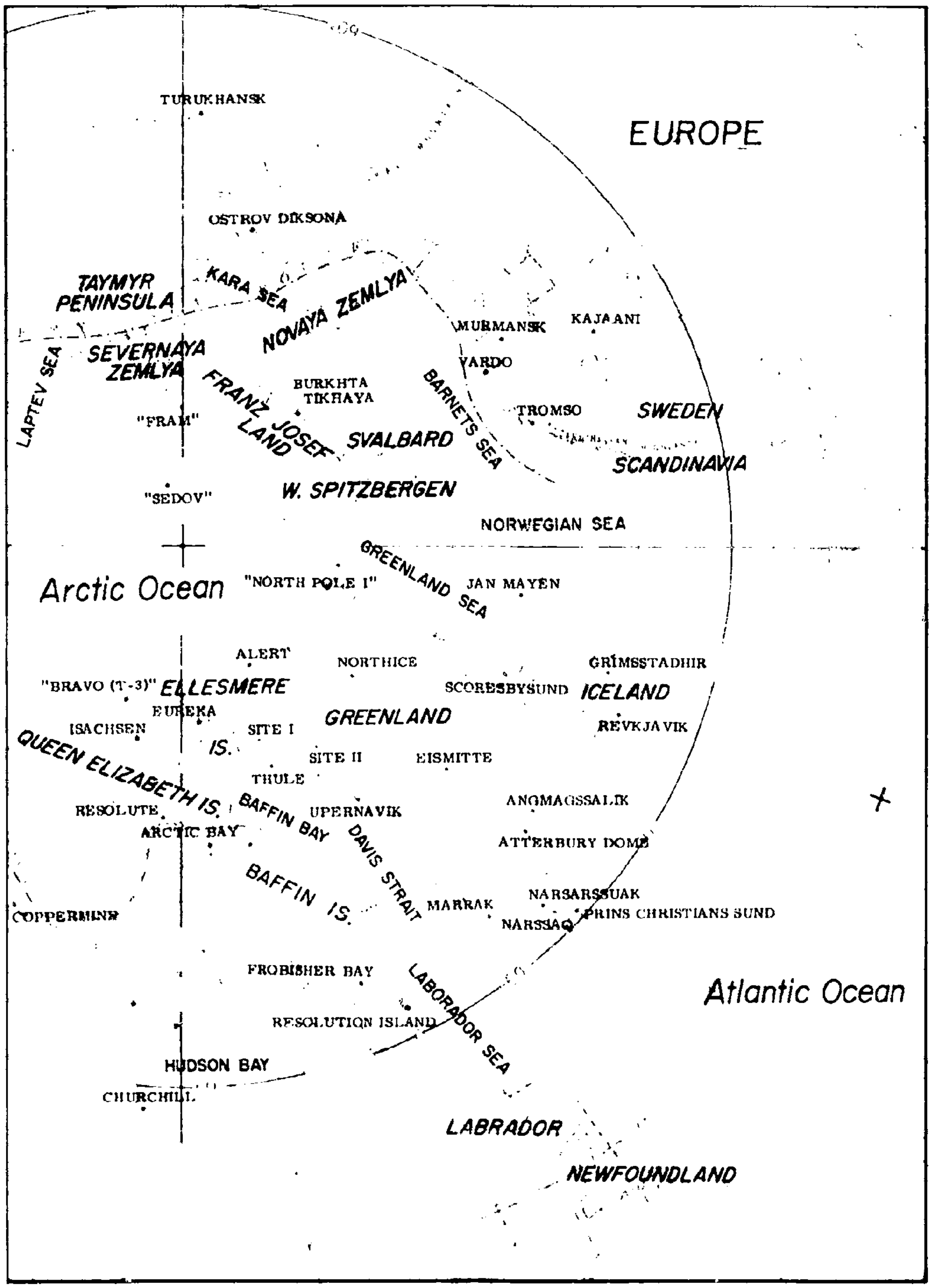



**Figure 1.1. Geographical Location Map of the Arctic.**

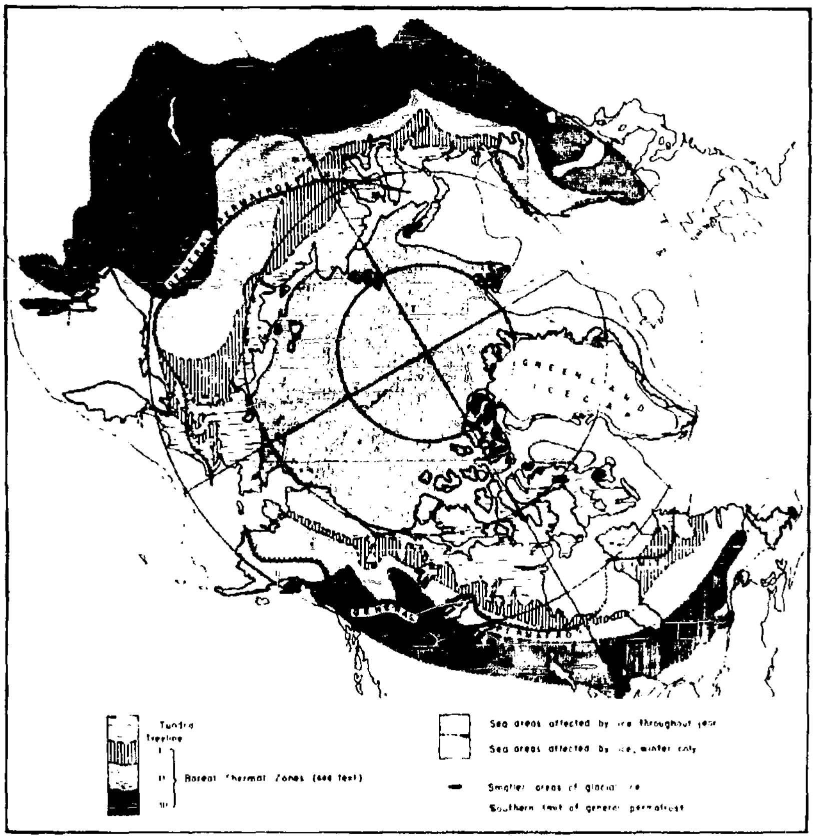



Figure 1.2. Some Geographical Distributions of Arctic "Surface Cover" Due to Climatic Controls. (After Hare and Orvig [3].)

up and rafting of the ice often occurs, the ice is somewhat thinner than in the inner Arctic and the summer melt is a bit greater. In summer large open areas appear near the shores and the coastal waters become navigable.

The amount and distribution of open water are influenced to a considerable degree by the prevailing wind pattern so that large fluctuations may occur from month to month or year to year. Generally the southern portions of the Chukchi, East Siberian, and Laptev Seas are navigable from late June or July through August or September. The Kara Sea stays open until October. The shipping season in the Beaufort Sea is short and confined to the late summer.

Shipping operations along the Northern Sea Route are important to the Soviet economy, and much effort is expended in charting ice conditions by aerial reconnaissance and in using icebreakers to keep the sea lanes open. The undertaking of annual supply missions to DEW-line stations along the Alaskan and Canadian coasts has spurred increased attention to ice problems in the American sector of the Arctic.

Other bodies of water which are covered in whole or in part with sea ice during at least a portion of the year are the Barents, Norwegian, and Labrador Seas, Baffin Bay, Hudson Bay, and the Northwest Passage. The Barents Sea is nearly free of ice in August but in late April or May, at the time of greatest ice extent, only the southwestern portion remains open. The southern and eastern sectors of the Norwegian Sea remain open throughout the year, though an ice strip of varying width hangs along the coast of

Greenland.

The eastern two-thirds of the Labrador Sea is ice free throughout the year, except along the Greenland coast. During the winter the ice grows outward from the Baffin Island, Labrador, and Newfoundland coasts to a width of a few hundred miles. The most favorable conditions for navigation in Baffin Bay occur in August and September. From November through April the bay is ice covered.

Despite earlier opinion to the contrary, Hudson Bay is completely frozen from January to June. The change from open water to ice and snow cover in late fall and early winter is marked by dramatic changes in weather conditions at stations on the east shore of the bay. The Northwest Passage is mostly icebound, though some navigation is possible in this region in August and September.

With the exceptions noted above, the polar seas may be regarded for three seasons of the year as solid snow-covered surfaces similar in their physical and meteorological characteristics to adjacent land areas. In summer, on the other hand, the seas are best pictured as melting sheets of ice which maintain air temperatures at or near the melting point.

1.1.2 Glacier Ice

Glaciers occupy only about 10 to 15 percent as much area as the pack ice of the polar seas, and more than four-fifths of this area is contained in the massive Greenland Icecap. The remaining glaciers are found on Ellesmere and Baffin Islands and other smaller islands in the Queen Elizabeth group (Canadian Archipelago); on Novaya Zemlya, Iceland, Jan Mayen, Svalbard, Franz Joseph Land, and Severnaya Zemlya; and in mountainous regions of Scandinavia, Alaska, and western Canada. The locations of many of these glaciers appear in figure 1.2.

By their nature glaciers occur in conjunction with high ground or, as in the case of Greenland, are themselves mountains of ice. The subglacial floor in Greenland is shaped like a saucer with a large portion of the interior dipping below sea level. The average thickness of the ice is of the order of 5,000 feet and the maximum known thickness is 11,000 feet.

In its effects on radiation and on heat and moisture transfer glacier ice is much like sea ice except that at high elevations temperatures stay below freezing and the ice remains covered with snow throughout the year. At lower elevations strong inversions may develop near the ice boundary in summer so that above freezing temperatures are common at instrument level.

1.1.3 Tundra

This is the name given to the treeless plains of northern Canada, Alaska, Siberia, Russia, and Scandinavia. The boundaries of the tundra are shown in figure 1.2.

From October through May the tundra is decked with snow and therefore has no unique effect on meteorological processes. In summer, when the snow is gone, the exposed surface consists of a marshy soil, covered with a dense growth of mosses, lichens, and small shrubs. The subsoil of the tundra is permanently frozen. It is this perennially frozen ground, or permafrost as it is now called, which is primarily responsible for the swampy nature of the tundra.

During most of the year, then, the surface of the tundra is undifferentiated from the previous surfaces we have been considering. From June through September, however, it is vastly different, possessing a much higher absorptivity and a greater ability to convert the absorbed radiation into sensible heat.

1.1.4 .The Boreal Forest

Roughly speaking the first patchy areas of woodland begin where the temperature of the warmest month reaches a figure of 10° C. (50° F.). The tree line may be more accurately delineated by taking into account the temperature of the coldest month as well.

Three forest zones are depicted in figure 1.2, Thornthwaite's potential evapotranspiration[1] serving as the defining parameter. In Zone I tundra and thin woodland are intermingled. Zone II consists of open woodlands of spruce, pine, or larch carpeted with mosses and lichens. The main Boreal forest of conifers and some hardwoods is contained in Zone III.

The length of the period of snow cover in the forested regions varies from 8 months in the north to 6 months in the south. In the southerly zone the snow usually melts by mid-April and the ground remains essentially bare until late October. Clearly, this belt has polar characteristics during only half the year.

[1] The amount of water that would evaporate from a plant-covered soil if there were no restriction in water supply in the root zone.

## 1.2 Surface Properties

From the meteorological standpoint the most important properties of the surfaces discussed in the preceding section are their albedo, heat capacity, and thermal diffusivity, since, along with changes of phase, these properties determine the manner in which heat is partitioned and stored. The product of the heat capacity and the square root of the thermal diffusivity defines an additional property of importance, the conductive capacity. The first factor determines the heat that can be stored in unit volume per unit temperature change and the second factor is known from heat conduction theory to be proportional to the depth of penetration. Thus their product, the conductive capacity, measures the total heat capacity of the affected volume. When the conductive capacity is large, the temperature change accompanying heat addition or subtraction is small.

Table 1.1 contains estimates of the foregoing properties for the various types of surfaces considered in the previous section.

The albedo of snow is high, ranging from 50 percent to 80 percent or even higher depending on the freshness. Ice too has a high albedo, though the figure given here is probably larger by 10 percent to 20 percent than the average value for the pack ice in summer when sizeable leads and puddles are present. Because of the low elevation of the sun, the albedo of water surfaces is relatively large in the Arctic, ranging from 4 percent when the sun is at its zenith in the subarctic in summer to 100 percent when the sun is on the horizon. The figure of 15 percent is believed to represent a reasonable average value. The tundra and forest areas have the smallest albedos. Where lakes and ponds are prevalent, the values are somewhat larger than given.

TABLE 1.1
Physical Properties of Various Surfaces.

| Surface | Albedo percent | Heat capacity cal./cm.$^3$/°C. | Thermal diffusivity cm.$^2$/sec. | Conductive capacity |
|---|---|---|---|---|
| New snow | 80 | 0.03 | 0.006 | 0.002 |
| Old snow | 50 | 0.22 | 0.003 | 0.012 |
| Ice | 70 | 0.45 | 0.012 | 0.05 |
| Tundra | 10 | 0.7 | 0.003 | 0.038 |
| Forests (green) | 5 | 0.6 | 0.004 | 0.04 |
| Forests (snow covered) | 25 | values appropriate for old or new snow | | |
| Water | 15 | 1.0 | 100 | 10 |

The variation in heat capacity from one type of surface to another is not great except that new snow has a much smaller capacity than the other surfaces because of its low density. The thermal diffusivities are likewise fairly uniform except for water in which case the conduction is assumed to be governed by eddy rather than molecular processes and is therefore very large.

The conductive capacity of new snow is small indicating, for instance, a tendency for the surface temperature to decrease rapidly when heat is lost by radiation. On the other hand, the ocean has an exceedingly large conductive capacity and therefore undergoes only slow temperature variations. The other surfaces considered also have rather small conductive capacities, indicating that their temperatures tend to fluctuate substantially in response to varying radiation conditions.

## 1.3 Orographic Features

The varying height and configuration of the earth's surface is a major factor in weather prediction. Effects of mountains can be seen on all scales of motion from the global circulation patterns to local wind systems of the foehn and katabatic types. Storms undergo characteristic changes in direction and intensity upon approaching mountainous terrain; in addition, amounts and distributions of clouds and precipitation are much influenced by orographic features.

Because of the relatively large extent of the polar seas and the existence of lowlands and plains over much of northern Canada, Russia, and eastern Siberia, the Arctic is not a region in which orography plays a dominant role. In at least two sectors, however, the effects of high ground are of primary importance - Greenland and Alaska.

More than 1,500 miles long and 600 miles wide, rising to above 10,000 feet at its highest point and averaging 7,000 feet in elevation, the Greenland Icecap presents a formidable barrier to the air flow in its vicinity and the effects, in terms of the mean circulation and the behavior of individual storms, are clearly apparent. The mountain ranges of southern Alaska rise abruptly from the shore, reaching average heights of about 8,000 feet. Individual peaks attain heights of 15,000-20,000 feet. Like the Greenland Icecap, the Alaskan ranges exert a considerable influence on pressure systems. In addition, they cause a sharp division in climate between interior and coastal Alaska. In summer the Brooks Range of northern Alaska is also of climatic

significance, since it acts as a barrier separating warm air in the Yukon Valley from the cold air which lies over the polar seas. Highest peaks in this range are close to 10,000 feet and the mean level is approximately 5,000 feet.

In Europe and Asia there are a number of mountain systems or highlands which are worthy of mention, though their effects on weather are not as pronounced as in the previously mentioned regions. The Scandinavian Highland runs the length of Norway and western Sweden, with an average elevation of about 4,000 feet and isolated peaks rising as high as 8,000 feet. The Ural Mountains constitute a narrow north-south divide of nearly comparable heights. At their northern end they make an S-shaped bend and link with a ridge which runs the length of Novaya Zemlya.

Highlands of a few thousand feet elevation in central Siberia are overshadowed by a complex of higher ranges - the Verhoyansk, Cherskiy, Anadyr' Ranges, and Kolyma Mountains - further to the east. These ranges average 5,000 feet and above over large areas and have individual peaks which reach 10,000 feet.

## 1.4 Duration of Daylight and Darkness

An important facet of the arctic environment is the extreme variation in the length of day that occurs over the course of a year. As is well known, the regions north of the Arctic Circle experience a period of varying duration in summer when the sun is continuously above the horizon and a period in winter when it is continuously below. At the pole itself the "day" and "night" are approximately of 6 months duration each.

Table 1.2 contains information on the duration of sunlight for the middle of each month at selected latitudes. The figures apply to level, unobstructed ground and may differ for stations located on hills or in valleys.

From the table alone one does not get a correct impression of the amount of useful light at high latitudes. There are two reasons for this. A first is that the arctic twilight is extremely prolonged. For 20 days after the setting of the sun at the pole on September 24, it is possible to read a newspaper by twilight - under clear skies. Complete darkness does not set in until November 12th. The period of total darkness ends on January 30, and the twilight grows brighter until the reappearance of the sun on March 19. Thus, there are only about 80 days of real night at the pole.

Secondly, the absence of sunlight is not as great a handicap as one might think; for in winter the moon is always highest in the sky for the longest duration, at the time of full moon. Furthermore, the reflection of the moonlight on the snow cover adds to the brightness of the landscape. These factors make it possible to travel safely across the snow in the middle of the polar night and even to land airplanes by moonlight under extremely favorable circumstances.

TABLE 1.2
Total Possible Duration of Sunlight on the 15th of Each Month.

| Month | North Latitude | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 60° | | 65° | | 70° | | 75° | | 80° | | 85° | |
| | hrs. | min. | hrs. | min. | hrs. | min. | hrs. | min. | hrs. | min. | hrs. | min. |
| January | 6 | 43 | 5 | 02 | - | -- | - | -- | - | -- | - | -- |
| February | 9 | 12 | 8 | 28 | 7 | 20 | 5 | 10 | - | -- | - | -- |
| March | 11 | 44 | 11 | 40 | 11 | 33 | 11 | 23 | 10 | 50 | 9 | 56 |
| April | 14 | 34 | 15 | 11 | 16 | 09 | 17 | 56 | 24 | 00 | 24 | 00 |
| May | 17 | 08 | 18 | 43 | 22 | 41 | 24 | 00 | 24 | 00 | 24 | 00 |
| June | 18 | 49 | 21 | 53 | 24 | 00 | 24 | 00 | 24 | 00 | 24 | 00 |
| July | 18 | 05 | 20 | 15 | 24 | 00 | 24 | 00 | 24 | 00 | 24 | 00 |
| August | 15 | 41 | 16 | 39 | 18 | 15 | 23 | 19 | 24 | 00 | 24 | 00 |
| September | 12 | 55 | 13 | 07 | 13 | 26 | 13 | 57 | 15 | 10 | 18 | 15 |
| October | 10 | 13 | 9 | 46 | 9 | 06 | 7 | 58 | 5 | 00 | - | -- |
| November | 7 | 34 | 6 | 16 | 3 | 52 | - | -- | - | -- | - | -- |
| December | 5 | 56 | 3 | 42 | - | -- | - | -- | - | -- | - | -- |

# 2. CLIMATIC FEATURES OF THE ARCTIC

Weather conditions in the Arctic are sufficiently variable so that climatic averages cannot in themselves provide the basis for accurate weather prediction. Nevertheless, the forecaster will find a knowledge of climatic conditions useful as a background for prediction, especially when assigned to an unfamiliar area.

The present chapter attempts to summarize some of the more important climatic features of the Arctic. Throughout, the object is to paint a broad, rather than a detailed, picture of the polar climate. First the mean state of the arctic circulation is described for different levels and seasons, along with the related temperature conditions. Various weather elements are then considered in turn, and their regional and seasonal variations discussed. Elements considered are wind speed, cloudiness, precipitation, snow depth, fog, and visibility.

## 2.1 Mean Circulation

In order to depict the major features of the circulation at the surface and aloft, use will be made of sea level isobaric charts and of upper-level height-contour maps. The upper-level contours will, in nearly all instances, afford an accurate representation of the wind flow aloft, but due to frictional and orographic influences some caution must be exercised in inferring surface winds from the sea level pressure patterns.

### 2.1.1 Sea Level Pressure

The mean sea level pressures for the four seasons, as given by Namias [7], are shown in figures 2.1 a-d. In January (fig. 2.1a) the arctic circulation is controlled by four main pressure cells - low pressure areas in the Aleutian and Iceland regions and highs situated over Siberia and northwest Canada. Both lows have eastward extensions which follow regions of open water. The two highs are joined by a thin ridge spanning the Chukchi and Beaufort Seas. The circulation over the pole itself is largely dominated by the cyclonic flow about the Icelandic low.

Over regions of elevated topography the pressure gradients are fictitious and do not accurately depict the wind flow near the surface. Thus, the gradient wind flow over central Greenland cannot be expected to conform to the isobars in figure 2.1a but will be determined by the 700-mb. height-contours of figure 2.2a instead. The strong pressure gradient over southeastern Alaska is also fictitious, lying over the mountain ranges which divide the continental high from the low in the Gulf of Alaska.

April finds the Siberian high much weakened and displaced eastward. On the other hand, the high over North America maintains its strength and shifts closer to the pole. It is at this time of year that the concept of a polar anticyclone is most nearly fulfilled. Important changes also occur in the low pressure cells between winter and spring. The Aleutian low diminishes greatly in intensity, and its extension in the Gulf of Alaska becomes relatively prominent. The Icelandic low also fills considerably and appears to shift northeastward because of greater pressure rises along the coast of Greenland.

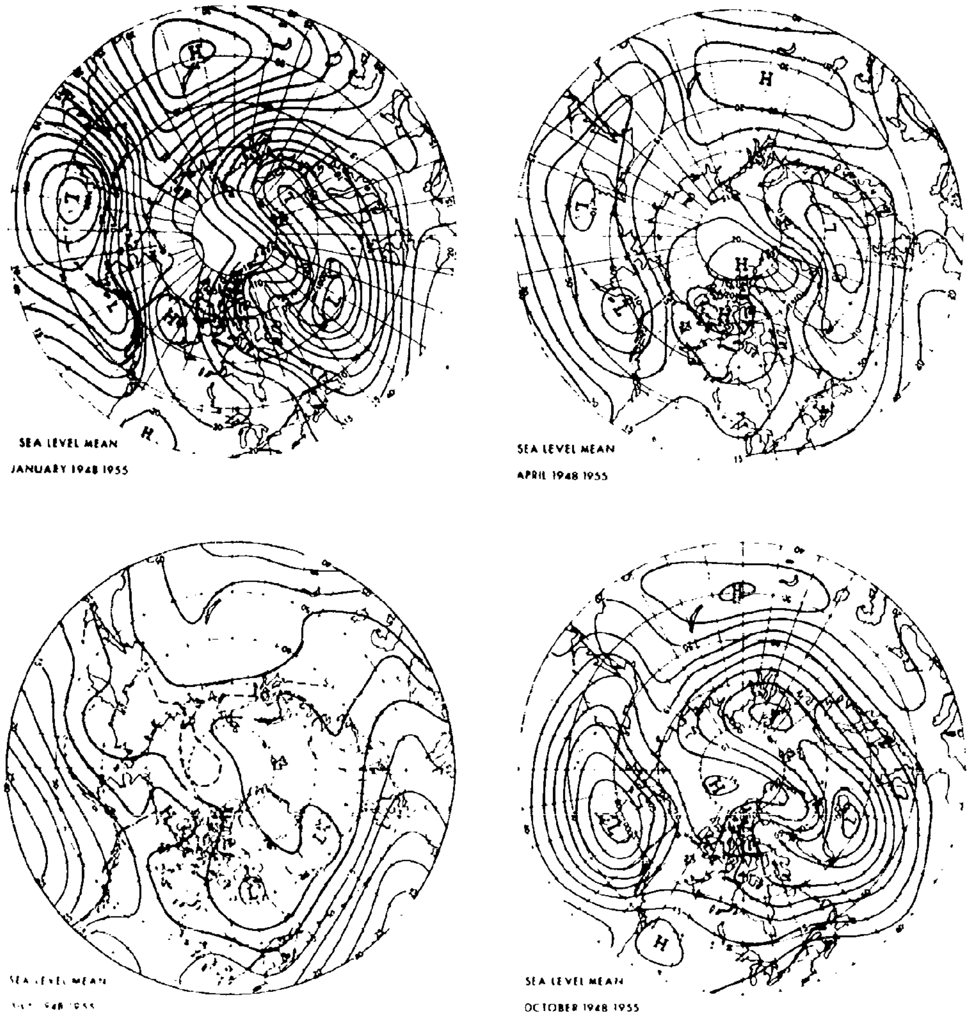
The pressure pattern over the polar cap in July is ill-defined and still subject to some uncertainty. The often assumed semipermanent polar anticyclone is clearly lacking, but as yet it is not certain whether a cyclonic circulation exists in its stead. According to recent evidence a weak trough extends outward from the Asiatic low towards the center of the Arctic Ocean. The presence of a ridge of high pressure over the Beaufort Sea is well confirmed.

In summer the Aleutian low fades away to a mere trough in the isobars, and the Icelandic low remains barely detectable. A more prominent low develops near the southern tip of Baffin Island. The Siberian high disappears completely, and low pressure prevails over the heated Asiatic landmass.

By October the Aleutian low has regained its full intensity but is single-centered and located east of its winter position. The Icelandic low is nearly at full strength and also slightly east of its winter position, the extension in the region of Novaya Zemlya being particularly pronounced. The Siberian high has reappeared and attained moderate intensity by October. The high in the American sector is located over the Arctic Ocean to the north of its wintertime position and is still gaining strength.

### 2.1.2 700-mb. Heights

Figures 2.2 a-d depict the seasonal changes in the upper-level circulation. In January there are three main troughs and low cells in the Arctic and subarctic. Centers of lowest contour height




ISOBARS ARE LABELED IN MB [illegible] COMPARED TO THE ORIGINAL FIGURES. [illegible]

Figure 2.1. "Normal" Distribution of Sea Level Pressure. (After Namias [7].)

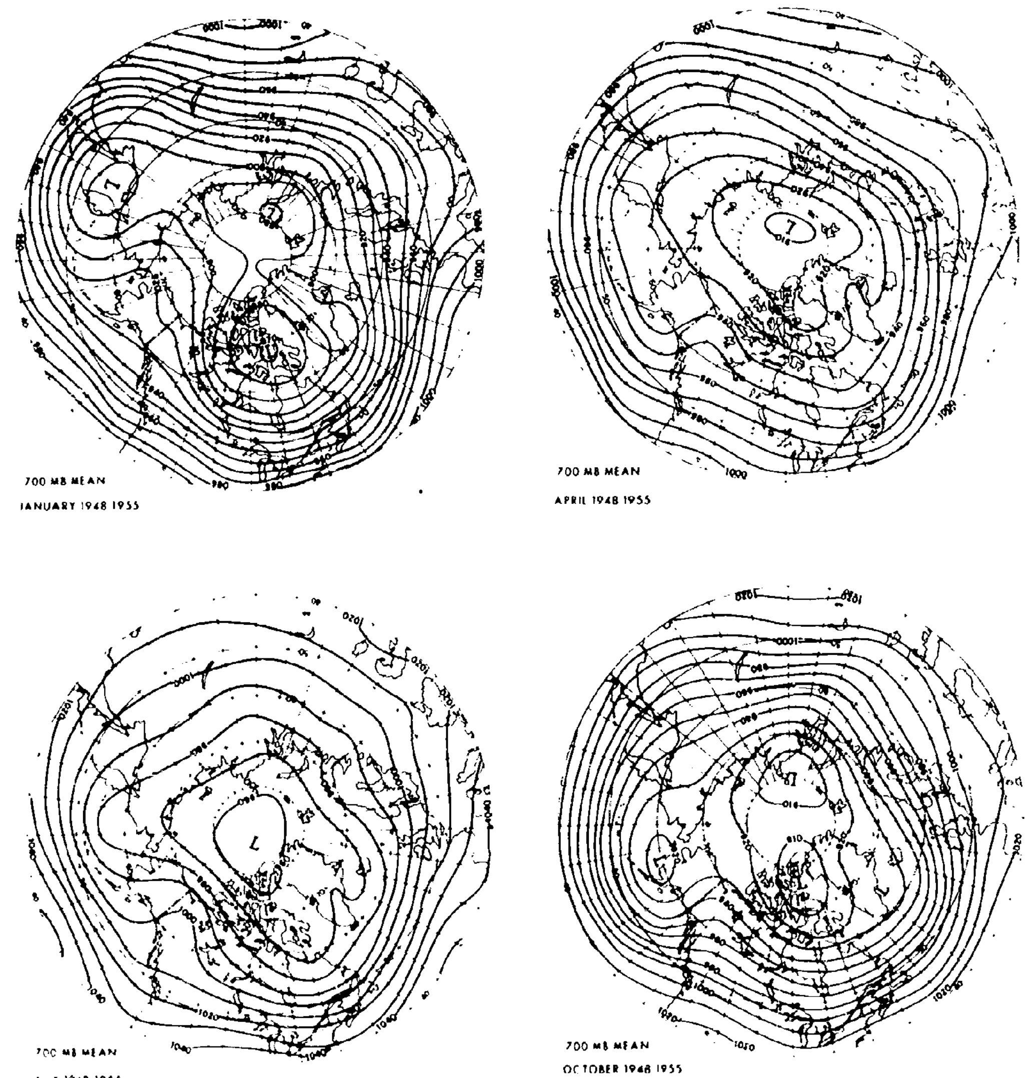



Figure 2.2. "Normal" Distribution of 700-mb. Height. (After Namias [7].)

occur over Baffin Island and near Novaya Zemlya. The most pronounced trough is found with the slightly weaker low near Kamchatka. Each of the upper lows is associated with a main low pressure cell or the extension of a cell at the surface. The most prominent upper-level ridge appears over Alaska. Weaker ridges may be noted between Greenland and Norway and over Siberia.

In April the flow over the polar regions becomes more zonal as the multicellular pattern is replaced by a single low cell near the pole. Troughs still persist in the Baffin Island and Kamchatka regions. The Novaya Zemlya trough is displaced westward to Scandanavia, and, in conjunction with the greater prominence of the surface low in the Gulf of Alaska, a noticeable trough appears in that region at the 700-mb. level. The most prominent ridges aloft are found over Alaska and Greenland.

In July the flow aloft is somewhat weaker but still follows a wave-like path about a single low center near the pole. The Kamchatka trough has shifted eastward to the Bering Sea, a minor trough still appears over the Gulf of Alaska, the trough in the Baffin Island region once again remains fixed, the trough over Scandanavia has retrograded to near Iceland and a new trough has developed over central Asia. As at the surface, the main ridge is found over the Beaufort Sea.

The upper circulation strengthens in October and closed low cells reappear. Main trough and ridge positions are virtually the same as in summer. A noteworthy feature is the continued eastward displacement of the Kamchatka trough towards Alaska even as it tends to be reestablished along the Asiatic coast.

2.1.3 30-mb. Heights

The more prominent features of the 700-mb. circulation patterns are maintained in the upper troposphere, and the winds generally increase in strength up to the level of the tropopause (200 to 350 mb.). In proceeding upward through the lower stratosphere a gradual weakening of the circulation is observed, and troughs and ridges become less pronounced. Higher in the stratosphere wind regimes are encountered which are partially or wholly distinct from the flow patterns at lower levels. The height contours of the 30-mb. surface (23 to 24 km.) in figures 2.3 a-d bring out the major circulation features in the middle stratosphere at the various seasons.

The polar circulation in winter (fig. 2.3a) is dominated by an intense circumpolar vortex. Within this vortex there exists near 70° N. a band of strong winds often referred to as the arctic stratospheric or polar-night jet stream. This jet stream is entirely separate from the familiar polar front jet stream of the middle latitudes. From figure 2.3a it is apparent that wave-like perturbations exist in the polar-night vortex, and from a comparison of this figure with figure 2.2a it is evident that these waves are merely smoothed versions of the largest scale waves present in the troposphere. Thus, the stratospheric flow pattern at this season is not entirely independent of the pattern at lower levels.

In springtime (fig. 2.3b) a feeble remnant of the polar vortex still hovers near the pole, but generally by late May the westerlies disappear completely, and after a brief period of stagnant flow the easterly regime of summer becomes established (fig. 2.3c). The easterlies are essentially undisturbed and, except in the Tropics, cannot be related to the tropospheric flow.

The westerly vortex generally reappears in late August or September and dominates the fall circulation over the polar cap (fig. 2.3d). The stratospheric jet is weakly developed at this season, and the wave-like perturbations which characterize the winter pattern are still in the formative stage.

2.2 Mean Temperature

Various features of the temperature distribution, both at the surface and aloft, are reviewed in this section. As a broad introduction to temperature conditions at the surface, charts of mean temperatures for January and July are first presented. A more comprehensive view of temperature behavior is afforded by graphs of monthly mean maximum, mean minimum, and extreme temperatures for selected stations.

The vertical temperature distribution is represented by means of typical mean soundings and by meridional mean cross sections for various seasons.

2.2.1 Surface Temperature, Winter and Summer

Coldest surface temperatures in winter (fig. 2.4a) occur in the valleys of northern Siberia and over the interior of Greenland. The extreme cold of the Siberian area may be attributed to the clear skies and dry, subsiding air which generally prevail there and to the presence of moun-

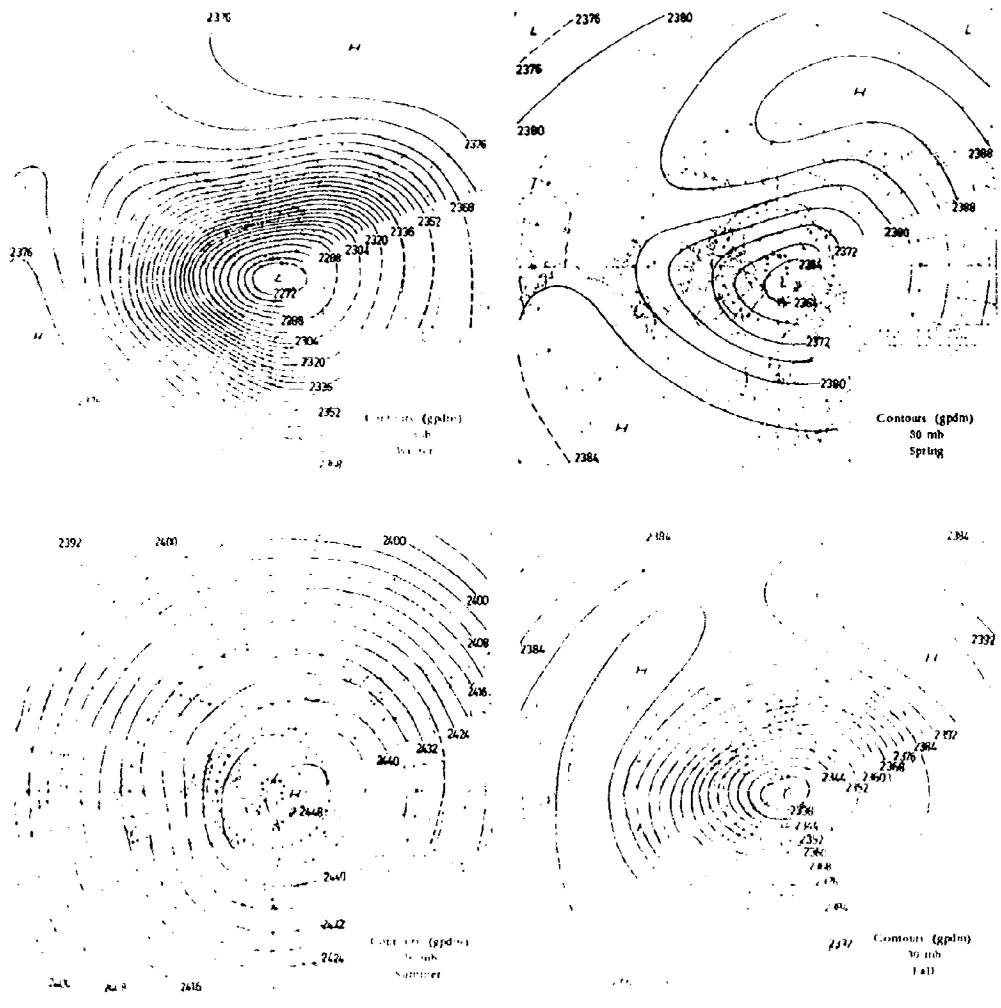



Figure 2.3. "Normal" Distribution of 30-mb. Height. (After Wege, Leese, Groening, and Hoffman [12].)

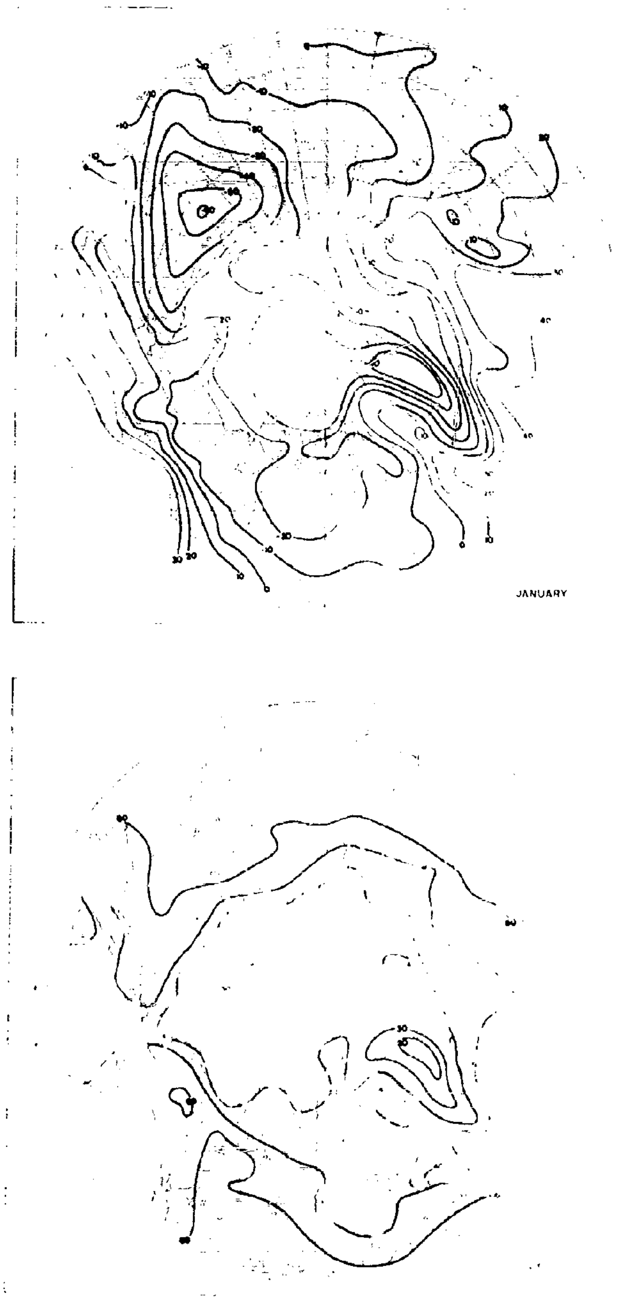



Figure 2.4. Mean Surface Temperature (°F). (After Pettersen, Jacobs, and Haynes [11].)

tains to the south and east which prevent inflow of air from the Pacific at lower levels. The exceptional coldness of interior Greenland is largely an effect of elevation. Because of more ready access to regions of warmer air, less extreme cold is observed over the Arctic Ocean and the North American Continent.

The presence or nearness of open water is clearly reflected in the thermal distribution. A broad tongue of relatively mild temperatures extends from the Atlantic to the Barents Sea. A smaller tongue of mild temperatures stretches northward through Davis Strait to Baffin Bay.

Mean surface temperatures in July (fig. 2.4b) likewise conform to the nature of the underlying surface. Temperatures close to the melting point prevail over the pack ice of the Arctic Ocean and along the fringes of the Greenland Icecap. Over the interior of the icecap temperatures well below freezing are observed. Cool temperatures dip southward over Bering Sea, Hudson Bay, Baffin Bay, and other bodies of open water. Over continental interiors the long hours of sunshine produce warm temperatures even close to the Arctic Circle. Mean temperatures of 60° F. and above occur over central Alaska, northwest Canada, and a vast region of northern Russia and Siberia.

### 2.2.2 Mean Daily Maximum and Minimum Temperatures and Extreme Temperatures

More detailed information concerning temperature behavior in the Arctic is contained in the graphs of mean temperature, mean daily maximum and minimum temperatures, and monthly extreme temperatures appearing in figure 2.5. In winter a great variation in absolute minimum temperature is seen to exist, ranging from near 0° F. at coastal and island stations facing on the Atlantic to as low as -90° F. in river valleys of eastern Siberia. Stations only a short distance removed from the moderating effects of the oceans (Gambell, Anchorage, Upernavik, Kajaani) exhibit absolute minima in the -20° F. to -40° F. range, and temperatures below -40° F. have been recorded over vast portions of the North American and Eurasian Continents and the Arctic Ocean. The coldest temperature on record is a reading of -93.6° F. at Verkhoyansk in Siberia, though it now appears that another Siberian station, Oymyakon, is the persistently coldest spot in the Northern Hemisphere. By way of contrast the coldest temperature ever recorded over the North American Continent was a reading of -81° F. at Snag Airport in the Yukon. Temperatures as low as -84.6° F. have been recorded in central Greenland (Wegener Expedition) and as low as -62° F. over the pack ice (Fram Expedition).

In summer absolute minimum temperatures vary relatively slightly, hovering near the freezing mark at most localities. An exception is central Greenland where temperatures as low as -19° F. have been measured in July.

Absolute maximum temperatures of 90° F. and above are not uncommon in the continental interiors in summer. At Verkhoyansk the temperature has risen to 94° F., making an absolute range of 188° F., the largest on earth. Most arctic coastal and island stations experience temperatures in the 50's, 60's, 70's, and even the 80's in summer. Over the pack ice, on the other hand, absolute maxima are in the upper 30's, and over central Greenland the temperature fails to reach the melting level.

### 2.2.3 Mean Vertical Temperature Distribution

Although the polar regions are often regarded as source regions for arctic air masses, it must be realized that the concept of an air mass as a homogeneous body of air is a considerable oversimplification and that no single sounding can adequately portray the structure of the polar atmosphere at a particular season. In fact, a certain degree of horizontal temperature variation is characteristic of most polar air masses, coldest temperatures near the surface usually occurring close to centers of anticyclones while aloft they are more usually found in regions of low pressure.

Many of the features of the thermal structure are brought out by the mean soundings for January and July in figures 2.6a and 2.6b. Even in the mean a strong inversion exists at all interior stations during the winter. It generally reaches to a height of 1 to 2 kilometers, and the temperature increase between bottom and top may exceed 10° C. At coastal stations (Tromso) a more normal lapse rate exists in the lower levels.

The tropopause in winter lies near the 9 kilometer level. Above the tropopause the temperature becomes more or less isothermal at the more southerly stations. In the extreme north it slowly decreases to a minimum at a height of about 30 kilometers.

In summer the inversion disappears at inland stations but still persists in much weakened form over the pack ice and along the shores of

the polar seas. The tropopause is somewhat higher in summer than in winter, averaging about 10 kilometers. In the first few kilometers above the tropopause the temperature rises sharply. Higher up the lapse rate is essentially isothermal.

The figures well support the earlier statement that considerable horizontal temperature differences occur in the polar atmosphere. This is true both in the troposphere and stratosphere.

2.2.4 Mean Meridional Cross Sections of Temperature and Wind

The average temperature and wind conditions in the Arctic are further elaborated by means of the meridional cross sections in figures 2.7a and 2.7b. Although large in individual cases, tropospheric temperature gradients over the polar region in winter (fig. 2.7a) are small on the average and, in accordance with the thermal wind relationship, the wind changes only slowly with height. Weak easterlies, reaching to heights of less than 5 kilometers (500 mb.), prevail near the surface north of 70 degrees latitude; feeble westerlies lie above. The main feature of interest in the temperature pattern is the inversion which extends to heights of 1 to 2 kilometers throughout most of the polar and subpolar area.

In winter the polar stratosphere contains a unique temperature and wind regime which is separated from the regimes of tropical and temperate regions by a band of relatively warm temperatures, extending from the tropopause to about 25 kilometers and girdling the globe at latitudes between 50° N. and 60° N. North of this warm belt temperatures undergo a steady and pronounced decrease to the pole, the coldest temperatures of all occurring near the pole at the 30 kilometer level (10 mb.). Within the region of pronounced cooling the westerly winds increase rapidly with height and a westerly stratospheric jet stream appears at latitudes of 60° N. to 70° N. The core of this jet is believed to lie at about 60 kilometers height and 55° N. latitude.

Mean temperature and wind gradients remain small in the polar troposphere in summer. The easterlies near the surface shrink in depth to less than 2 kilometers and shift to the latitude belt between 60° N. and 75° N. The low-level inversion diminishes greatly in intensity and extent, occurring only over the Arctic Ocean to heights of about 1 kilometer.

The temperature gradient in summer reverses above the tropopause so that the warmest temperatures are found at the pole itself. In connection with this reversed temperature gradient, the westerlies decrease with height and switch over to easterlies in the layer between 15 and 20 kilometers. The core of the easterlies appears to lie in the Tropics at a height of about 50 kilometers so that the easterly regime at high levels in summer is by no means a peculiarity of the polar region.

The warmth of the polar stratosphere in summer is explained by the maximum of insolation at high latitudes and the absorption of the short wave radiation by atmospheric ozone. With the absence of solar radiation in winter, strong cooling occurs, and the temperature swings to the opposite extreme. Dynamic factors may also be involved in the maintenance of the winter cold pool.

2.3 Surface Winds

The general features of the wind circulation over the Arctic have already been discussed in connection with the charts of mean pressure. In this section we will consider the seasonal and geographical variations in wind speed and the frequency with which gales occur at different localities.

Three main factors influence wind speeds in the Arctic: (1) the pressure gradients associated with cyclones and anticyclones, (2) the static stability of the air in the layer near the ground, and (3) features of the local topography. Average wind speeds tend to be large in areas where cyclonic storms are frequent and relatively intense, and small in regions where anticyclonic conditions predominate. However, the strength of the wind corresponding to a given pressure gradient is much affected by the stability of the air. When an inversion is present very little mixing occurs between the inversion layer and the faster moving layers above. Thus, the momentum lost to the ground by friction is only weakly replenished from above, and the surface wind speed becomes much reduced relative to the gradient wind speed. On the other hand, under unstable conditions the surface wind may attain speeds almost as great as at the gradient wind level.

Many of the aspects of the behavior of wind speed in the Arctic can be explained solely in terms of characteristics of large-scale pressure systems and variations in static stability. Local effects, however, are often of great importance, especially where strong winds are concerned.

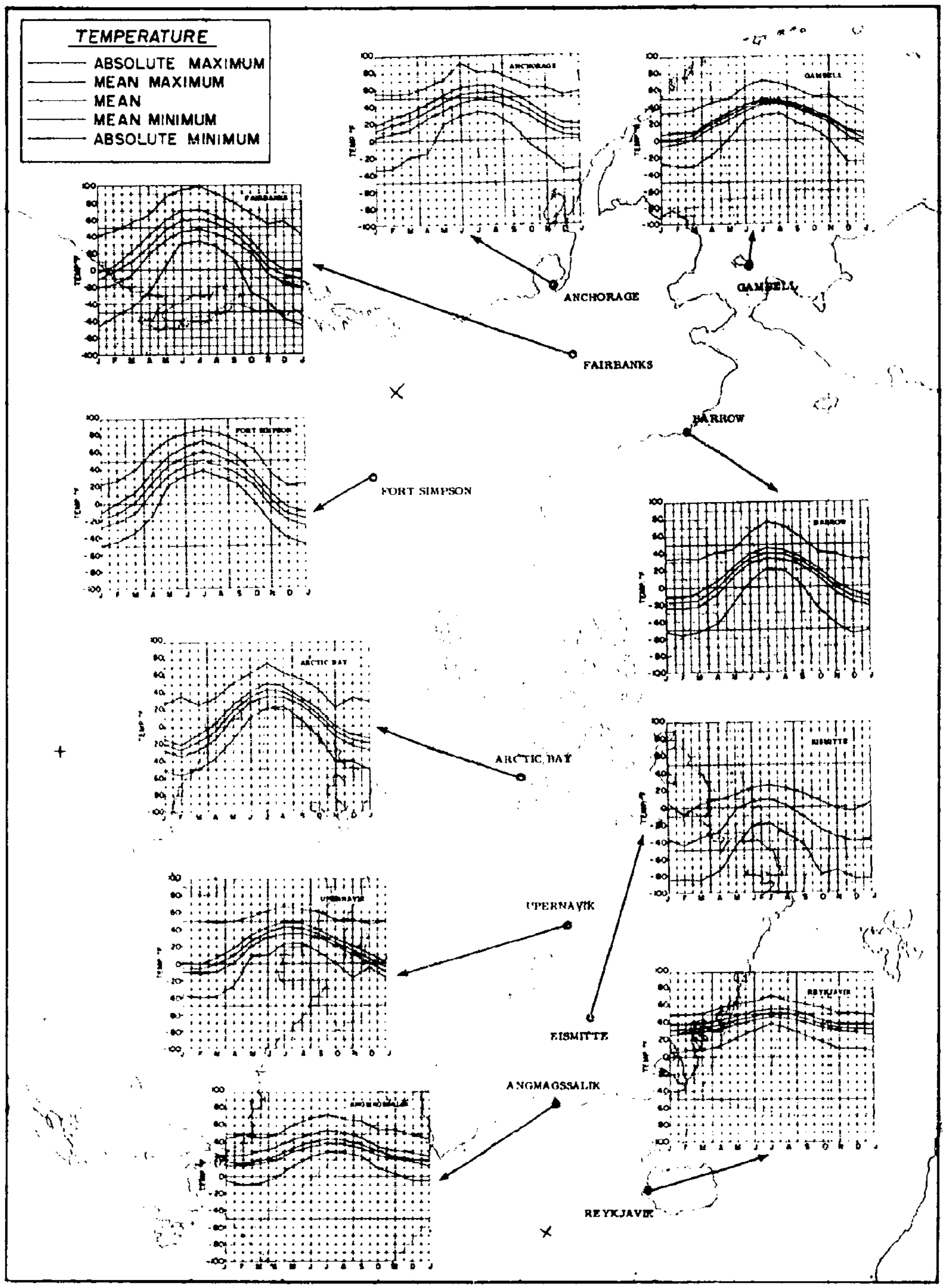



Figure 2.5. Monthly Temperature Distributions at Selected Stations.

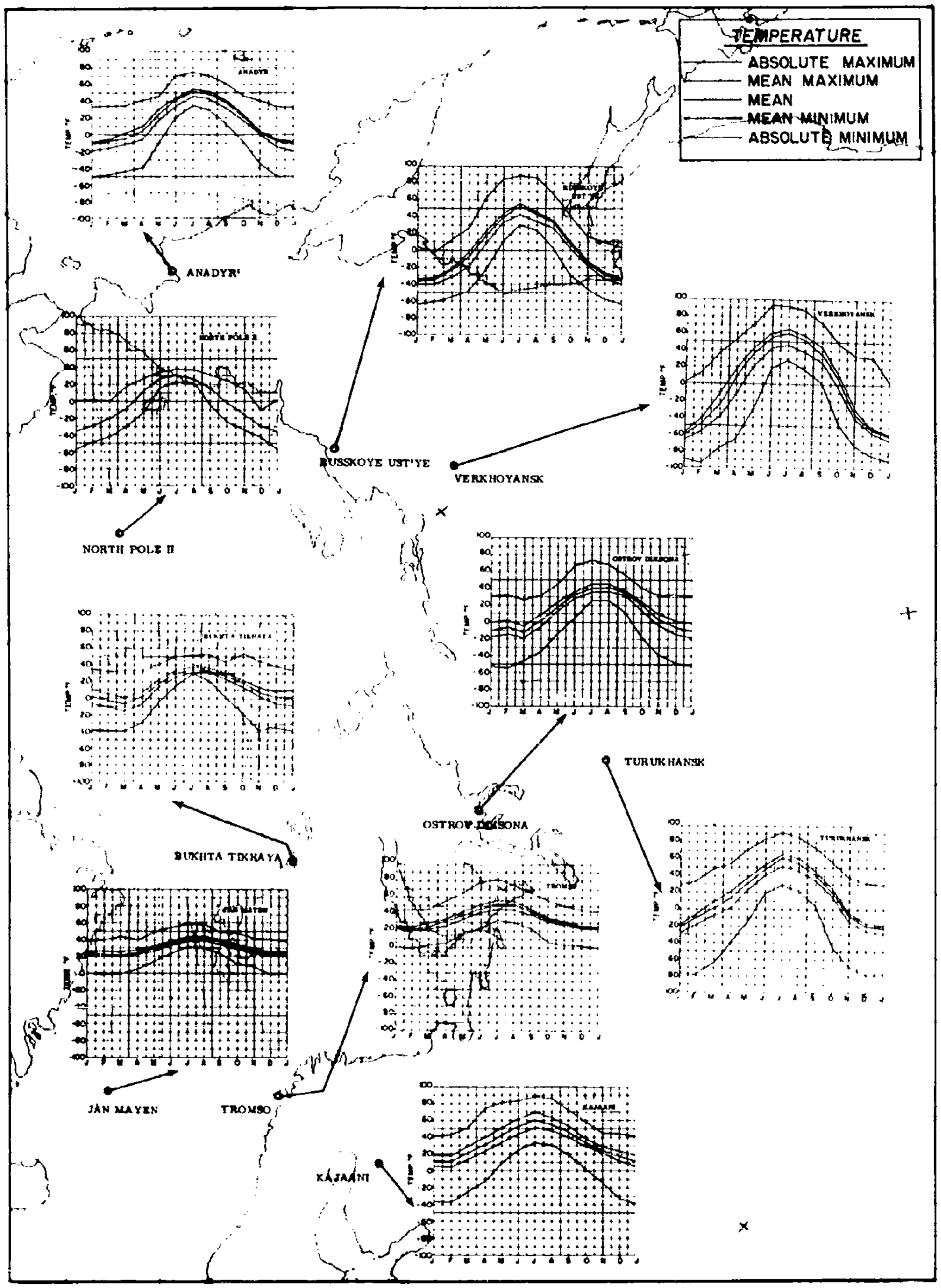
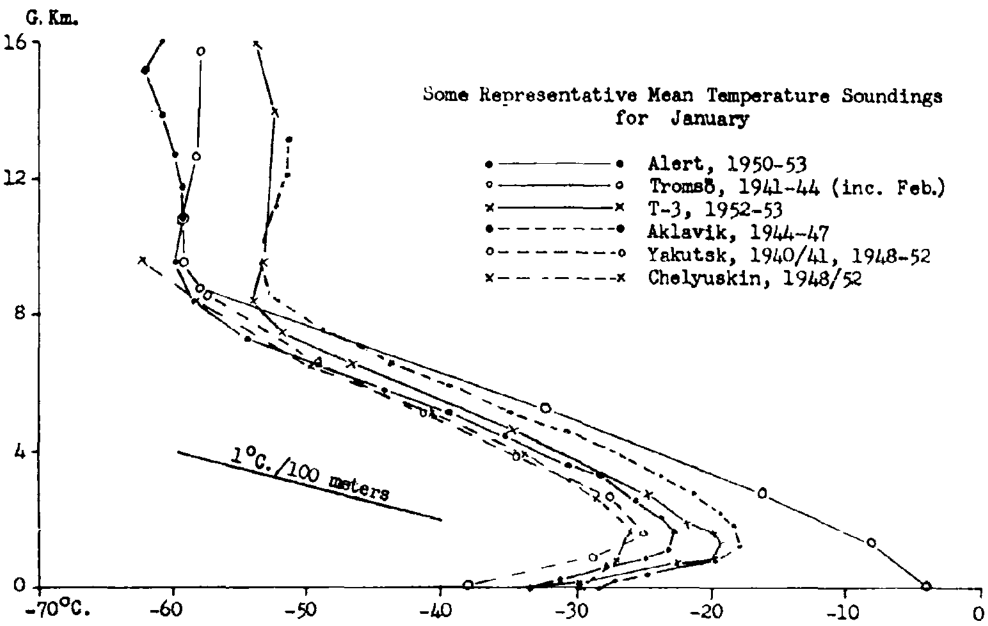



Figure 2.5. Monthly Temperature Distributions at Selected Stations.



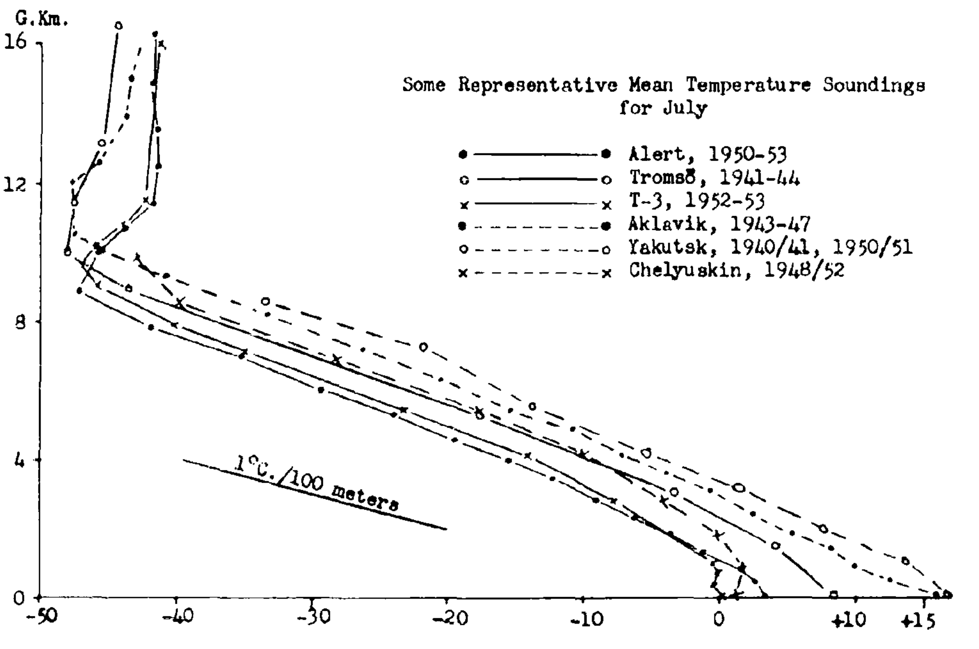




Figure 2.6. Some Representative Mean Temperature Soundings. (After Hare and Orvig [3].)

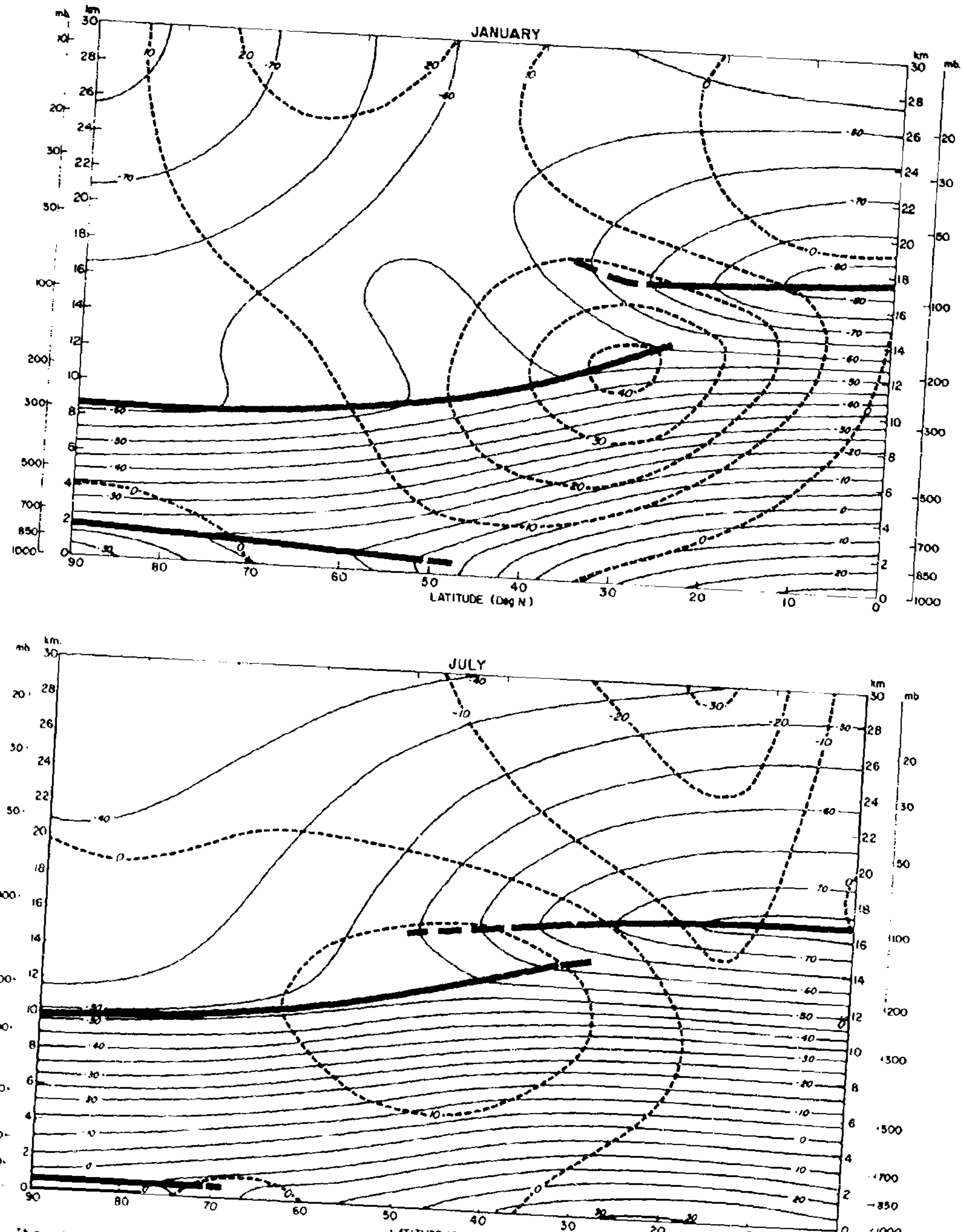



Figure 2.7. Mean Meridional Cross Sections of Wind and Temperature.

Because of these local effects, connected with features of the topography, considerable care must be exercised in selecting representative stations to depict wind behavior in the Arctic, and even with careful selection no small sample can be expected to portray adequately the full range of behavior. With this caution in mind, graphs of average wind speed and frequency of gales are presented in figure 2.8.

It is immediately apparent from the graphs that mean annual wind speeds tend to be greatest at exposed coastal locations in regions of relatively high storminess (Jan Mayen, Vardo, Ostrov Diksona). At these same locations storms are most intense in winter and seasonal variations in stability are relatively minor; consequently, wind speeds are distinctly greater in winter than in summer. The effect of exposure is clearly evident when winds at Angmagssalik and Anchorage are compared with those at the preceding stations. Although both sets of stations are in coastal regions of frequent and sometimes intense cyclonic activity, winds at the latter two stations are generally light because of sheltered locations.

If local effects are excluded, the general statement can be made that in the yearly mean, lightest winds occur in continental interiors (Verkhoyansk, Fairbanks, Watson Lake). This circumstance may be attributed to the effect of the inversion in reducing wind speeds in winter and to the presence of comparatively weak pressure gradients in summer when the inversion is absent. The effect of the inversion in reducing the wind is so pronounced that nearly all inland stations show an annual minimum in wind speed during the winter or early spring.

A number of stations (Barrow, Anchorage, Upernavik) display complex variations in monthly mean wind speeds. In some cases the complexities are due to irregularities caused by the short period of record. In others they are the result of complex variations in pressure gradients and stabilities. For instance, certain stations near the Arctic Ocean experience a semiannual variation in stability with minima in spring and fall. At such stations (for example, Barrow) the wind tends to be strongest at these seasons and to reach minima in winter and summer.

In general, gales are most frequent in regions where mean wind speeds are greatest. However, because of the effects of topography in producing excessively high winds at certain localities, there is by no means an exact correspondence. Vardo and Wales, for instance, have nearly identical mean annual wind speeds, yet gales are more than three times as common at Wales. The effect of local topography in producing excessive winds is further brought out by comparing the annual number of gales at two pairs of neighboring stations in Greenland, Upernavik and Marrak, and Angmagssalik and Atterbury Dome. Threefold to fivefold variations are seen to occur in short distances.

The wind speed over the polar sea averages close to 10 miles per hour and is characterised by an irregular variation which will require many more years of data to clarify. No doubt regional differences in wind behavior will be found to exist so that reports from a single drifting station are not sufficient to define the wind regime of the whole basin. However, it does seem well established that gales are infrequent over all sectors of the Arctic Ocean. Winds of gale force (>32 m.p.h.) are observed only a few times a year, and speeds of 40 miles per hour appear to be close to an upper limit for the pack ice. The dearth of extreme winds is due in part to the high frequency of inversions and in part to the absence of topographic effects. The early conclusion that the lack of strong winds indicates a lack of cyclonic storms is not supported by recent synoptic evidence.

In central Greenland, contrary to other interiors, the wind is greatest in winter and least in summer. In this region the inversion persists throughout the year, and the pressure gradient is the controlling factor in the annual cycle. From the 700-mb. charts presented earlier it is evident that the stronger gradients occur in winter.

## 2.4 Cloudiness

Geographical and seasonal variations in cloudiness in the Arctic depend less on the behavior of frontal systems and cyclonic storms than on the character and changes in character of the underlying surface. The melting of the pack ice in summer is conducive to the formation of a low-level inversion and to the development of persistent fog and stratus. Mean cloud cover in excess of 90 percent may be observed during some summer months and the amount generally exceeds 80 percent. In autumn, following the freezing over of the ice, the cloudiness diminishes rapidly and reaches minimum values of 30 to 40 percent in winter and early spring.

Over land areas the variation in cloudiness is more complex, consisting at many stations of

double maxima and minima. The primary minimum generally occurs in winter. A secondary maximum follows in May or June in connection with the melting of the snow cover. Once the snow is gone, the land warms, lapse rates become less stable and the amount of low cloudiness is diminished. The summer minimum is short-lived, and by early fall the primary maximum is attained at most stations. With the freezing of lakes and other bodies of water in late fall, the cloudiness undergoes a rapid decrease. The secondary minimum in summer may be lacking at coastal and insular stations which are under the influence of flow from the pack ice or open water.

Over and along the fringes of the Norwegian and Barents Seas the seasonal variation in cloudiness is slight, the amount being large at all seasons. Some stations report average yearly values in excess of 80 percent. Because of the prevalence of inversions, stratus is the dominant cloud type in summer. In winter the large cloud amounts are due in part to the great amount of cyclonic activity and in part to the formation of stratocumulus decks when air from the pack ice streams over the relatively warm ocean waters. Since the cloudiness spreads inland a considerable distance, interior stations in the regions adjoining these seas possess relatively large cloud amounts in winter. At these stations the primary minimum comes in early summer and the late spring secondary maximum is much suppressed. Many of the foregoing comments are illustrated by the graphs in figure 2.9.

Records from the Fram, Barrow, and Bukhta Tikhaya exhibit the summer maximum and winter minimum typical of stations over and at the edge of the pack ice. Anchorage, Pond Inlet, Hay River, Turukhansk, and other land stations show the secondary maximum at the time of spring melt. Large annual amounts with little month to month variation are evident at Reykjavik, Jan Mayen, and Angmagssalik. Kajaani furnishes an example of an inland station with relatively large wintertime cloudiness and suppressed secondary maximum in late spring. Somewhat unique is Verkhoyansk, an interior station with single pronounced summer maximum and winter minimum.

## 2.5 Precipitation

Because of the blowing and drifting of snow, accurate measurement of precipitation is difficult to achieve in the Arctic and data regarding amounts must be accepted with some reservations. However, the uncertainties are not sufficiently great to obscure the broad features of the precipitation distribution.

In general, annual precipitation amounts increase southward from the pole. Over and along the fringe of the Arctic Ocean annual amounts (including both rain and the water equivalent of snow) average less than 5 inches and rise to about 15 inches at more southerly locations in Eurasia and North America. Much higher amounts are measured at coastal spots adjoining the Atlantic Ocean and the open waters of adjacent seas; amounts of 40 to 60 inches or even higher being common in Norway, Iceland, and southern Greenland.

Over most of the Arctic the maximum precipitation occurs in late summer and the minimum in winter or spring. Such an annual cycle clearly reflects the effect of air temperature in determining the capacity of the air to hold water vapor. The only regions which depart significantly from this regime are the aforementioned areas bordering the open seas. In these areas temperatures are higher and there is a plentiful moisture supply so that the degree of cyclonic activity becomes a more important factor. Since the storminess is at a peak in fall and winter, maximum precipitation is obtained at these seasons. Minimum amounts (but not generally below those of other portions of the Arctic) are measured in summer.

The annual cycle at selected stations is illustrated in figure 2.10.

### 2.5.1 Snow Depth

Because of their effect on energy and moisture transfer processes at the surface, the amount, extent, and duration of snow on the ground are important elements in the arctic climate. Some idea of the monthly variation in snow depth may be gained from the graphs in figure 2.11. It must be noted that snow depths undergo extreme variations in short distances as a result of drifting and that only broad conclusions can be drawn from the records of selected stations. For instance, Narssaq on the southern tip of Greenland reports an average snow depth of 4 inches in March, while Prins Christians Sund less than 200 miles away reports a depth of 60 inches. These differences due to exposure must be kept in mind in interpreting data on snow depth.

Over the Arctic Ocean snow begins to accumulate on the ice in late August or September and reaches a depth of about 1 foot by late spring. The snow generally begins to melt in June and disappears almost entirely in July.

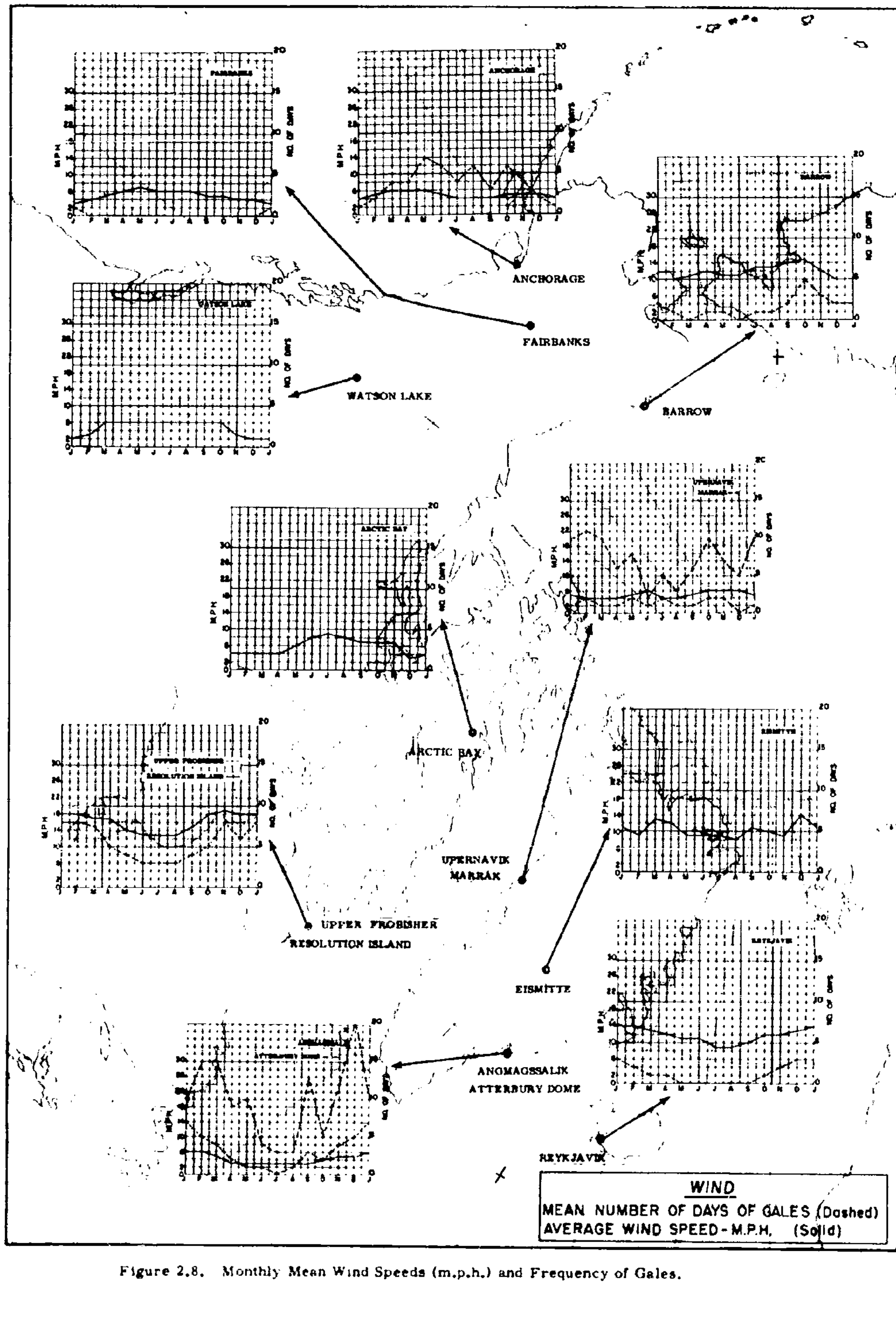



Figure 2.8. Monthly Mean Wind Speeds (m.p.h.) and Frequency of Gales.

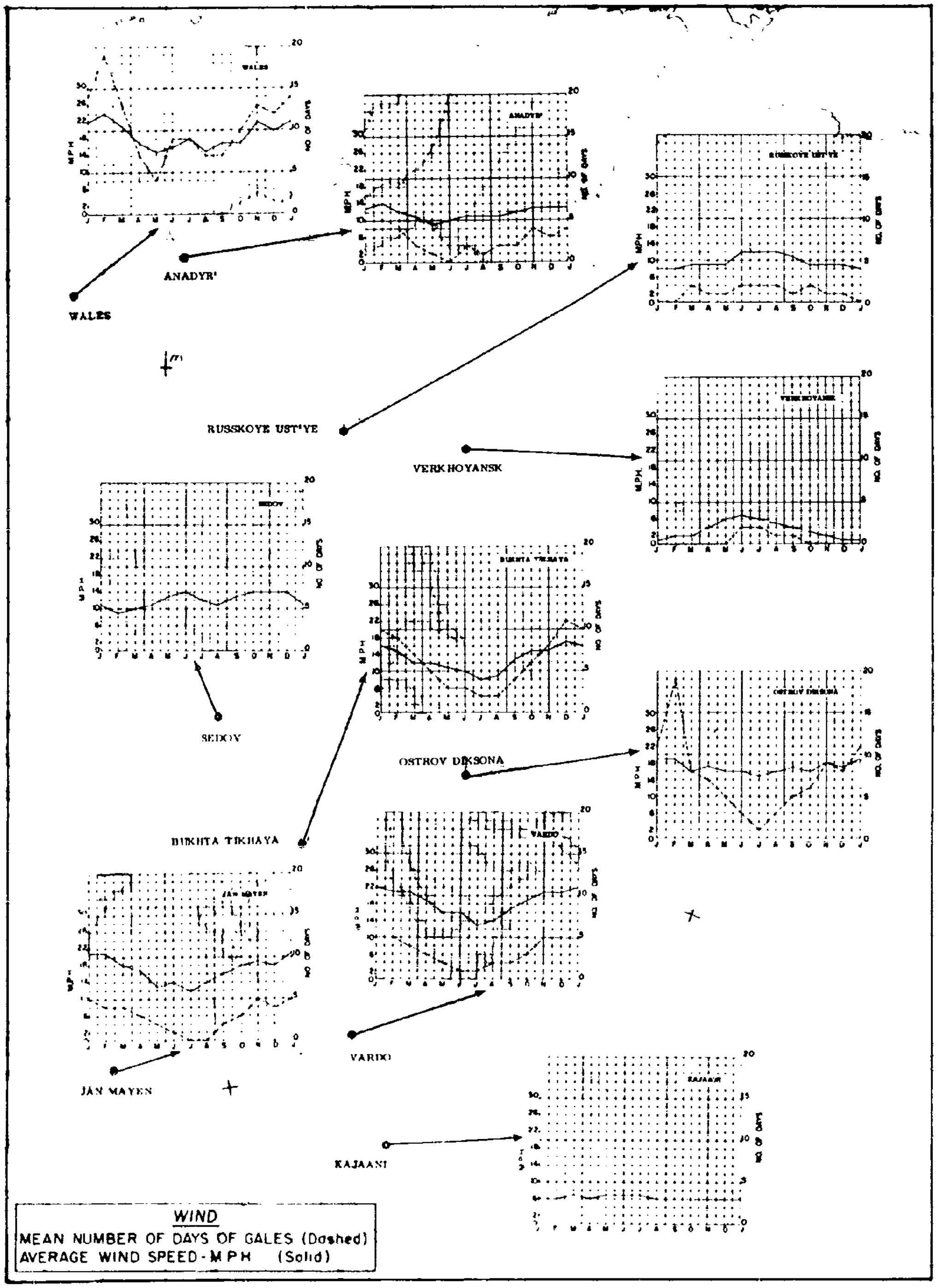



Figure 2.8. Monthly Mean Wind Speeds (m.p.h.) and Frequency of Gales.

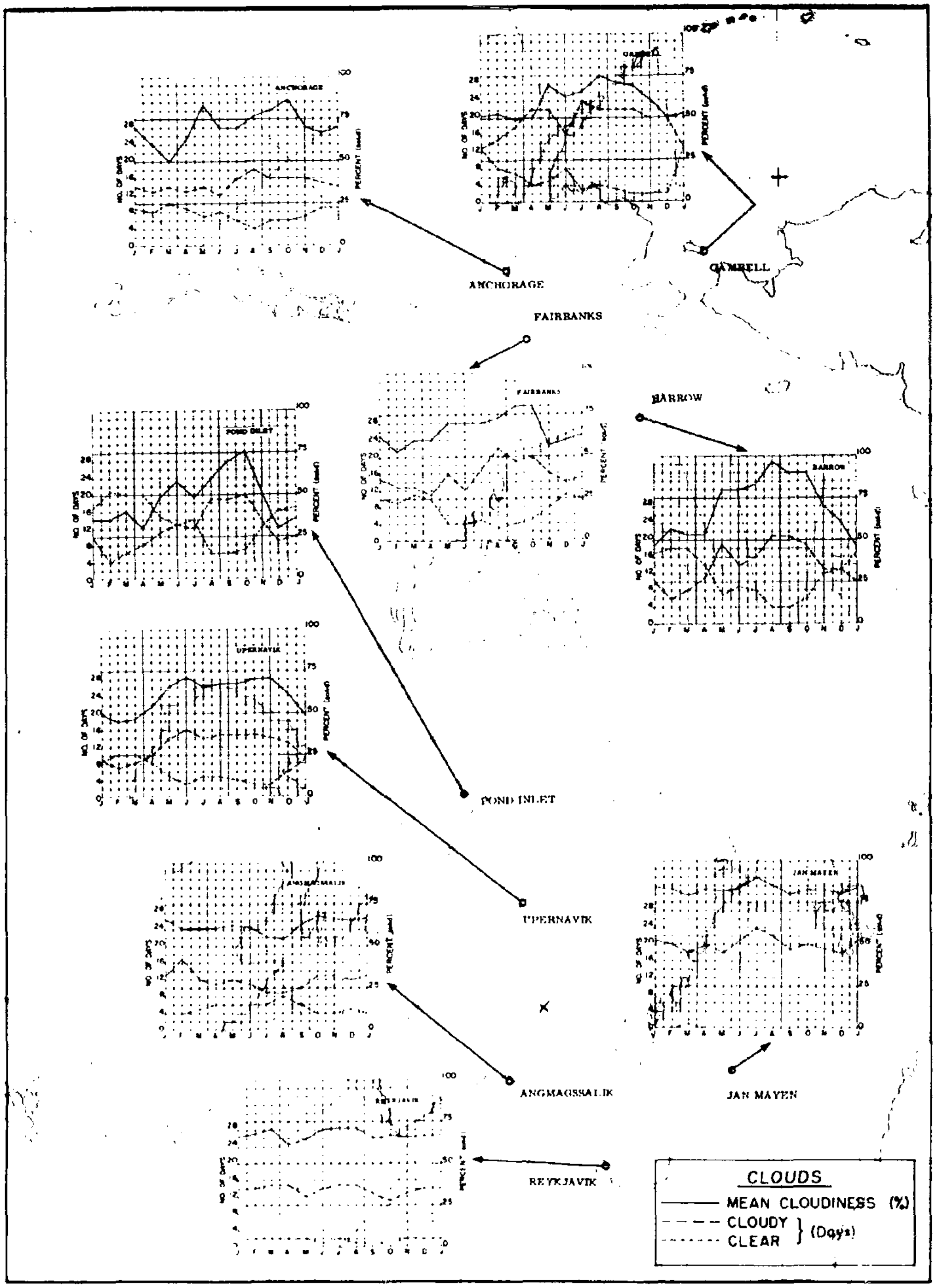



Figure 2.9. Monthly Mean Cloud Cover (%).

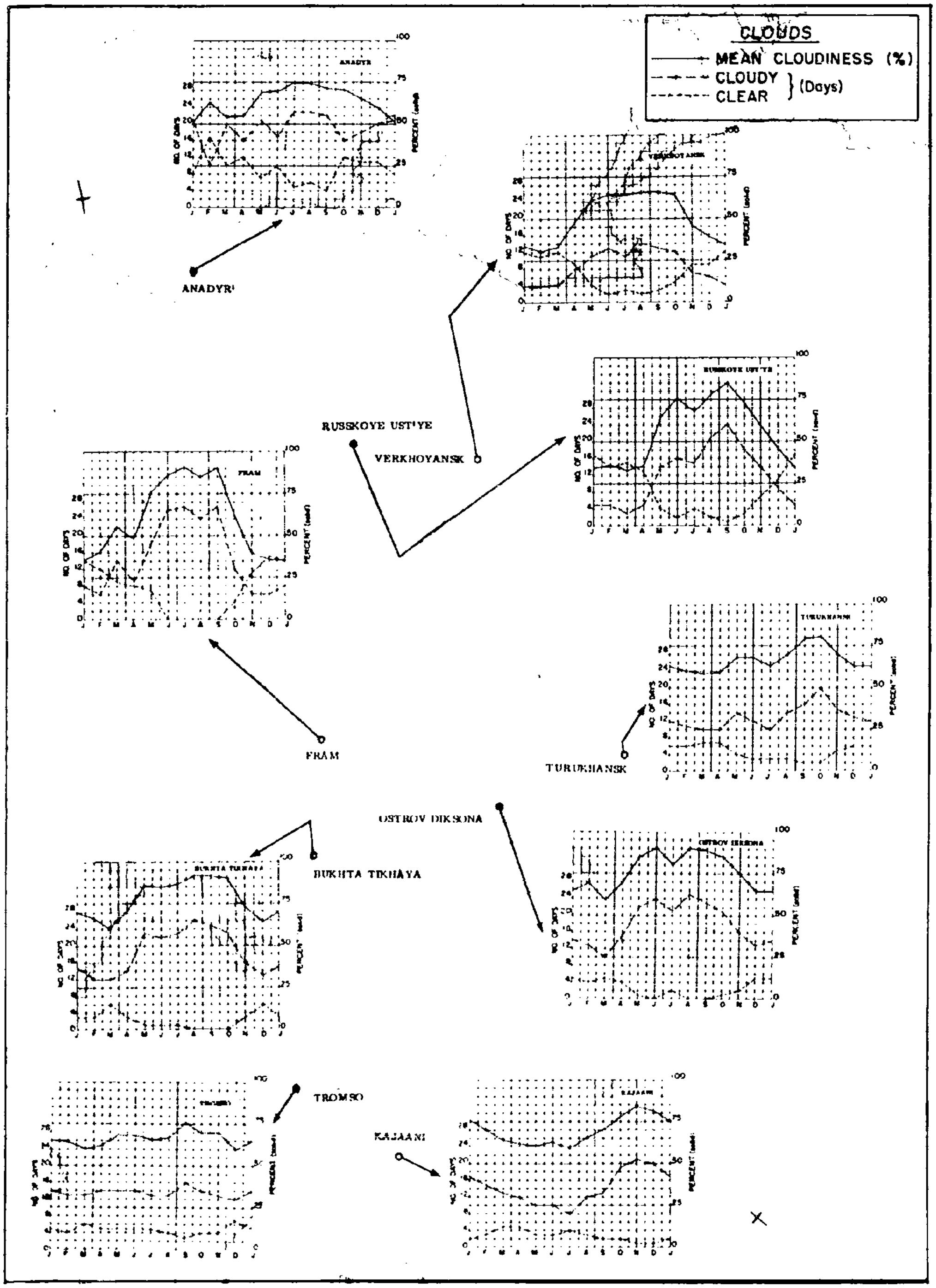



Figure 2.9. Monthly Mean Cloud Cover (%).

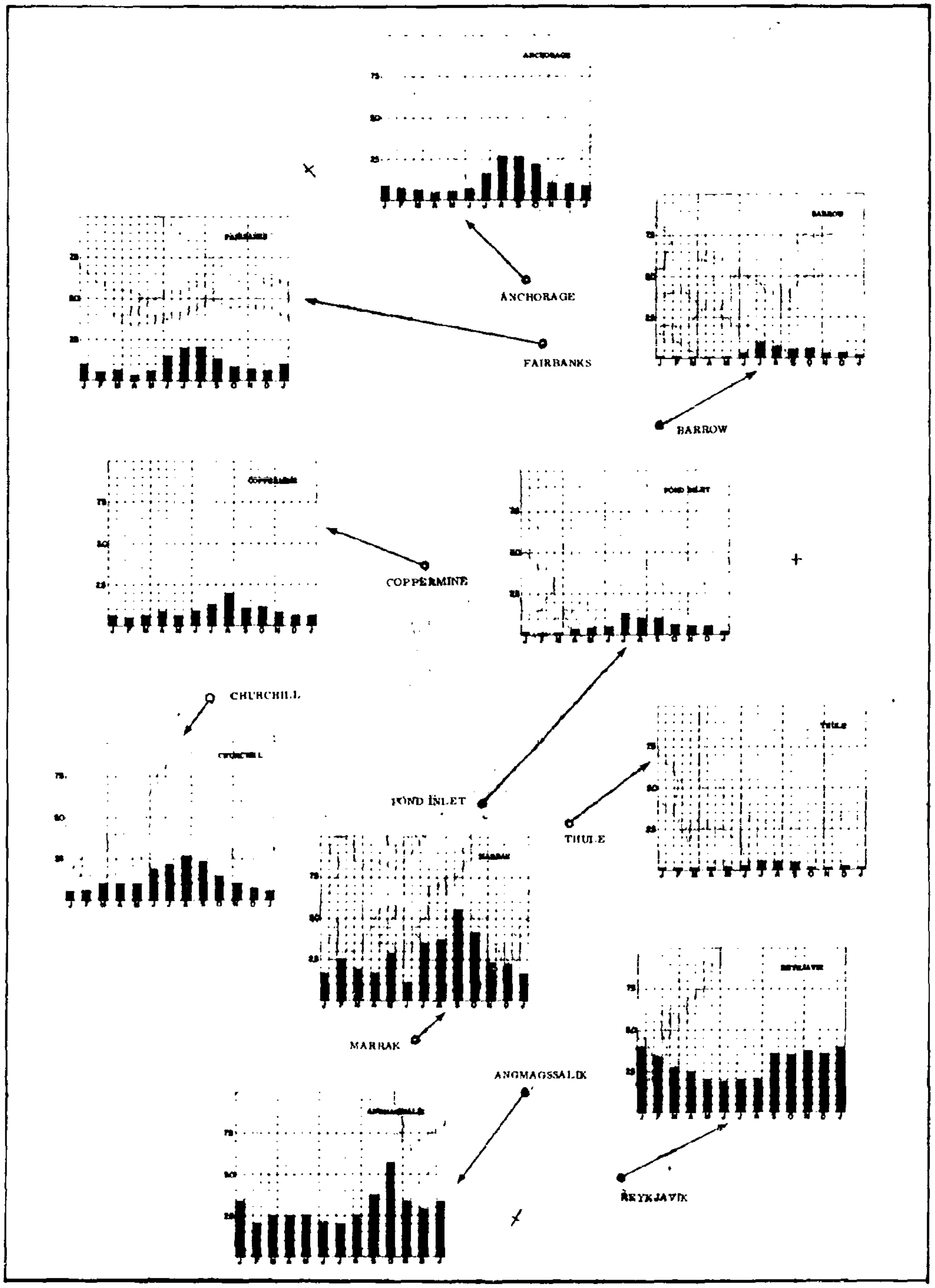



Figure 2.10. Monthly Mean Precipitation (Water Equivalent in Inches).

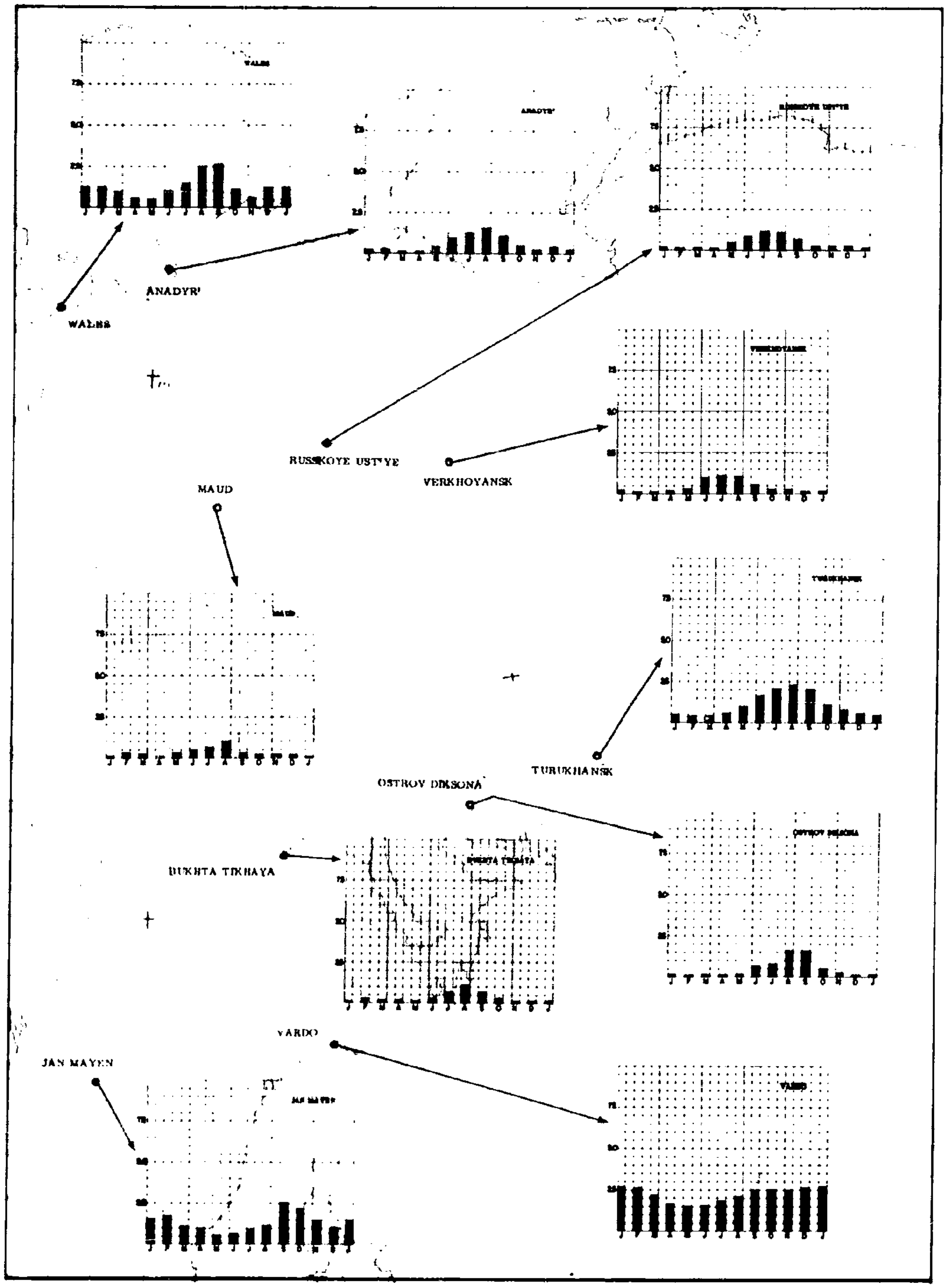



Figure 2.10. Monthly Mean Precipitation (Water Equivalent in Inches).

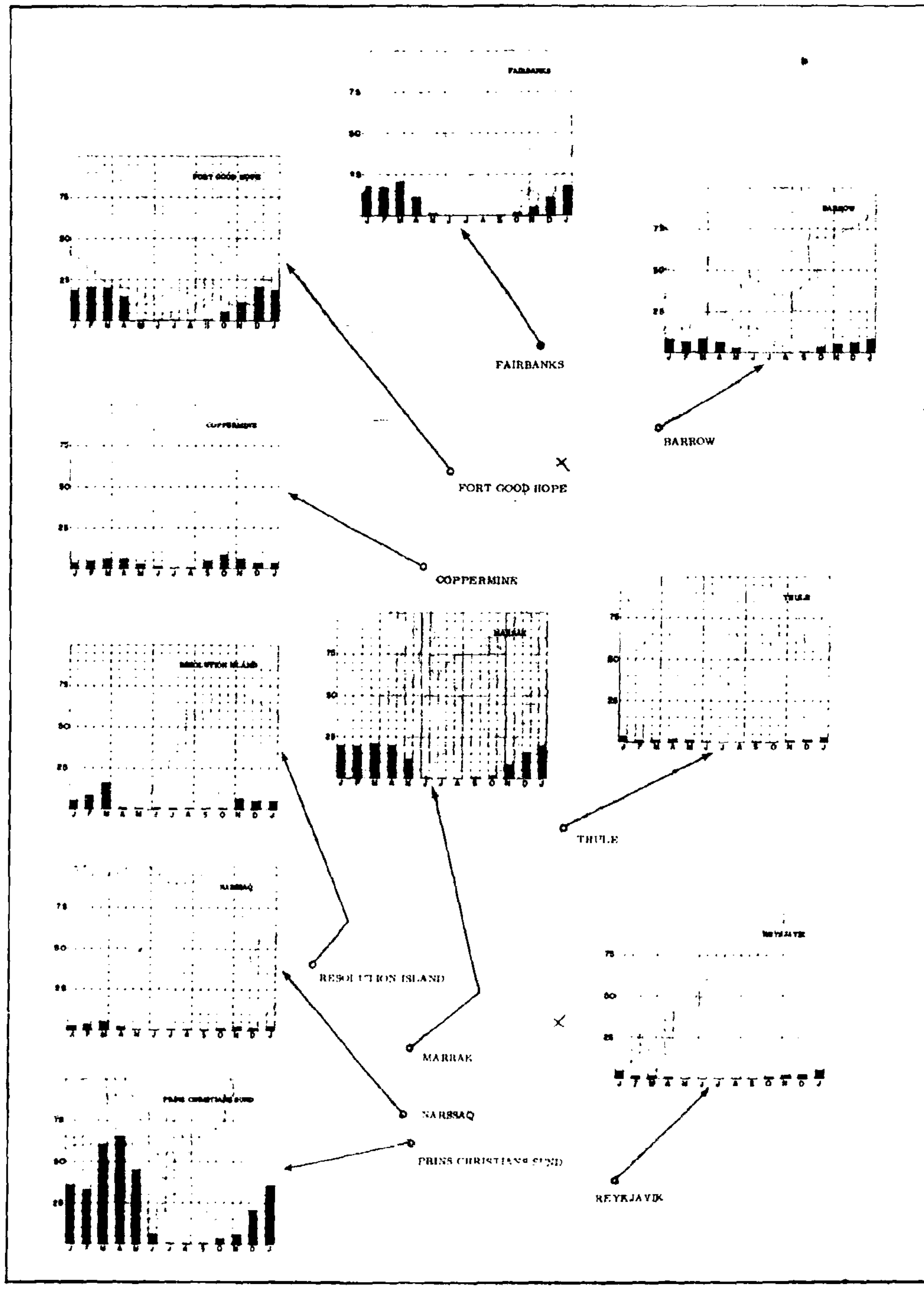



Figure 2.11. Monthly Average Snow Depth (Inches).

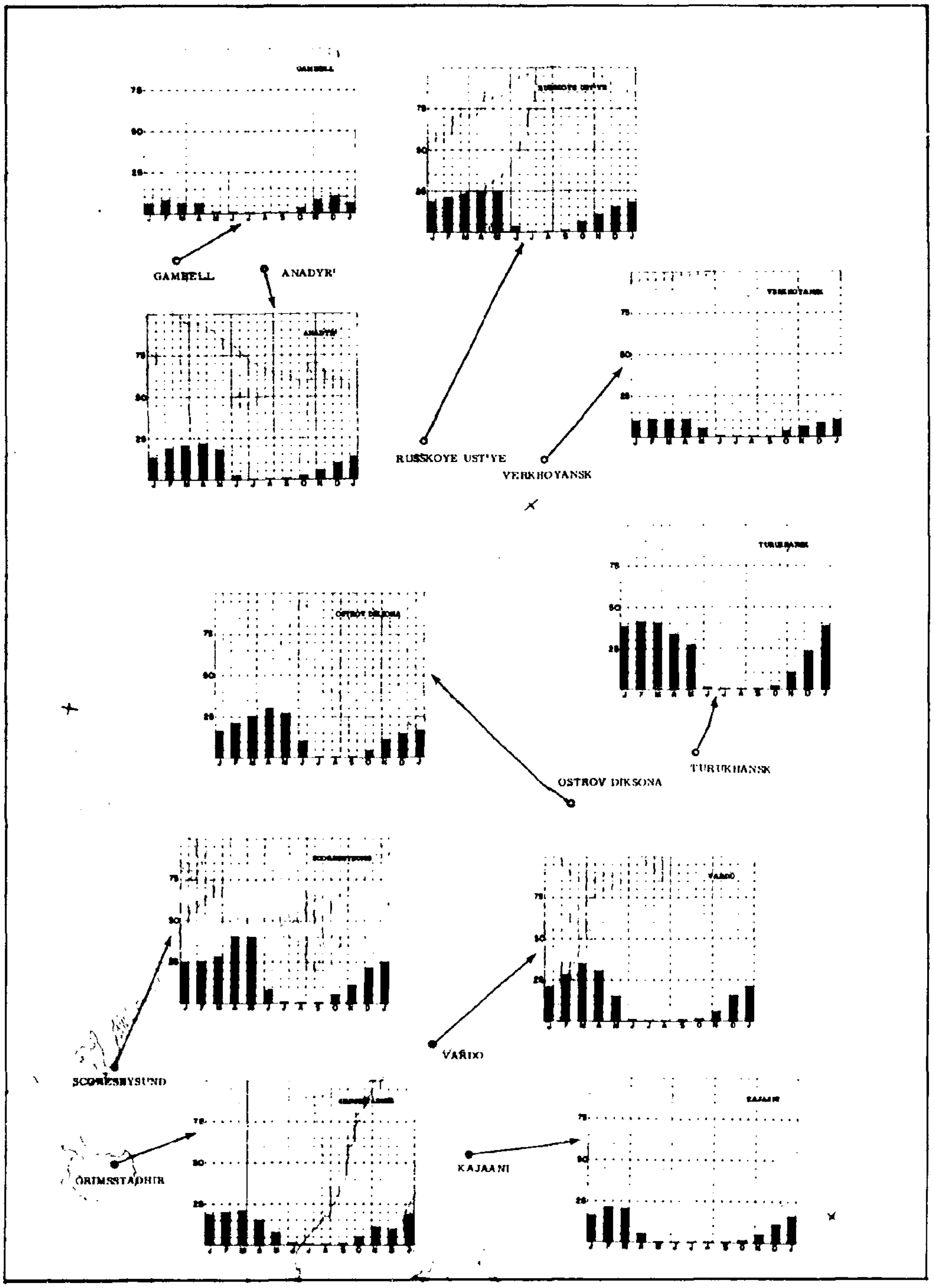



Figure 2.11. Monthly Average Snow Depth (Inches).

An especially long snow season also exists over northern Siberia where the ground is first covered in late September or October and does not become bare again until June. At most stations in this area the snow achieves maximum depths of 1 to 2 feet in March or April, and substantial depths persist throughout May. Scattered stations in the Queen Elizabeth Islands and along the coast of Greenland experience similarly prolonged snow seasons. Usually these stations are characterized by heavy accumulations with depths as great as 5 feet reported at some places.

Over interior portions of the Arctic snow cover generally persists from October to May and reaches a depth of about 2 feet in late winter and early spring. At the outer edges of the Arctic the snow usually disappears by the end of April.

Coastal stations in Iceland receive only slight accumulations. Inland stations may have depths of almost 2 feet (Grímsstadhir) with only a short period of bare ground during the summer. Small accumulations are also observed along the south coast of Norway. However, amounts gradually increase to the north and at Vardo, for instance, a depth of nearly 3 feet is attained in March.

From the brief data considered it would appear that the snow depths at arctic stations depend on temperature, precipitation, and (most importantly) on the local vagaries of the wind. Until more representative snow measurements are available, many of the foregoing remarks must be regarded as tentative.

## 2.6 Fog and Visibility

Fog occurs with greatest frequency over the Arctic Ocean and along the adjoining coasts, many stations in these areas report this phenomenon on more than 100 days per year. Fog occurs much less frequently at most inland localities.

At all oceanic and coastal stations the fog is most frequent in summer and least in winter. The summer fog is of the advection type, being caused by the chilling of relatively warm, moist air in its passage over the melting ice surface or cold waters. It is extremely patchy, and a daily occurrence may often represent, at most, a few hours of low visibility.

The fog regimes at inland stations are quite different than at coastal stations and are of two different types. Some stations exhibit the greatest frequency of fog in the fall or early winter. The fogs at these stations belong mainly in the radiation category and are connected with the cooling of the ground and the development of pronounced inversions. At other stations the maximum is reached in mid-winter at the time of coldest temperatures. The fogs in these cases are ice fogs caused by the addition of moisture to the air by human activities.

The graphs in figure 2.12 bring out some of the more salient features of the fog distribution. Many stations (the Maud, Jan Mayen, Barrow, Bukta Tikhaya, and Ostrov Diksona) furnish examples of large annual frequencies and summer maxima. The lesser annual frequencies and fall maxima at inland stations are illustrated by the records for Fort Good Hope, Kajaani, and Turukhansk. Fairbanks with its extensive military and aircraft operations is an outstanding example of a station with a sharp peak in fog occurrence in winter due to the formation of ice fog.

Verkhoyansk shows a double maximum probably as a result of radiation fog in the early fall and ice fog in the winter. Other stations (not shown) also display more complex distributions. Stations slightly removed from the coast are subject to both the advection fogs of summer and radiation fogs of fall and winter. At such stations double maxima or flat distribution may occur.

Finally it should be remarked that the few stations in figure 2.12 cannot adequately portray the intricacies of the fog distribution. Fog, like wind speed and snow depth, is much affected by the local topography and large variations in behavior may occur in short distances.

### 2.6.1 Other Visibility Reducing Factors

The polar atmosphere is relatively uncontaminated by smoke, dust, and other impurities so that falling snow, blowing snow, and ice crystal haze constitute the only additional visibility reducing factors of importance. Of these only blowing snow is of major importance.

Information on the frequency of blowing snow is scanty. Coastal stations of Russia and Siberia report 100 or more days of blowing snow per year. During the winter months snow may blow on more than half the days.

Statistics on blowing snow at Point Barrow on the Alaskan coast are similar to those at

Siberian coastal stations. During the winter of 1957-1958, a normal winter as far as wind speeds were concerned, blowing snow was observed on 73 days in the 6 month period from November through April and on 23 percent of the hourly observations.

Since the frequency of blowing snow depends mainly on the frequency with which the wind exceeds a certain critical speed - generally about 10 knots in the Arctic - a sharp decrease in the number of occurrences is to be expected at interior stations where winds are characteristically light in winter. As the wind rises above the critical value the visibility lowers and usually becomes less than 1 mile when the wind attains a speed of 25 knots. Thus, the graphs on wind speed (fig. 2.8) give at least a crude idea of the locations at which blowing snow may be a frequent hazard in aircraft operations.

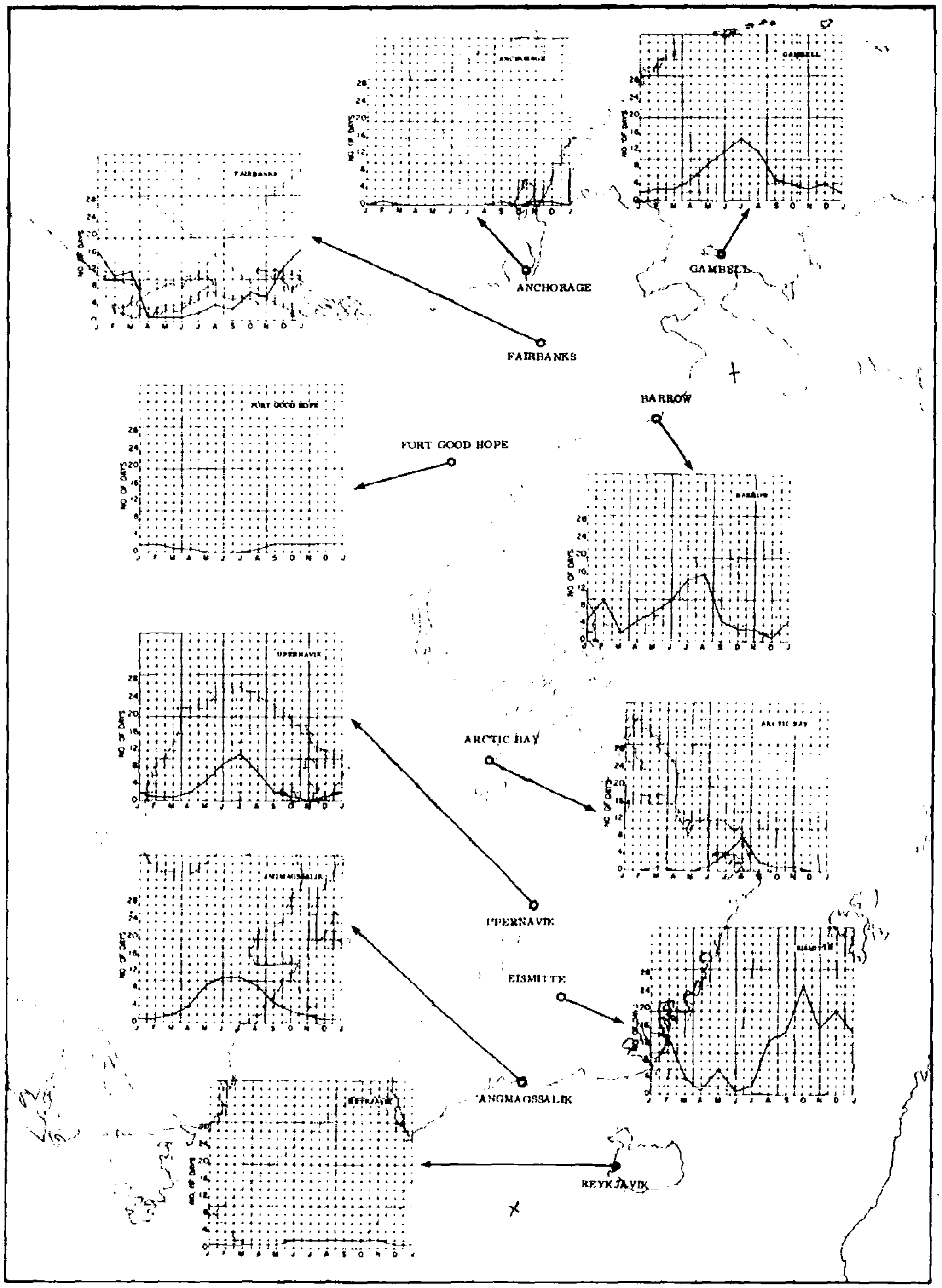



Figure 2.12. Mean Number of Days with Fog.

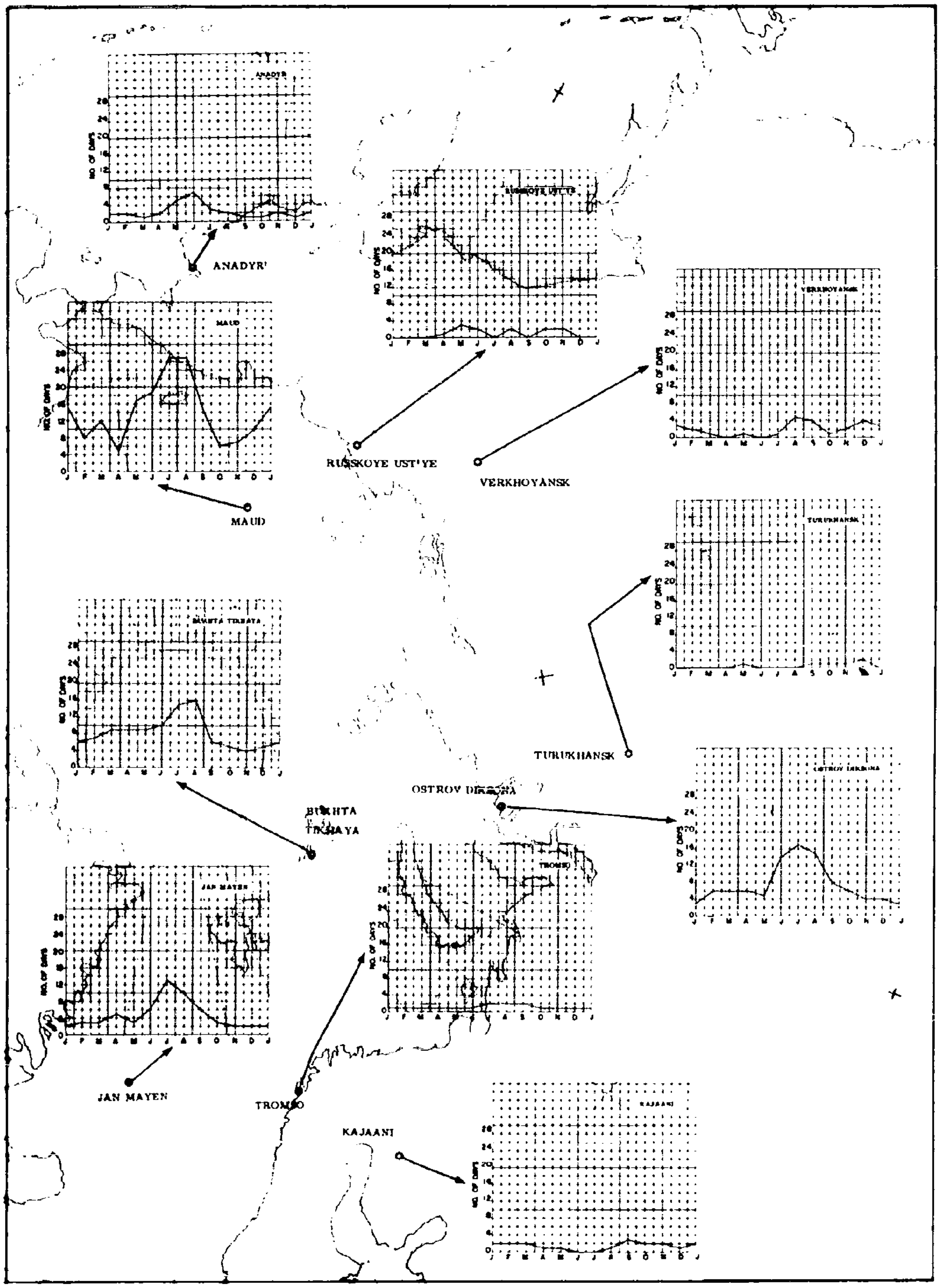



Figure 2.12. Mean Number of Days with Fog.

# 3. THE STRUCTURE AND BEHAVIOR OF ARCTIC WEATHER SYSTEMS

Until fairly recently it was believed that the circulation at low levels in the Arctic is characterized by a more or less permanent anticyclone. This view is now known to be incorrect. Weather data gathered from drifting vessels, ice stations, air reconnaissance, and an expanded network of land stations make it abundantly clear that the Arctic is a region of moderate synoptic activity. Migratory high and low pressure areas, frontal systems, upper waves, and jet streams--the familiar synoptic entities of middle latitudes--appear regularly, if not frequently, north of the Arctic Circle. These, and other features of the arctic circulation, will be discussed and illustrated in this chapter.

## 3.1 Baroclinic Disturbances

This term refers to large-scale cyclonic disturbances which owe their origin primarily to strong thermal contrasts and which exhibit a considerable degree of thermal asymmetry. We will understand it to include both wave cyclones and others in which the thermal contrasts are not sufficiently sharp to permit easy identification of fronts.

In most areas and seasons, temperature differences are greatest at middle latitudes so that the main regions of cyclonic activity lie outside the Arctic. Cyclones over the polar cap are mostly remnants of storm systems which have entered the Arctic during the final or occluded stage of the life cycle; and once imbedded in the cold air they gradually fill and disappear at the surface, while aloft they may persist for long periods as thermally symmetric cold lows.

Despite the preponderance of cold low structures in polar regions, there are also numerous instances of baroclinic developments. Outstanding examples of these occur along the northern coasts of Siberia, Alaska, and Canada during the summer season. Ultimately they are connected with the strong thermal contrasts that develop along the continental shores due to the different responses of land and ice surfaces to the solar radiation. The land absorbs a large fraction of the incident radiation and in this way warms appreciably. The ice reflects a greater proportion of the insolation and, furthermore, cannot warm above the melting temperature of 32° F.

Although the land-ice boundary is ultimately responsible for the summer storms, it should not be inferred that a more or less permanent frontal zone parallels the arctic coast in summer and that cyclone development is confined to a coastal strip. The differing properties of the land and ice surfaces set up and maintain the broad thermal contrasts that later sharpen into fronts. The locations of the fronts in any given case are determined as much by the state of the general circulation as by the locations of heat sources and sinks.

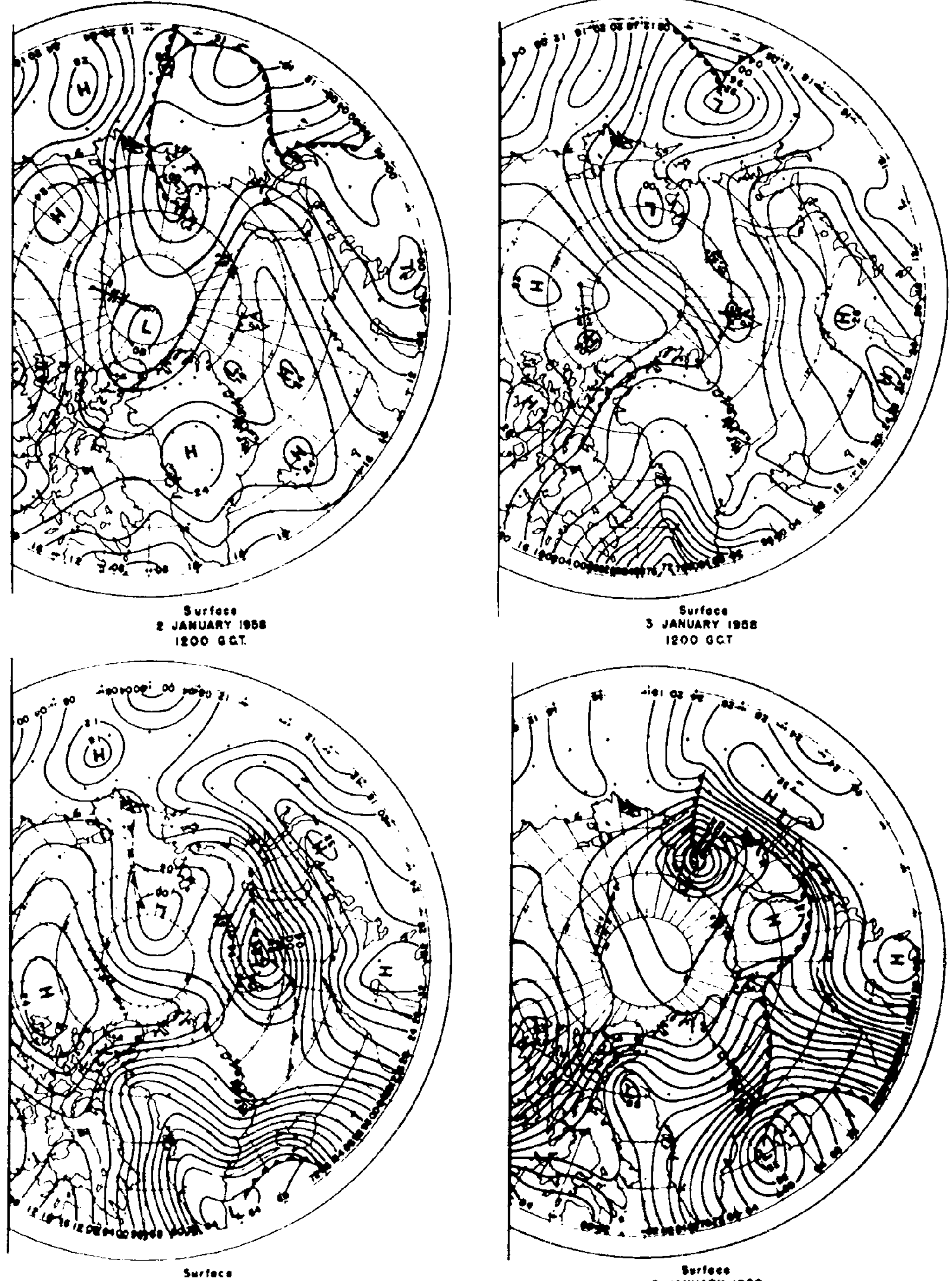
A second region of important baroclinic development is found off the east coast of Greenland during the cold season. Developments are often complex in this region, being more in the nature of rejuvenations of old systems from the Atlantic and North America than entirely new formations. Vorticity generation in the lee of the Greenland ice barrier, and heating of arctic air along the edge of the ice pack contribute to the developments in this area.

An example of the less commonly observed spontaneous type of development is shown in figure 3.1. A weak low not connected in any obvious way with earlier features of the flow appeared off Greenland on the map for 2 January 1958. During the next 2 days it deepened somewhat and moved to the vicinity of Spitzbergen. At that stage a rapid intensification occurred, accompanied by a frontogenesis, and on the map for 5 January a deep conventional-looking wave cyclone was present near the Taymyr Peninsula. Further minor wave activity then took place on the newly formed frontal system.

In addition to these favored regions of baroclinic development, there are many other areas where thermally asymmetric lows appear sporadically. The large-scale flow pattern is of vital importance in determining these less regular occurrences. When the polar circulation is extremely anomalous, as sometimes happens for periods of several weeks, it is possible for a normally anticyclonic region to become the seat of vigorous cyclonic activity.

## 3.2 Frontal Systems

Opinions regarding arctic fronts, among forecasters with experience in polar analysis, range through a broad spectrum. At one extreme we find the skeptics who question the existence of true fronts in the Arctic; at the other end of the spectrum there are those who maintain that a more or less permanent "arctic front" oscillates back and forth along the outer edge of the



Surface
2 JANUARY 1958
1200 GCT

Surface
3 JANUARY 1958
1200 GCT

Surface
4 JANUARY 1958
1200 GCT

Surface
5 JANUARY 1958
1200 GCT

Figure 3.1. Example of Spontaneous Development of a Low - January 2, 1958 through January 5, 1958.

Arctic. The truth of the matter lies somewhere between these extremes.

There can be no doubt that the phenomenon which the middle latitude analyst marks on his map as a front exists also in polar regions. There are several reasons, however, why fronts are not as prominent in the Arctic as in more southerly latitudes, and these probably account for some of the skepticism regarding their existence. First, except during the summer months, fronts are not as frequent in the Polar Basin as in many other parts of the globe. And such fronts as do appear during the cold season are usually well occluded and in the process of decay.

Another factor which tends to make surface fronts indistinct is the cold air "film" which blankets much of the surface at all seasons. This film may mask rather substantial temperature fluctuations in the free atmosphere. Over the pack ice in summer, for instance, surface temperatures hover close to 32° F. in all sectors of a storm, due to the contact of the air with the melting ice surface. However, the cooling spreads upward relatively slowly so that frontal structure may be preserved aloft. Although the fronts in such cases are actually upper fronts, they are sufficiently close to the surface to be entered in a conventional manner on the surface chart.

In rejecting the opposite view--that there is a more or less permanent "arctic front" present at northerly latitudes--we enter on the delicate subject of frontal philosophy which it is not our intention to air at length here. We will just note that all meteorologists agree that low-level fronts must continually fracture or dissolve in order to allow for the necessary exchange of heat between high and low latitudes. Thus, any given frontal system has only a limited life span, and birth and decay (frontogenesis and frontolosis) are ever present. Because of its brief existence, we cannot speak of a certain front as being the "polar front" or the "arctic front" unless it is a member of a family of fronts which can be clearly identified on the basis of latitude, temperature, or some other objective property.

Even a cursory examination of historical weather maps reveals that the number of separate frontal systems analyzed from Equator to pole can vary within wide limits. On some days fronts may be lacking over the entire length of North America; at other times as many as five or six frontal systems may be present. And the temperatures associated with the fronts can run the gamut. Under these circumstances, it is an oversimplification of the facts to say that the atmosphere is composed of three air masses--tropical, polar, and arctic--separated by the polar and arctic fronts. The analyst should not force the atmosphere into a prescribed frontal mold, but should decide each case on its own merits. In so doing he should follow the rule that the simplest frontal analysis is usually best. It is not good practice to enter a front to account for every small variation in temperature, moisture, wind, or weather.

In view of the foregoing remarks the use of the terms "arctic front" and "polar front" will be avoided here. However, as stated previously, fronts of various description do appear regularly in the Arctic. Moreover, there are noticeable regional and seasonal differences in their frequencies, as may be seen by reference to figures 3.2 and 3.3.

During the cold season there are two areas where fronts are drawn quite frequently in or near the Arctic. A first area covers the Greenland, Norwegian, and Barents Seas, and adjoining portions of the Arctic Ocean. The fronts in this area are usually occluded, although other frontal structures are analyzed as well. Quite often a front is drawn in a trough that extends from near Iceland to Novaya Zemlya. This trough reflects the major storm track that passes through the region and is pronounced even on the mean monthly charts. It is the practice of entering fronts in this trough that has given rise to the term "Atlantic arctic front." Since the wisdom of drawing a front in this trough varies from case to case, it should not be regarded as a permanent frontal zone.

A second area of high frontal frequency in winter runs in an arc from the Alaska Range in southern Alaska to the Canadian Rockies. The front in this area is obviously a regional phenomenon, connected with the presence of a mountain barrier and the very different properties of the surfaces on opposite sides of the barrier. In many instances this front may more correctly be regarded as the boundary of a shallow cold layer dammed up against the mountains. The surface front in such cases is due to the slope of the terrain and not to the slope of the cold air. Consequently, it is questionable whether it should be regarded as a true front. At other times, however, the boundary slopes and extends to about the 700-mb. level, justifying the drawing of a front.

Sharp fronts in this area are invariably associated with the approach and passage inland

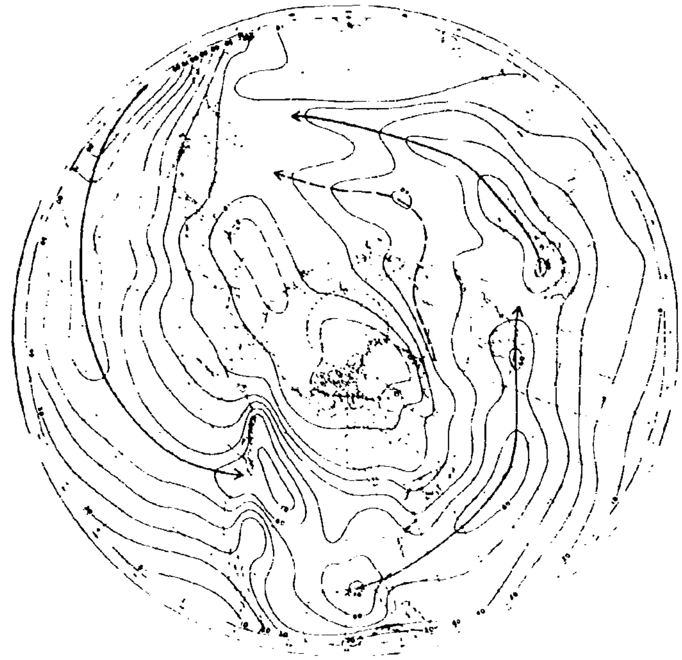
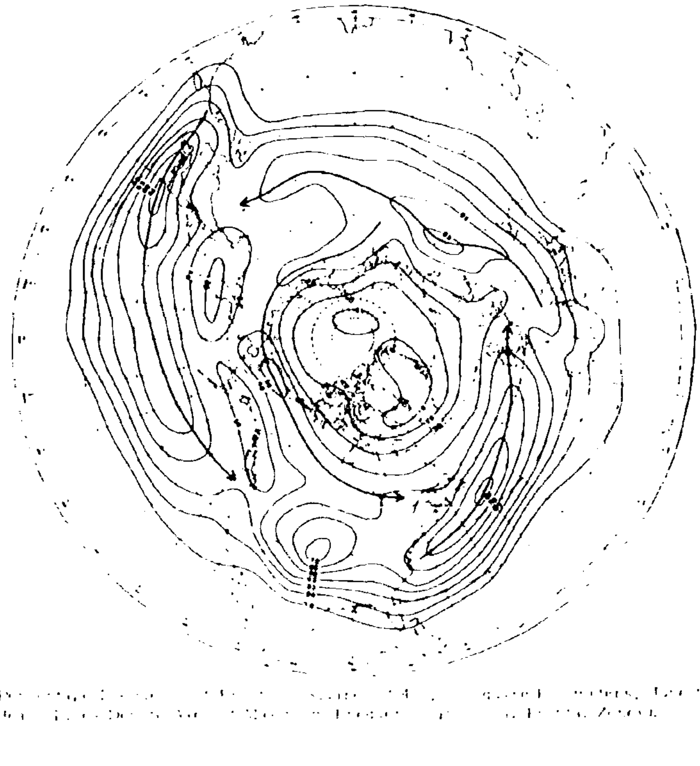


Figure 3.2. Percentage Frequency of Fronts in Squares of 400,000 Square Kilometers, December through February. Heavy Lines Denote Axes of Maximum Frequency (Principal Frontal Zones).



[illegible] August. [illegible] Zones.

of occluded cyclones from the Pacific. When a series of cyclones enters the coast, the first produces a low-level frontogenesis over and to the lee of the mountain range, while succeeding lows induce waves on the resulting front. The occluded front from the Pacific lies roughly normal to the so-called "arctic front", passing through the wave crest. South of the crest the occluded front extends to the surface; to the north it is located aloft over the wedge of arctic air. Since nearly all, if not all, waves which form along the "arctic front" owe their origin to disturbances moving inland from the Pacific, a proper understanding of the weather attending a particular wave requires that attention be given to both frontal systems.

Figures 3.4 and 3.5 give an example of the type of front under consideration and of the interaction of a Pacific disturbance with the front. A cyclone, which earlier had been instrumental in forming the sharp frontal zone, is located over Hudson Bay on the map for 13 February 1957. The front stretches from this cyclone westward and northwestward to a storm entering the south coast of Alaska. During the next 24 hours the Alaskan low moves inland along the preexisting front deforming it into a wave shape. The frontal system accompanying the storm sweeps eastward south of the low center. In the particular case shown here the front has been analyzed as a cold front. Ordinarily, it is drawn as an occlusion, and not uncommonly some sort of upper frontal structure is depicted to the north of the wave.

A fast-moving upper trough follows the cold front. The region of temperature concentration and high wind speeds at 500 mb. lies to the south of the "arctic front", indicating the shallow nature of this front and the importance of the upper wave in determining the associated weather. The temperature and wind cross sections, shown in figures 3.6 and 3.7, respectively, further bring out the distinctions between the two frontal systems. The front from the Pacific is diffuse at low levels, but is moderately strong above 850 mb. The temperature contrast is particularly large within the zone marked by the heavy lines in figure 3.6. In the upper troposphere, the frontal zone broadens, and its boundaries become indistinct. The jet core at 300 mb. is clearly associated with this more southerly frontal zone.

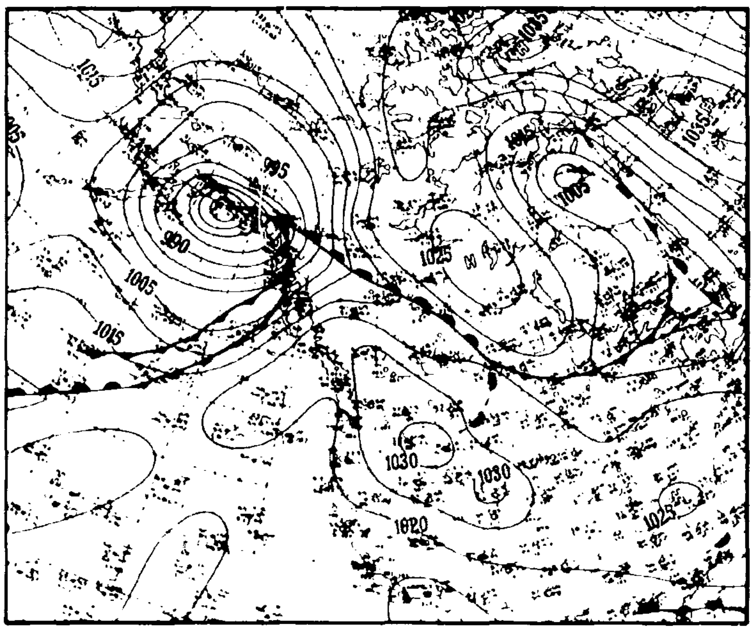



Figure 3.4. Surface Chart for 1230 G.C.T., 13 February 1957. (Northern Hemisphere Historical Series.)

The structure of the "arctic front" is very different. The temperature gradient is huge at and near the surface--35° F. in less than 50 miles--while already at 850 mb. it begins to weaken, and at 700 mb. disappears altogether. A relatively weak offshoot of the main westerly jet spreads down over the frontal surface. The winds rapidly shift to east in passing downward through the frontal zone into the cold wedge below.

In the foregoing illustration, the two frontal and wind systems are sharply differentiated. Not all cases are so clear cut, though the dual components are nearly always present. It has been helpful in describing the example to make use of the currently popular term "arctic front." However, since this phenomenon appears to be a product of the local topography rather than a feature of the planetary circulation, it would be preferable if it were given a more restrictive label.

In summer, a major zone of frontal activity appears along the rim of the Polar Basin from Siberia eastward to the Alaskan coast and thence southeastward to Hudson Bay. The reasons for this frontal activity have already been discussed. Typical examples of developments in this area will be presented in section 3.5.

### 3.3 Jet Streams and Tropopause

Jet streams are necessary counterparts of deep frontal zones, and wherever such zones occur, bands of high winds appear above them. In view of the regular appearance of fronts at high latitudes, jet streams are a common feature, too. Shallow fronts connected with topographical features are not, of course, accompanied by high level jets. On the other hand, definite wind maxima often occur aloft above diffuse baroclinic zones which are lacking surface fronts. Although there is not a one to one correspondence between fronts and jet streams at high latitudes, nevertheless, their frequencies are probably not very different.

In this connection, it is unfortunate that the term the "jet stream"--implying a single river of high winds encircling the globe at middle latitudes--is in such common use. Like the use of the terms the "polar front" or the "arctic front," it represents a gross oversimplification of the

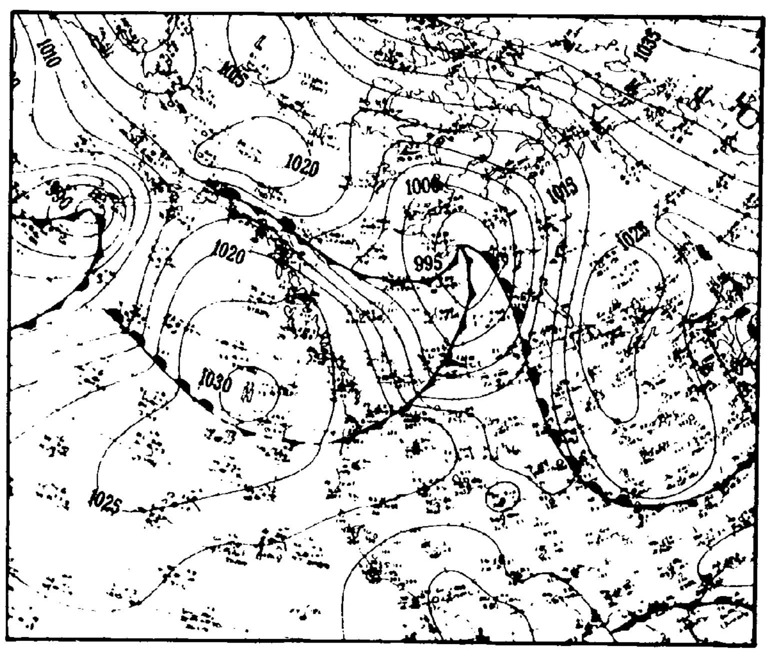



Figure 3.5. Surface Chart for 1230 G.C.T., 14 February 1957. (Northern Hemisphere Historical Series.)

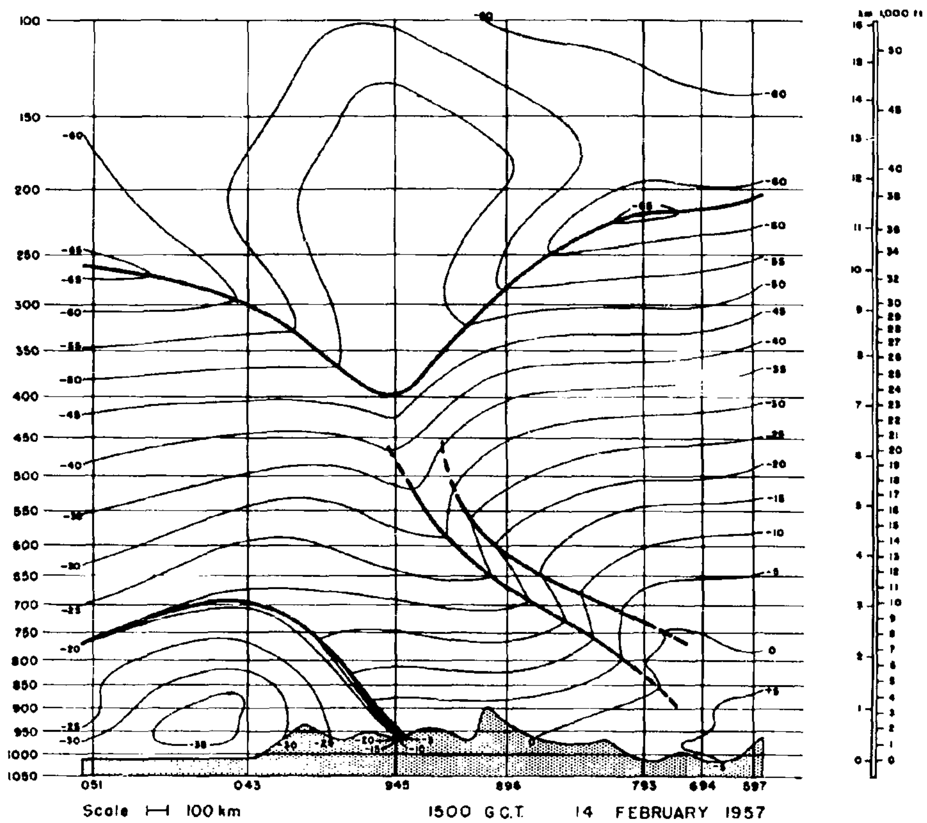



Figure 3.6. Cross Section from Medford, Ore., (597) to Sachs Harbor, N.W.T., (051). Thin Solid Lines are Isotherms in °C. Heavy Lines Represent Discontinuities.

atmospheric circulation as can easily be verified by examination of hemispheric maps. In reality, the atmosphere possesses at any given time a number of jets of varying length which may branch or merge but which seldom form a simple pattern. This behavior is consistent with the previously noted multiple and fragmentary nature of frontal systems.

The degree of organization of jet streams depends highly on the state of the general circulation, and occasional instances of very simple and stable jet configurations can be found. Even in such cases it is usual for the jet to occur in two or more separate bands, one of which may be located at high latitudes.

The foregoing remarks serve to emphasize the fact that jet streams are a regular feature of the arctic circulation.[1] Some of the characteristics of arctic jets during the winter season are shown in figure 3.8, which is a composite of nine individual cases. The jet core lies at about the 300-mb., or 30,000-foot, level at a mean latitude of 73° N. In each case a second, and usually stronger, jet (not shown) is present at lower latitudes.

Although the individual wind maxima were moderately intense by high latitude standards, the average maximum speed of 65 knots is not large in comparison to speeds observed in typical middle latitude jets. However, in extreme cases,

[1] Probably most of the currents of relatively high winds in the vicinity of the arctic tropopause would not qualify as jet streams, according to the strict definition recently proposed by the World Meteorological Organization.

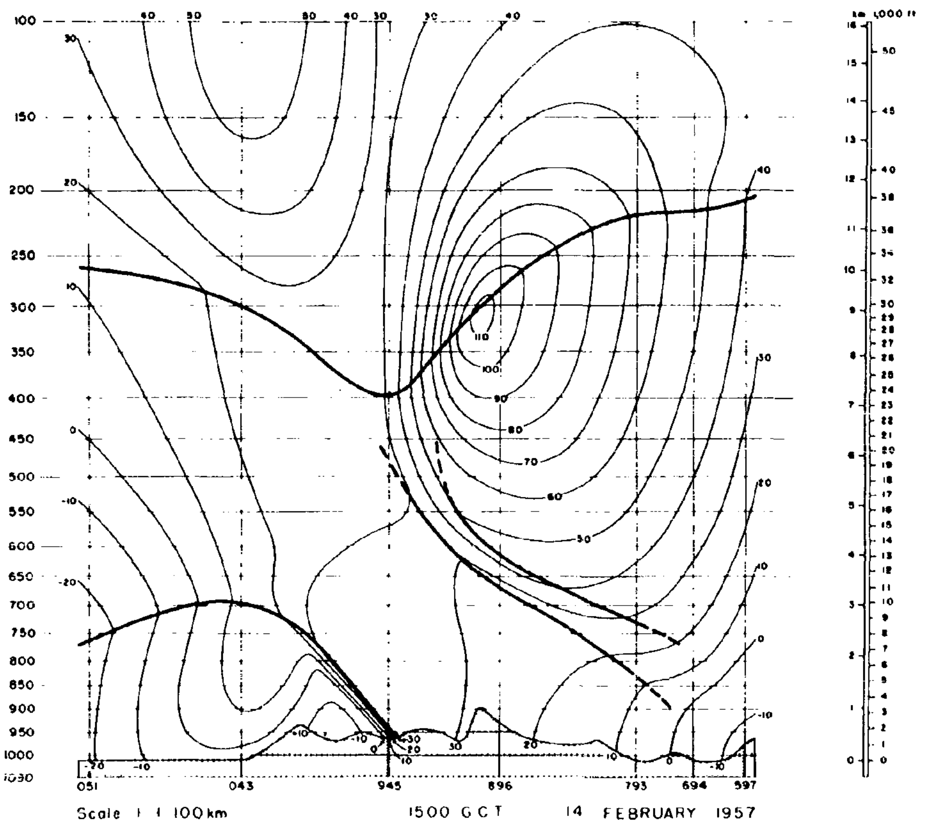



Figure 3.7. Cross Section from Medford, Ore., (597) to Sachs Harbor, N.W.T., (051). Thin Lines Are Isotachs of the Geostrophic Wind Component Normal to the Section (in Knots).

the figure of 65 knots may be greatly exceeded, and jet maxima in excess of 100 knots may occur north of the Arctic Circle at all seasons. These usually are the result of a confluent wind field which brings air currents from high and low latitudes into juxtaposition. Since the strongest northward surges of air at high altitudes are often associated with large, blocking ridges near Alaska or Scandinavia, the areas north of these regions are favored locations for the extreme cases.

Although an effort was made to preserve, on the composite, any thermal concentrations or frontal zones that appeared on individual sections, it is seen from figure 3.8 that the baroclinic zone associated with the jet is broad and not bounded by recognizable discontinuity surfaces. The temperature field at 300 mb. or 250 mb. can frequently be used as an aid in locating jet cores at middle latitude. To the north of the jet axis the temperature warms, while a band of cold temperature parallels the jet near or just south of the axis. There is some suggestion of the same structure in figure 3.8, but the features are not clearly defined.

In the upper left corner of the figure a separate area of high wind may be noted. This is the lower part of a recently discovered feature of the arctic atmosphere known as the "arctic stratospheric jet stream" or "polar night jet."

The tropopause in figure 3.8 varies in height from about the 250-mb. level south of the jet to slightly below the 300-mb. level to the north. Individual cases often show higher and lower values. In well-developed polar lows the tropo-

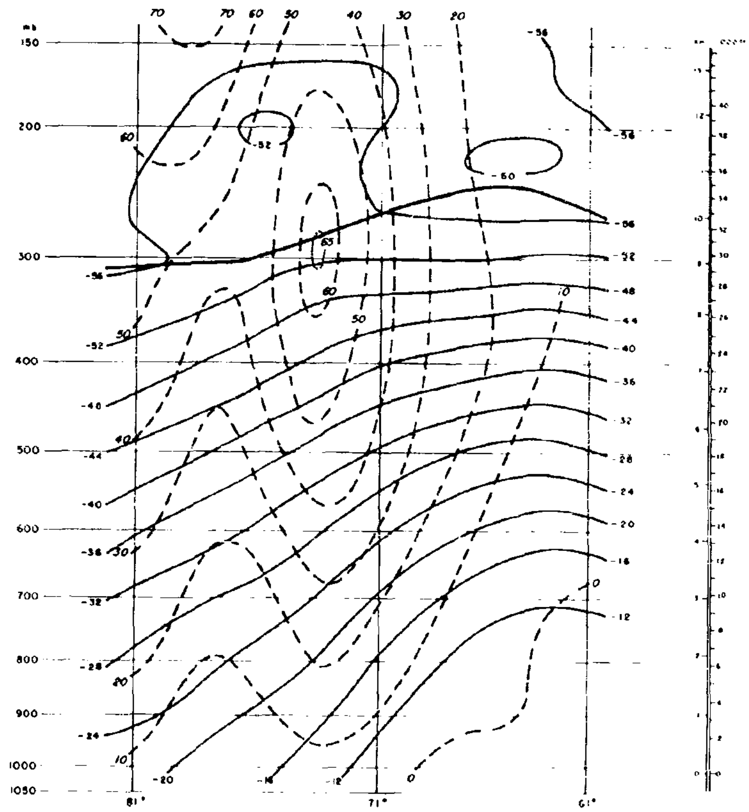


Figure 3.8. Mean Cross Section. Thin Solid Lines, Isotherm in °C.; Thin Dashed Lines, Normal Geostrophic Wind Components in Knots; Heavy Solid Line, Tropopause.

pause may sink to 400 mb, or below, and in winter may disappear altogether. The tropopause may also be indistinct near the jet stream, giving rise to a tropopause break. In such cases the higher segment of the tropopause will always lie on the anticyclonic side of the jet and the lower segment on the cyclonic side.

## 3.4 Distribution of Cloud and Precipitation in Baroclinic Disturbances

Very little has been published on this subject, and our treatment here will be confined to a brief review of the results of a study of the cloud distribution in frontal cyclones over the Arctic Ocean, as revealed by aerial reconnaissance, and of the relationship of clouds and precipitation to fronts in the North American Arctic, as revealed by 3-hourly surface observations.

The continual development and movement of baroclinic disturbances across the path of the Ptarmigan reconnaissance flight, during the first 3 weeks of July 1956, made it possible to locate each cloud observation at the appropriate point on the model displaying the fronts, jet stream, and cold low center shown in figures 3.9 and 3.10. From averaging of the data, cloud amounts were determined by 5,000-foot layers between the ground and 20,000 feet, and for the layer above. Only the distributions in the layers between 5,000 to 10,000 feet and 15,000 to 20,000 feet are shown in the figures.

In the layer 0 to 5,000 feet (not shown) cloud cover averaged nearly 100 percent and exhibited only a slight relationship to the synoptic pattern. The cloudiness in this layer, the characteristic summer stratus, thinned out somewhat near the southern edge of the diagram due to the nearness

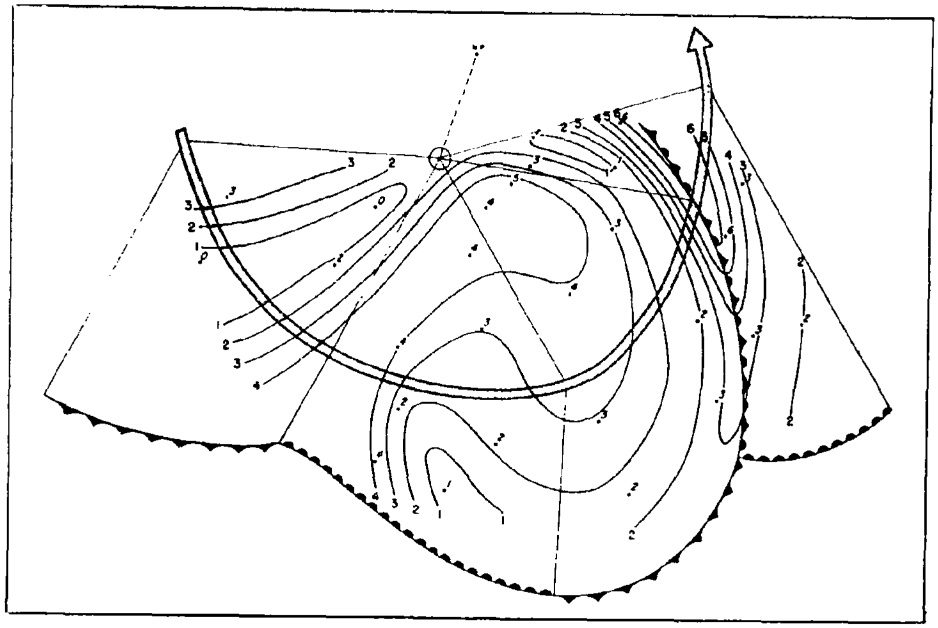


Figure 3.9. Composite Chart of Cloud Distribution in a Layer from 5,000 to 10,000 Feet. Mean Cloud Amounts in Oktas (Eights of Sky Covered) Are Plotted at Points of Averaging. Circle Denotes Low Center at 500-mb.; Arrow Denotes the Jet Axis at that level.

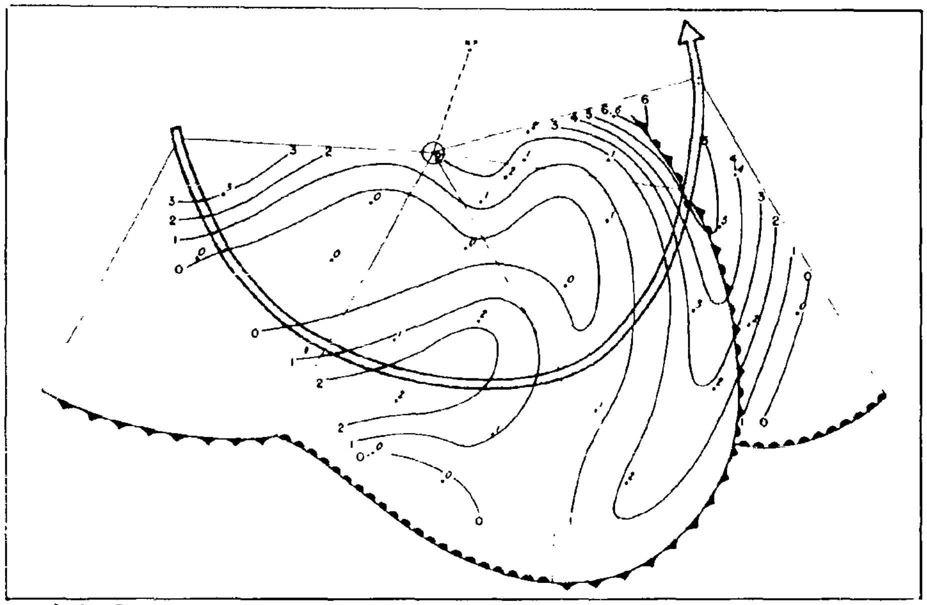


Figure 3.10. Composite Chart of Cloud Distribution in a Layer from 15,000 to 20,000 Feet. Mean Cloud Amounts in Oktas (Eights of Sky Covered) Are Plotted at Points of Averaging. Circle Denotes Low Center at 500-mb., Arrow Denotes the Jet Axis at that Level.

of land. Between 5,000 and 10,000 feet (fig. 3.9), cloudiness is extensive over the surface occlusion, near the crest of the wave, and in the southeast quadrant of the cold low. A strip of nearly clear skies lies behind the cold and occluded fronts, and a cloud-free area is located behind the peak of the wave. The clear strip shifts westward with height (fig. 3.10), and the cloudiness to the southeast of the upper low center disappears, evidently being connected with the low cloud deck in that region rather than with the main cloud masses of the storm systems. The total amount of cloudiness diminishes with height, but the patterns are essentially unchanged except for the differences just noted.

From the preceding discussion it is evident that, above the surface layers, the broad features of the cloud distribution in arctic disturbances are in agreement with accepted models for frontal-type cyclones. In matters of detail there appear to be some deviations from the models, and also individual cases may differ considerably from the typical or composite situation. When frontal surfaces are identifiable at upper levels it appears that, as a rule, the cloud mass is not inclined along the front but has a lesser inclination. Moreover, the cloud mass tends to be broken up into a number of discrete layers.

The range of individual variation in the relationship of clouds and precipitation to fronts is brought out by the examples in figure 3.11. Part of this variation is no doubt due to surface effects and part to variations in the humidity distribution and in the strength of vertical motions from case to case.

3.5 Illustrative Example

The foregoing remarks on the structural features of baroclinic disturbances will now be illustrated by reference to a specific case.

July 1956 was a month of above normal weather activity over the polar area, offering many examples of cyclogenesis along the northern shores of Siberia, Alaska, and Canada. Early in the month a wave disturbance moved rapidly eastward along the Siberian coast, from near the Taymyr Peninsula to the Chukchi Sea, where it underwent a spectacular deepening and spun northward to a position near 80° N. and 180° longitude. The central pressure on the surface chart reached a minimum of 975 mb. on the 6th of July. The upper-level low, on this date, was located almost directly above the surface system and was very intense, showing a central height of approximately 16,500 feet at 500 mb.

During the next 2 weeks weather events over and adjacent to the Polar Basin literally revolved about this mighty low, which gradually filled and wobbled erratically about in the area north of the East Siberian and Chukchi Seas. Being of typical cold low structure in its later stages, the low may be visualized as a large pool of cold air in cyclonic rotation. Within the center of the pool the temperatures were coldest, and they warmed progressively towards the rim. Because of the appreciable and nearly uniform temperature gradient between center and circumference, it is not correct to think of the cold pool as a more or less homogeneous arctic air mass in the classical sense.

During the week prior to the time of the first map shown here (fig. 3.12), a series of minor lows skirted the cold pool, passing along a frontal system at the outer rim. In each case these lows appeared to be initiated by perturbations within the cold pool itself or by disturbances which entered the Arctic from middle latitudes, and in each case they eventually were absorbed into the main circulation. Remnants of two such systems appear over the Canadian Archipelago and northern Greenland in figure 3.12.

The system of principal interest on the chart for July 14, however, is the incipient wave near the Chukchi Peninsula. During the next 2 days this wave deepened rapidly as it moved northeastward across the Beaufort Sea, and on the map for July 16 it appears as a deep occlusion over the Arctic Ocean to the northwest of the Canadian Archipelago. The thickness lines in figures 3.12, 3.13, and 3.14 bring out the important thermal characteristics of the disturbance. The surface low forms at the outer edge of the region of strong temperature contrast and during the occlusion process becomes engulfed in the cold air pool.

Over land areas the surface frontal passages were marked by pronounced temperature changes. For reasons explained previously, the situation was very different over the ice pack where temperatures fluctuated only a few degrees during the passage of the storm. Unfortunately, dropsondes from the Ptarmigan flight were not taken at the right places and times to illustrate the large temperature changes taking place just above the surface cold film. However, prevailing winds at Sachs Harbor, on the shores of Banks Island, were from the direction of the ice pack throughout the period so that soundings for this station can be used to illustrate the nature

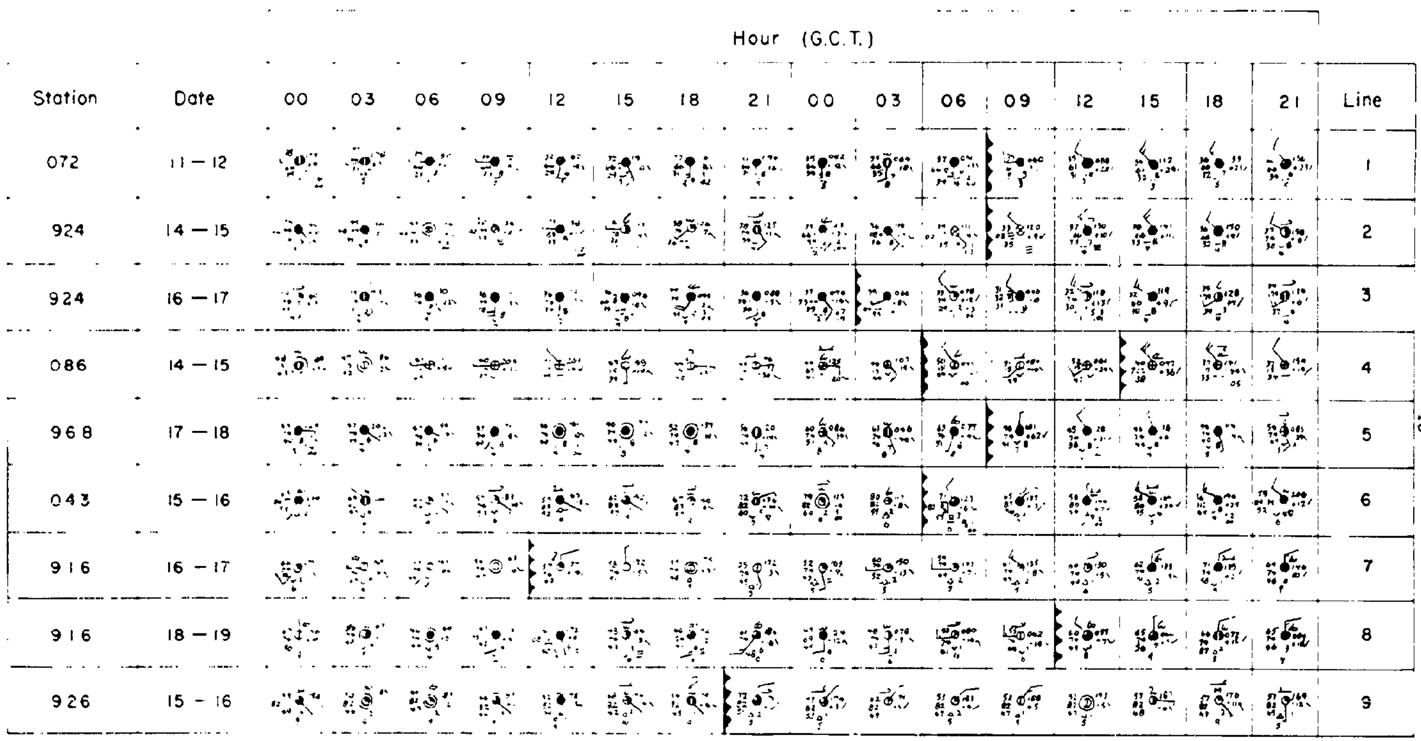



Figure 5.11. Plots of 3-Hourly Surface Observations in Usual Code, Illustrating Typical Frontal Passages. The Type of Front and Time of Passage Is Denoted by the Appropriate Symbol.

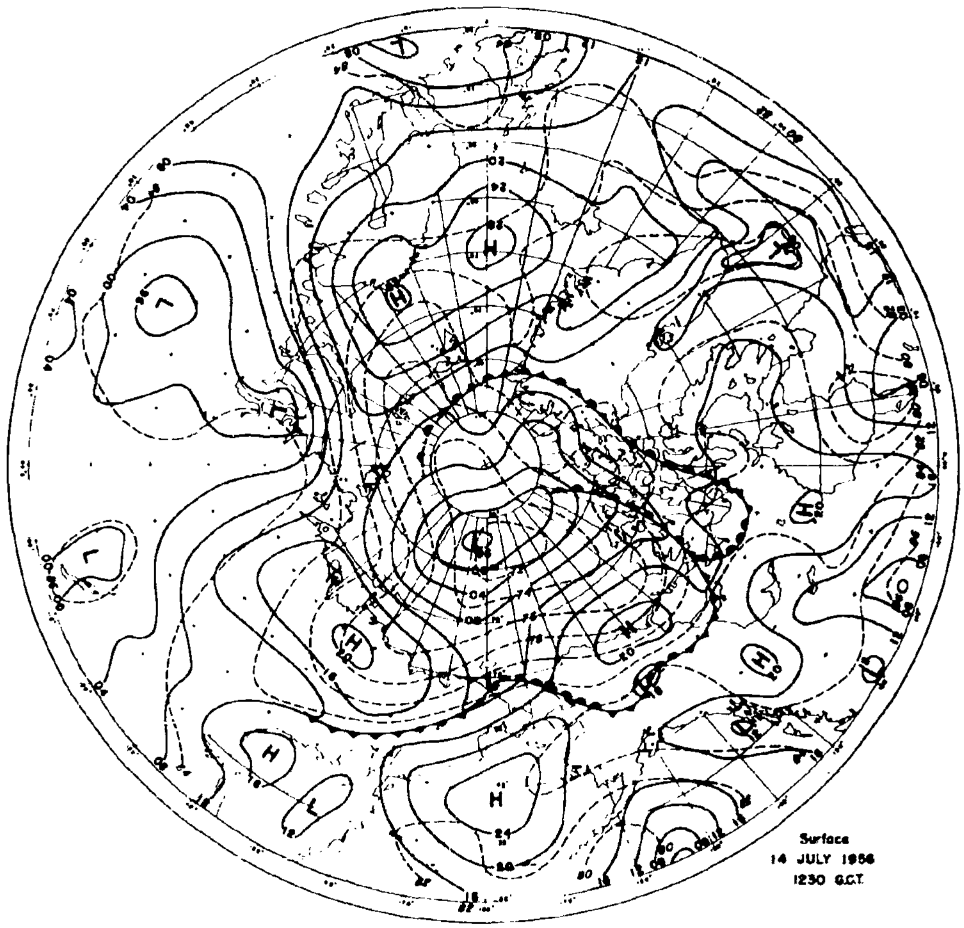



Figure 3.12. Surface Chart for 1230 G.C.T., 14 July 1956. Dashed Lines Are 1000- to 500-mb. Thickness at Intervals of 200 Feet.

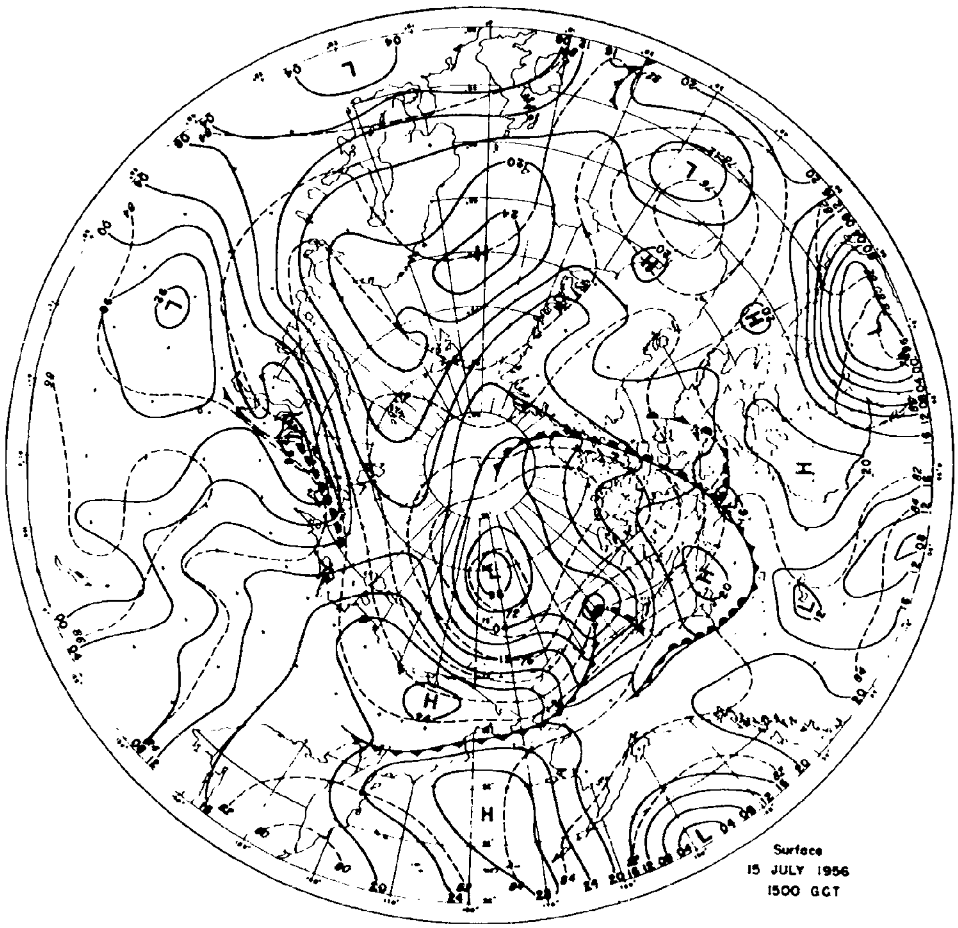



Figure 3.13. Surface Chart for 1230 G.C.T., 15 July 1956. Dashed Lines Are 1000- to 500-mb. Thickness at Intervals of 200 Feet.

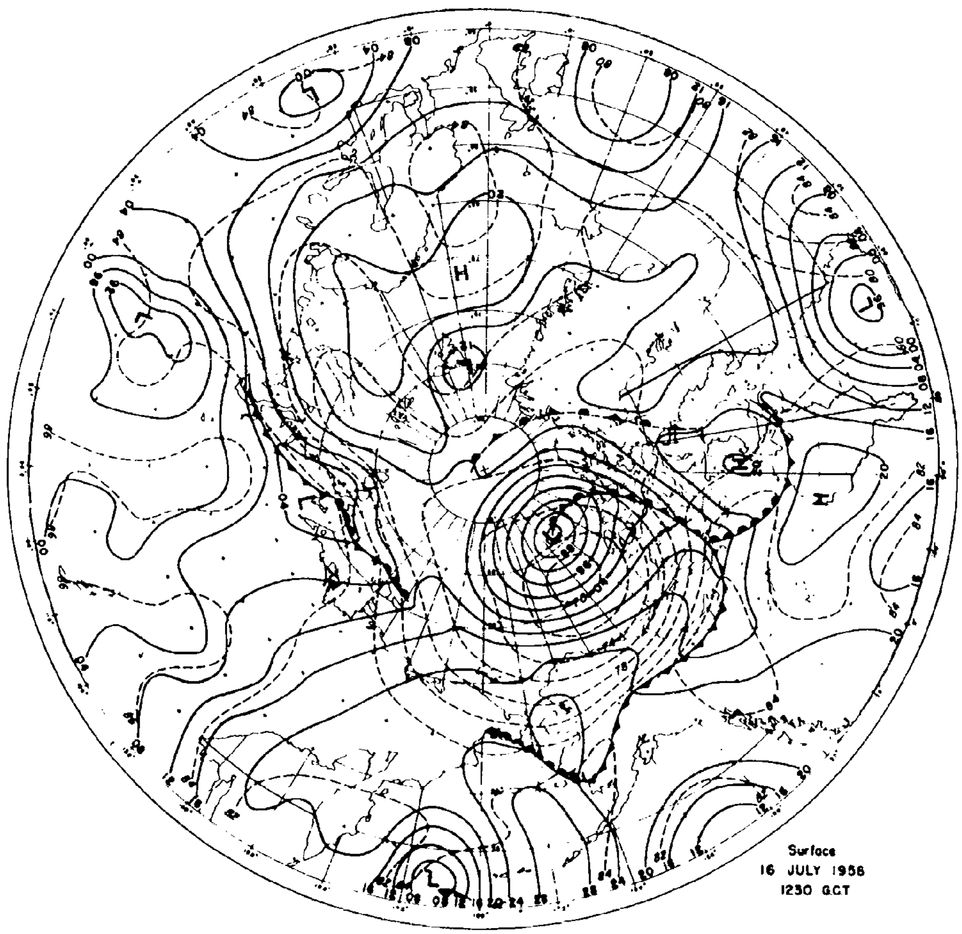



Figure 3.14. Surface Chart for 1230 G.C.T., 16 July 1956. Dashed Lines Are 1000- to 500-mb. Thickness at Intervals of 200 Feet.

of the temperature changes at the different levels.

From the soundings (fig. 3.15), it is seen that the surface temperature dropped only 5° F. (from 37° F. to 32° F.) with the frontal passage. At upper levels the change is much larger--20° F. at a height of only 2,000 feet. In cases such as this, in which surface temperatures are of little help in locating the fronts, isobaric configuration and pressure tendencies may be used to good advantage. Figures 3.12, 3.13, and 3.14 provide excellent examples of the relationship between fronts and isobars during various stages of cyclone development.

To illustrate three-dimensional aspects of the frontal, jet stream, and tropopause structures, and the distribution of clouds and precipitation in the baroclinic disturbance, we turn now to the situation of July 20, 1956. During the 4-day interval from the previous example, the cold pool drifted slowly southeastward and is found entering the western part of the Canadian Archipelago on the map for July 20 (fig. 3.16). The earlier occlusion vanished during the interim, while a later occlusion, which developed from a wave near Great Slave Lake, N.W.T. on the 18th, is present over northern Baffin Bay. The original low at the surface is still discernible as an elongation of the low pressure area in the direction of Mould Bay (072).

A second frontal system may be noted in figure 3.16, extending along the northern coasts of Canada and Alaska to a low pressure area over the East Siberian Sea. The beginnings of this front are to be found on the chart for the 15th (fig. 3.13) in the frontogenesis entered along

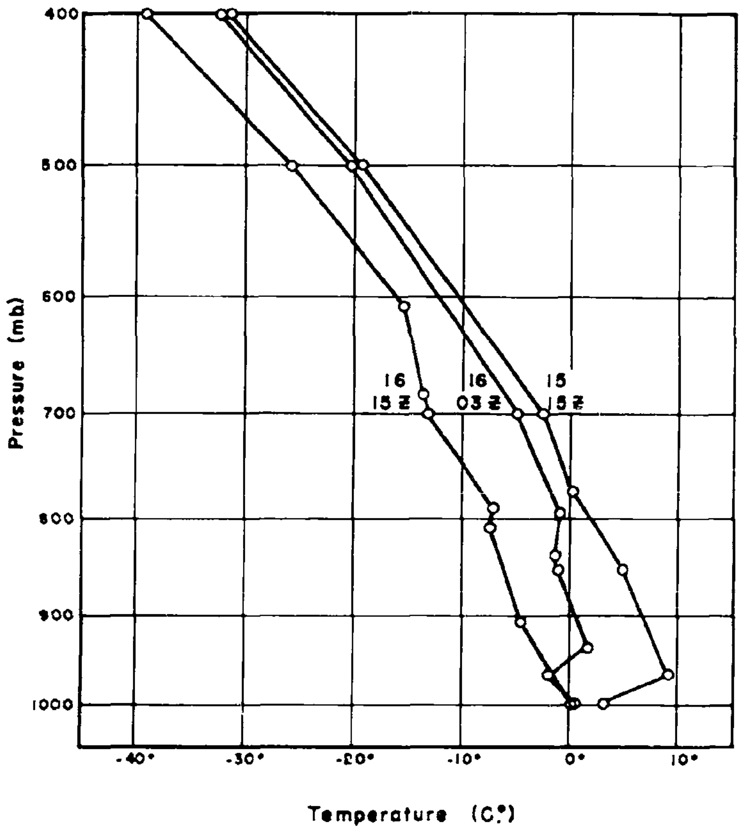



Figure 3.15. Successive Soundings at Sachs Harbor (051) During Period 1500 G.C.T., 15 July 1956 to 1500 G.C.T., 16 July 1956.

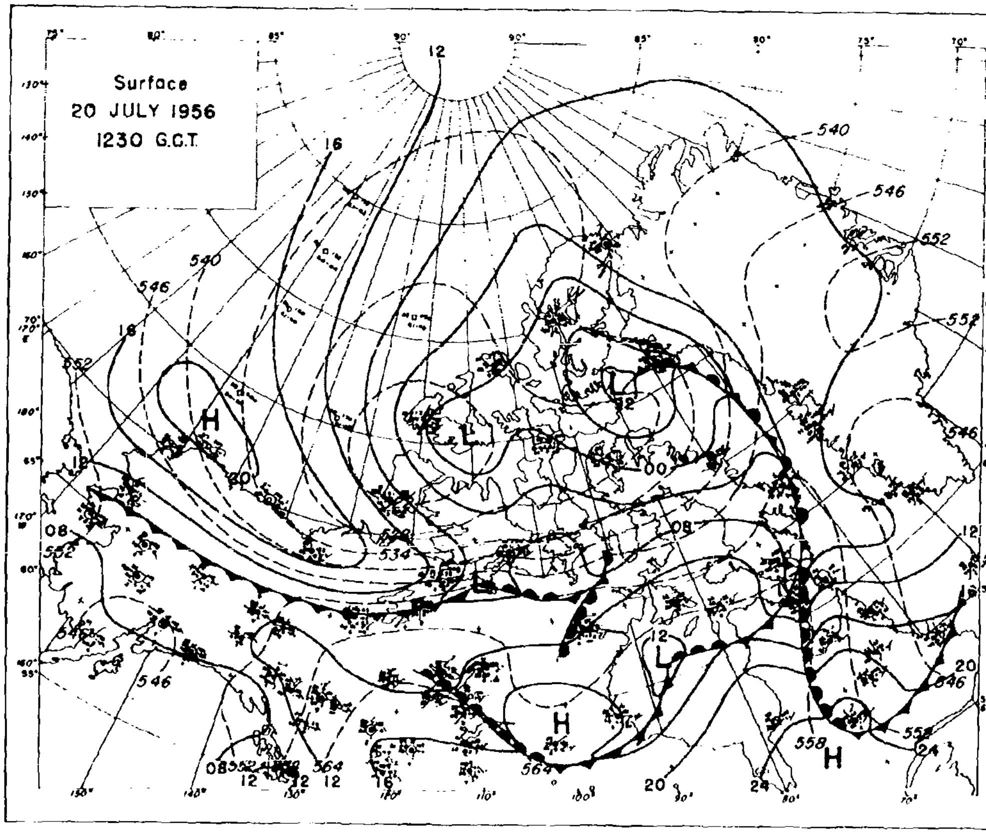



Figure 3.16. Surface Chart for 1230 G.C.T., 20 July 1956. Dashed Lines Are 1000- to 500-mb. Thickness Contours in Geopotential Decimeters.

the north Siberian coast. The frontogenesis gradually progressed eastward in association with minor disturbances along the periphery of the cold pool. One of these disturbances is located just south of Coppermine (938) on the map for the 20th (fig. 3.16).

Cross sections of temperature, wind, and cloud distribution are shown in figures 3.17 and 3.18. The sections extend from Norman Wells, Canada, (043), to Eureka, Ellesmere Island, (917). A main feature of the tropospheric temperature field is the cold air pool centered between Mould Bay (072) and Sachs Harbor (051). To the north the cold pool warms gradually; however, to the south the temperature gradient, especially at lower levels, becomes concentrated to the rear of a sloping frontal boundary. Further south in the warm air, temperature gradients are small but not absent, and in the upper troposphere it is impossible to distinguish between warm and cold air masses.

The tropopause in figure 3.17 is extraordinarily sharp, giving rise to a very simple and distinct temperature structure in the transition layer between troposphere and stratosphere. Above the tropospheric cold pool the tropopause dips to its lowest levels, and stratospheric air that is warm, relative to its surroundings, lies within the hollow. The coldest air in the stratosphere coincides with areas of high tropopause. In the transition layer the coldest temperatures are found along the tropopause, so that adjacent regions of both troposphere and stratosphere are relatively warm.

The jet core (fig. 3.18) is located on the

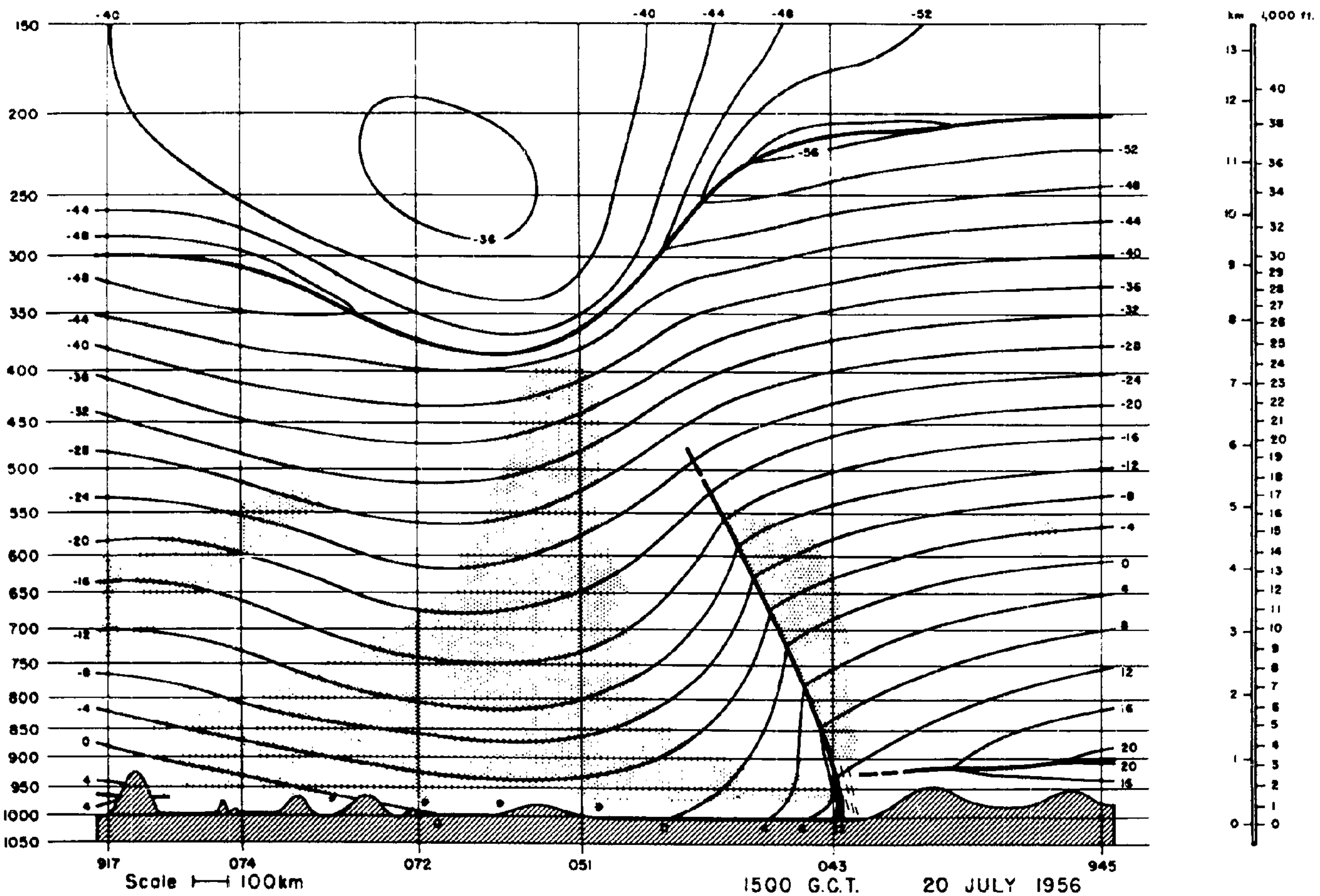



Figure 3.17. Cross Section from Eureka (917) to Fort Smith (945) Showing Temperature and Cloud Distributions at 1500 G.C.T., 20 July 1956. Thin Solid Lines Are Isotherms in °C.; Heavy Solid Lines Are Tropopause, Frontal, and Inversion Discontinuities. Cloud Layers Are Depicted by Stippling. Short Slanting Lines Represent Rain; Asterisks, Snow.

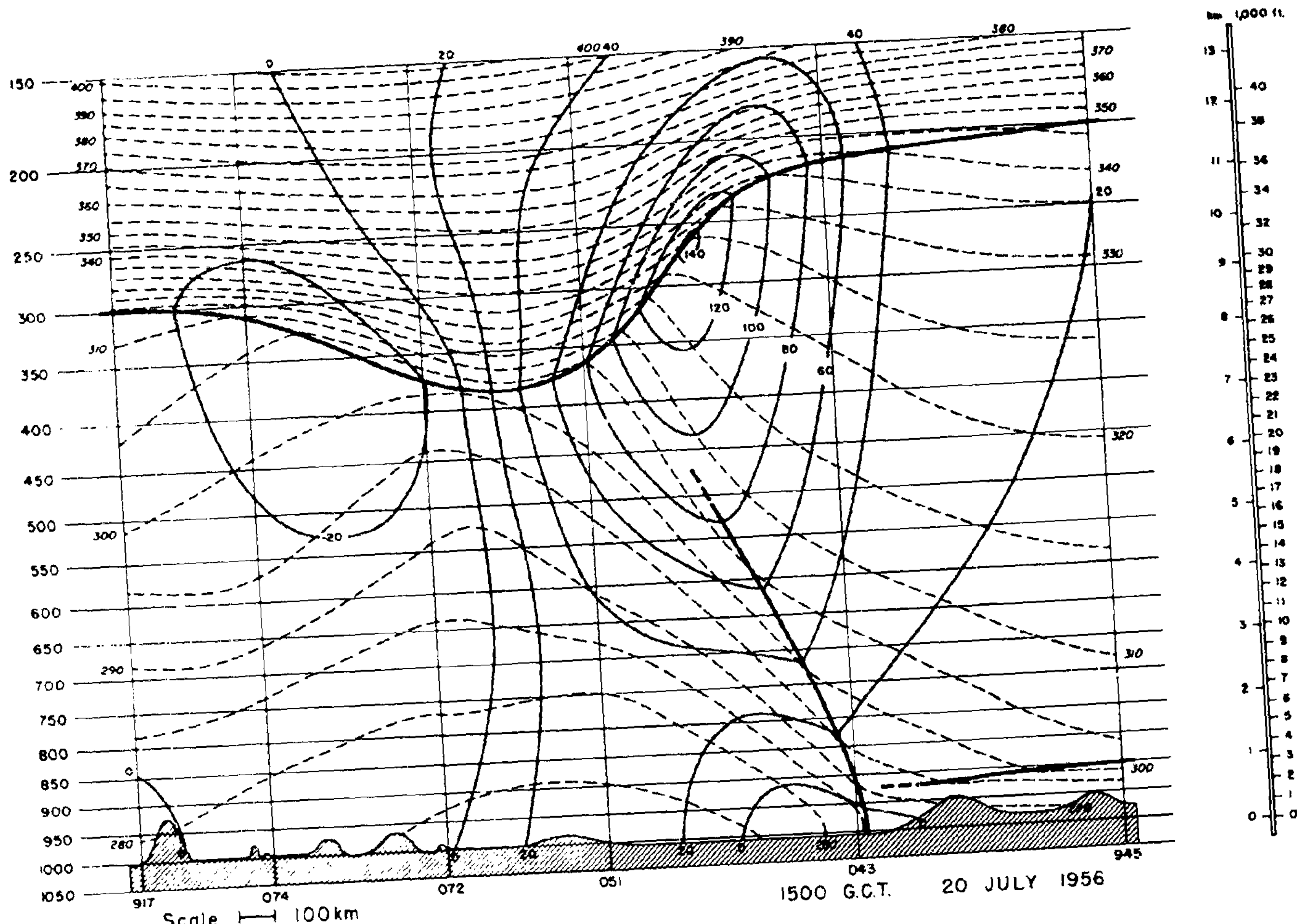



Figure 3.18. Cross Section from Eureka (917) to Fort Smith (945) Showing Potential Isotherms (Thin Dashed) and Normal Geostrophic Wind Components (Thin Solid) at 1500 G.C.T., 20 July 1956. Temperatures Are in °A., Wind Speed in Knots. Heavy Solid Lines Are Tropopause, Frontal, and Inversion Discontinuities.

tropopause at the 270-mb. level and lies above the region of strongest temperature contrast in the middle troposphere. The maximum wind (estimated geostrophically) is 140 knots, a high value by middle latitude standards even for winter storms. Isotachs of the strongest winds extend downward along the frontal boundary. The relationships between the jet axis, tropopause, and temperature field are particularly striking at the 250-mb. level (fig. 3.19). The observed coincidence between the jet axis, the tropopause, and the strip of cold temperature is characteristic of only the strongest middle-latitude systems and cannot be regarded as a usual feature of arctic disturbances.

The cloud distribution in figure 3.17 is considered to be typical of cases in which the section passes through both an active cold front and a cold low. Only scattered clouds appear in the warm sector. A cloud mass rises up along the cold front to a height of about 15,000 feet. Scattered, light showers fall from this cloud system. To the rear of the cold front, strong subsidence dissolves the frontal clouds. However, a low stratus or stratocumulus persists at the top of a mixing layer next to the ground, trapped there by the cap of subsiding air.

An extensive cloud mass pervades the interior of the cold low, rising to the tropopause near the low center. Stations within the inner region all report occasional snow. Except when a cold low is in the later stages of dissolution, overcast skies and patchy areas of precipitation are the rule.

3.6 Cold Lows

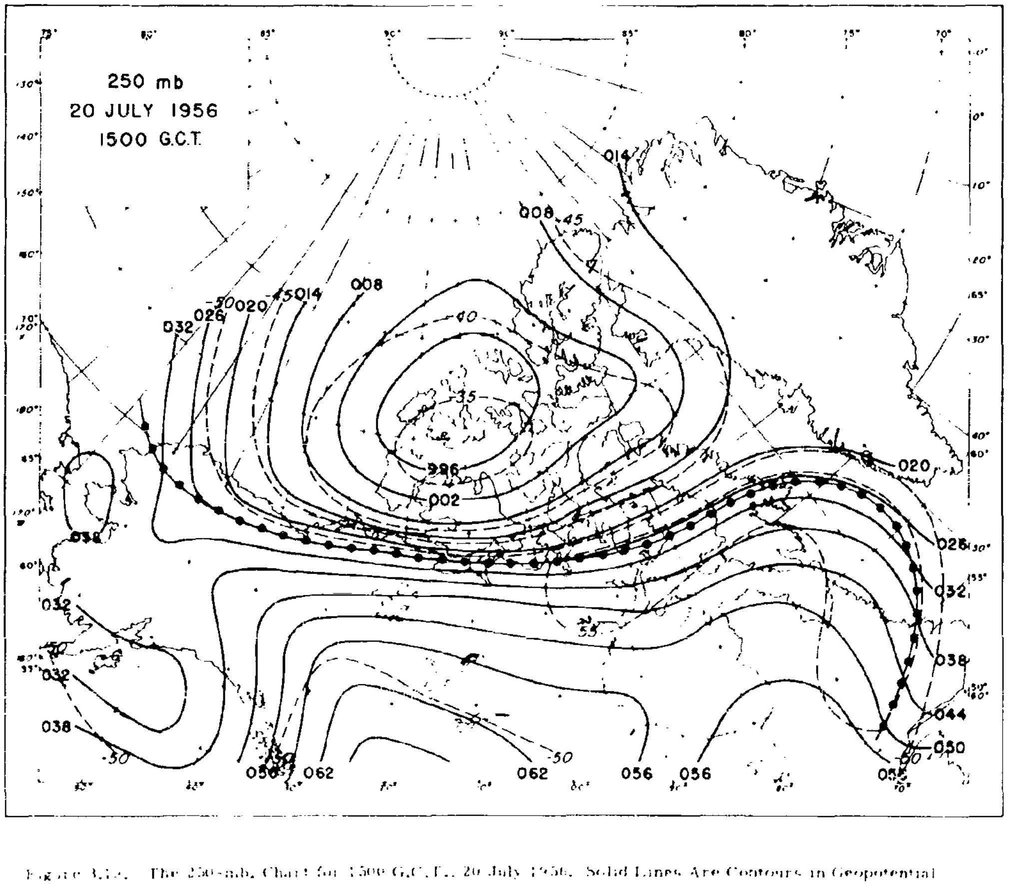



Figure 3.19. The 250-mb. Chart for 1500 G.C.T., 20 July 1956. Solid Lines Are Contours in Geopotential Meters (First and Last Digits Omitted), Dashed Lines Are Isotherms in °C., Beaded Line Represents Intersection of Tropopause with Pressure Surface.

The term "cold low" is applied to an upper-level cyclone which is more or less thermally symmetrical about the center of lowest pressure or height, the point of coldest temperature coinciding roughly with the low center. The circulation in the cold low is strongest at the tropopause and diminishes in intensity both upward and downward. The cyclonic circulation at the ground is often weak and may be lacking in some cases. As an example of a cold low, we may cite the quasi-permanent cyclone which hovered over the Polar Basin throughout the July series discussed in the previous section. Here we shall elaborate further on the characteristics of cold lows and examine a typical winter case.

It is often difficult to trace cold lows back to a distinct initial stage, partly because data is frequently lacking in critical areas and partly because the cold low normally represents an offshoot of a larger circumpolar vortex from which it may not always be clearly distinguished. In cases where a distinct origin may be noted, it appears that baroclinic developments play an important role in the formation. Thus, cold low formations north of Greenland and Spitzbergen in winter are connected with the northward migration of Atlantic depressions which, though occluded, still possess some degree of thermal asymmetry so that the cold low may be regarded, in large measure, as the end product of the occlusion process. The origin and maintenance of cold lows over the Polar Basin in summer are obviously related to the occlusion of baroclinic disturbances, as shown in the previous section.

The case of January 2, 1957, provides a good example of a cold low situation. The 500-mb. chart for the date (fig. 3.20) shows a deep low centered over Ellesmere Island with an inner temperature of -47.9° C. reported at Eureka (917). The isotherms do not coincide exactly with the contours, the center of gravity of the cold air being displaced somewhat to the northwest of the low center, and the thermal concentration being considerably larger to the south of the center than to the north. A band of relatively high winds encircles the low, with Thule, Greenland (202) reporting a speed of 50 knots. Further to the south, near Sachs Harbor (051), a reported wind of 55 knots marks the entry of a second jet into the cold low system. The cyclonic circulation exhibits the characteristic weakening downward, so that only a faint reflection of the upper

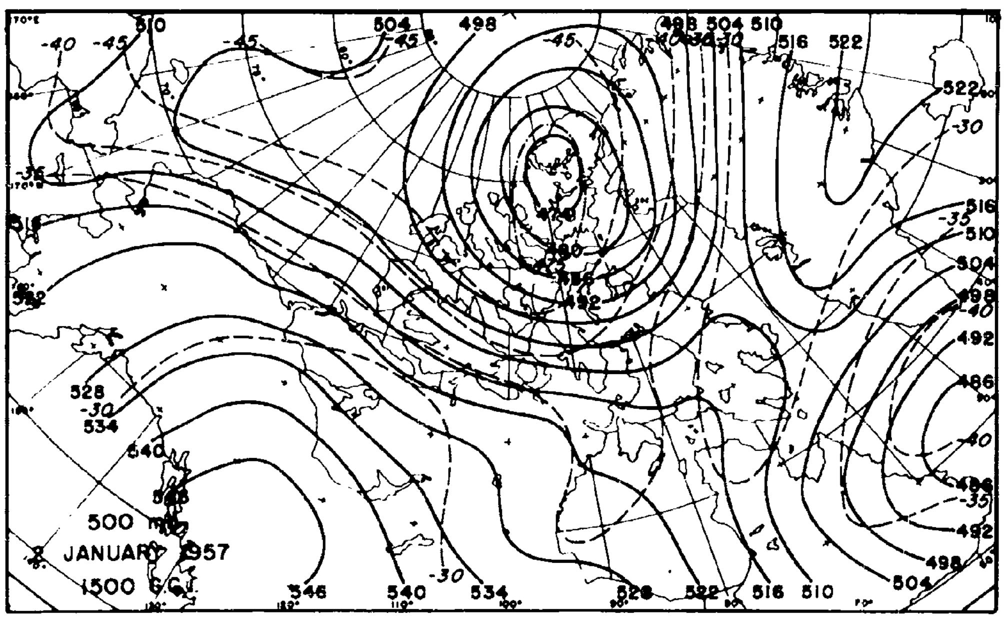



Figure 3.20. The 500-mb. Chart for 1500 G.C.T., 2 January 1957. (Isolines Defined in Fig. 3.19).

low may be detected on the surface chart (fig. 3.21) in the form of an extension of the trough from Baffin Bay.

The cross section in figure 3.22 gives added perspective to the temperature and wind structures. As in the summer case discussed previously, the "cold air mass" is not homogeneous but possesses a fairly uniform temperature gradient that extends from near the surface to about 400 mb. Distinct, though relatively weak, wind maxima lie above the zones of strongest temperature concentration near Isachsen (074) and Alert (082). The concentrations are not sufficiently pronounced to be regarded as fronts and are not connected with surface frontal systems.

Surface temperatures in the area of the cold low are controlled largely by cloud cover and wind speed. Thus, Eureka at the center of the low, with an overcast sky and a 16-knot wind, has a higher surface temperature than stations further removed from the center.

A second deep baroclinic zone without definite frontal structure is evident in the vicinity of Sachs Harbor (051). A more prominent high latitude jet stream is connected with this second zone, a maximum geostrophic speed of 100 knots occurring at the 300-mb. level. At the southern end of the cross section, the tropopause is located near the 250-mb. level and dips to the 400-mb. level upon approaching the low center. Near the center the tropopause is indistinct, as is often the case in winter situations.

During the cold season, weather conditions in cold lows are highly variable so that it is questionable whether it is permissible to generalize from a single case. However, certain aspects of the present example are considered to be representative. It will be noted from the surface chart that the sky is overcast beneath the low center aloft and that the cloudiness diminishes outward. Most peripheral stations report clear skies. At several stations (Isachsen (074), Resolute Bay (924), Thule (202)) close to the center, ice needles are observed, and at one of these (Resolute Bay) a trace of snow has fallen in the past 6 hours.

Generally speaking, cold lows are characterized by much less cloudiness and precipitation in winter than in summer. On occasions

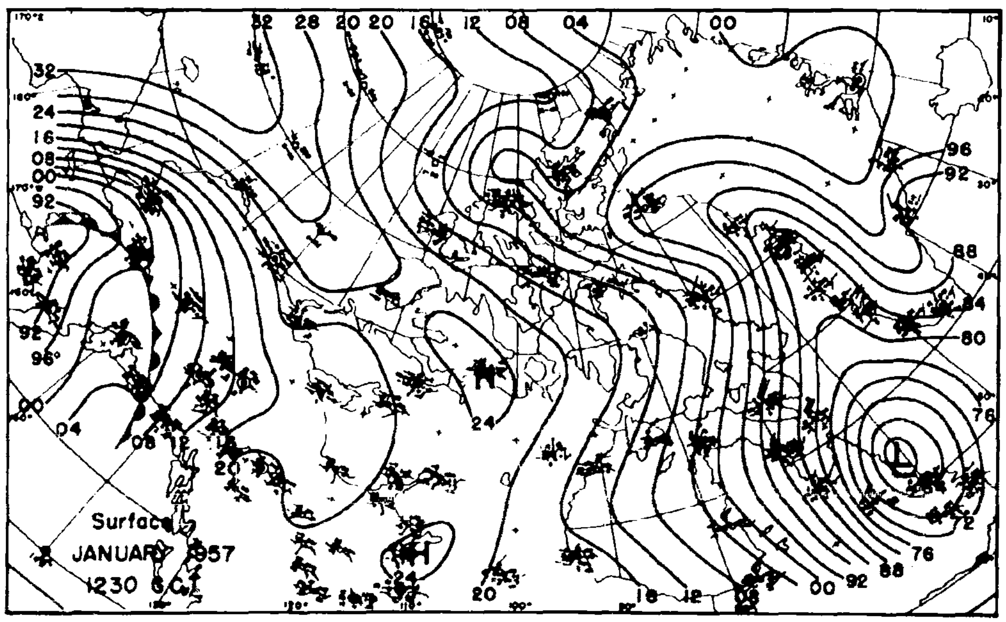



Figure 3.21. Surface Chart for 1230 G.C.T., 2 January 1957.

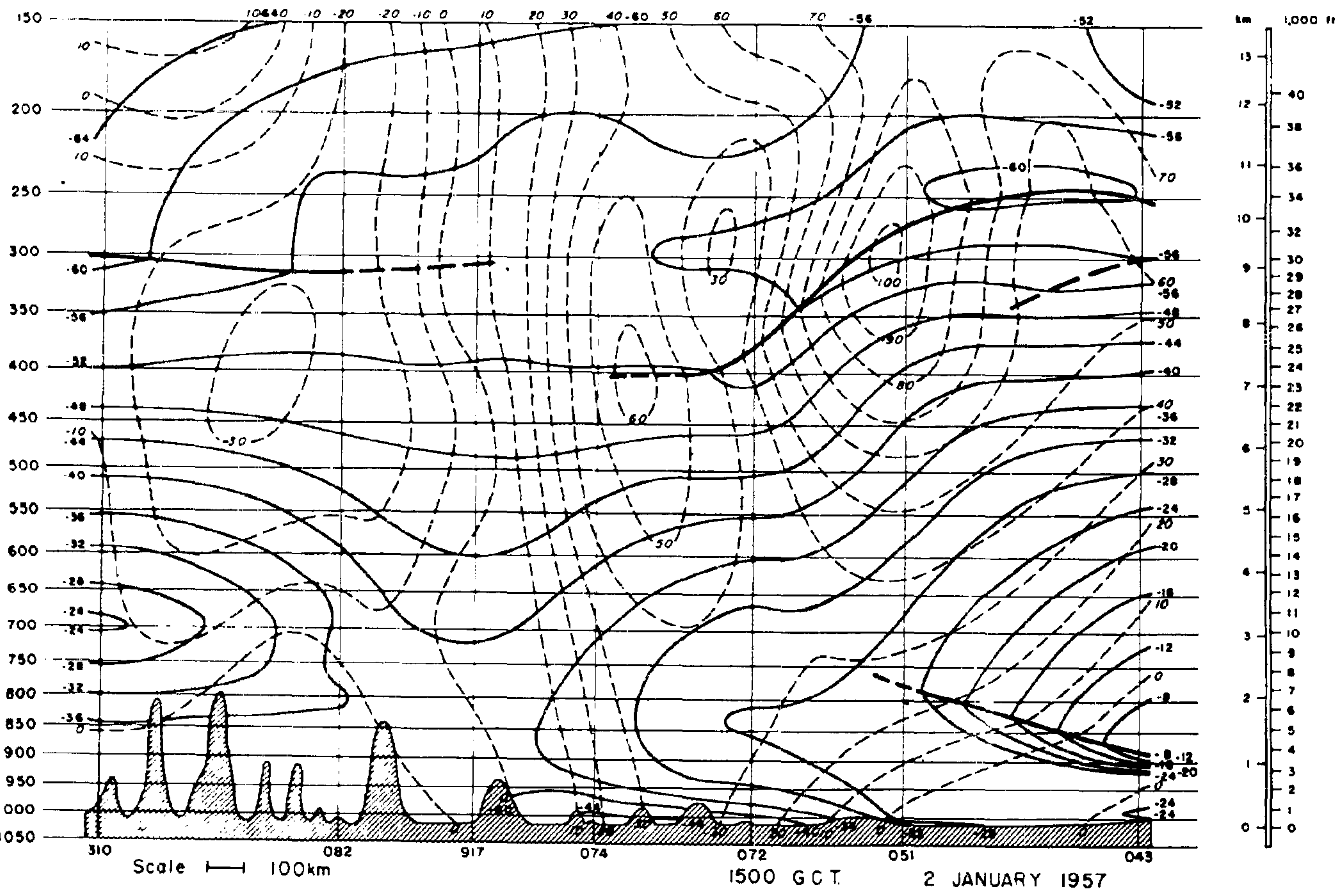



Figure 3.22. Cross Section from Nord, Greenland (310) to Norman Wells, Canada (043) for 1500 G.C.T., 2 January 1957. Medium Solid Lines Are Isotherms in °C.; Heavy Solid Lines, Discontinuities; Dashed Lines, Normal Geostrophic Wind Speeds in Knots.

skies may be clear throughout the area of the low, but usually in such cases a number of stations report ice needles so that, in fact, a form of condensation (or deposition) occurs.

The differences between weather characteristics in summer and winter can be accounted for by differences in temperature. At the low winter temperatures, the liquid water content of the air is extremely small, of the order of one-tenth the summer value in the Arctic and one-hundredth the value in temperate latitudes. Moreover, when temperatures are below -40° F. water vapor may be precipitated directly in the solid form without first passing through the liquid phase. Because of the very small amount of moisture available and the possibility of direct transformation from vapor to the solid phase, it is not surprising that condensation phenomena should appear different in the two seasons.

### 3.7 Low-Level Inversions

Low-level temperature inversions frequently occur over the entire polar region in winter and over ice and snow covered areas in summer. The physical processes that give rise to these inversions are in no manner peculiar to high latitudes. However, certain of the processes are particularly intense or prolonged at these latitudes so that arctic inversions are often marked by unusual depth, strength, and persistence. We shall review here, briefly, the main processes connected with the formation and maintenance of arctic inversions and shall present examples of some common types.

The great bulk of polar inversions may be classed as radiation inversions; that is, inversions formed primarily by a net loss of radiation from the surface. When a radiation deficit develops, the temperature of the surface drops quite rapidly thereby reducing the emission of long wave radiation and increasing the conduction of heat from below. Finally, a state of equilibrium is reached in which the net outgoing radiation is balanced by molecular heat conduction from subsurface layers and by eddy heat transfer from the atmosphere to the surface. In this equilibrium state, the air temperature is coldest next to the surface and increases upward to heights which vary from a few hundred feet to several thousand feet, depending on the individual situation.

An example of a typical radiation inversion is shown in figure 3.23 (A). Note the rapid increase of temperature with height in the lower 70 mb. (1025 mb. to 955 mb.--1,600 feet) and above this the isothermal layer extending to about 780 mb. (6,000 feet). The dual aspect of the inversion is a common feature and has led some investigators to introduce separate terms for the two components. The term "ground inversion" is often applied to the shallow, intense lower portion and "main inversion layer" to the thicker upper stratum. However, no physical reason has been offered for the separation, and it may be that the division in many cases is due to an economy in coding rather than to a real discontinuity.

The weather conditions at the time of the sounding are plotted above the ascent curve. Note the clear skies and light winds, factors which generally favor the formation of radiation inversions. The wind at gradient level is also shown.

A second class of arctic inversion, and perhaps the only other type of importance, is the so-called advection inversion. This type forms when relatively warm air blows over a cold surface that is not able to adjust its temperature to the air temperature. As an example of such a surface we may cite the top layer of the pack ice during the summer melt-season or the surface of an extensive snow field during periods of thaw. Heat loss by long-wave radiation gradually cools the upper warm portion or "nose" of the inversion so that its strength depends largely on the initial warmth of the air and on the length of time it has travelled over the cold surface.

The dropsonde in figure 3.23 (C), taken over the Beaufort Sea on 4 July 1956, affords a striking example of an advection inversion. Exact surface weather information is lacking, but from flight observations and analyzed charts it is known that a low overcast prevailed and that warm air was being advected northward in advance of a deepening cyclone.

Variations in sky cover, wind speed, and vertical motion produce important modifications of inversions. The effects of sky cover are particularly pronounced in the case of the radiation inversion. Clouds of sufficient thickness completely absorb the outgoing long-wave radiation from the surface and, being perfect absorbers, they emit as black bodies at their own temperature. If the cloud temperature is higher than the surface temperature--as will often be the case--the emission from the cloud base will be greater than from the surface, and the surface temperature will rise. The cloud top is cooled by radiation to space, but since the cooling is distributed throughout the cloud by mixing, only a minor

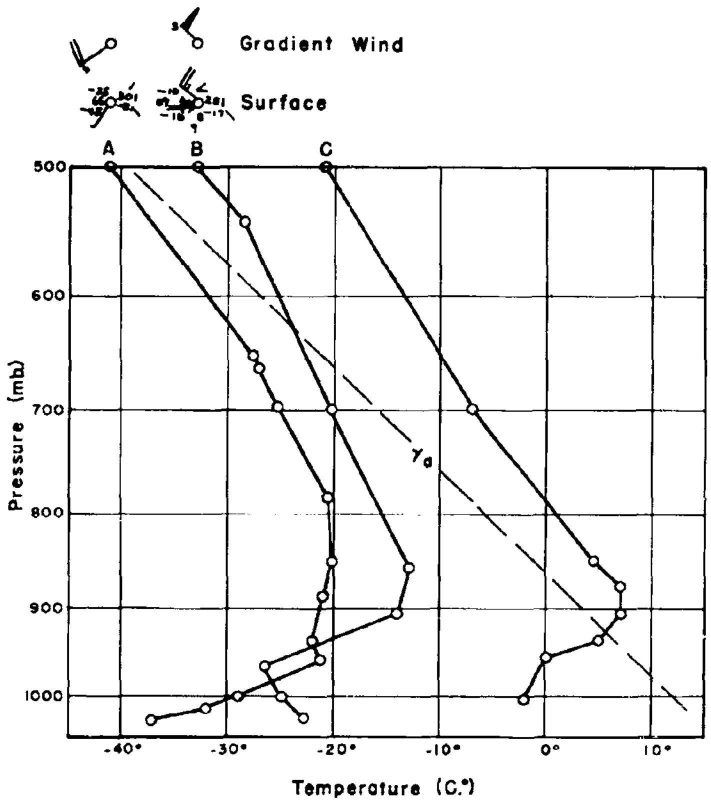



Figure 3.23. Examples of Different Types of Arctic Inversions.

inversion is formed at the cloud top. The net result of cloud cover, then, is to weaken and elevate the inversion.

Under stable conditions turbulent mixing gives rise to a downward flux of heat, the magnitude of which increases with increasing wind speed. Thus, if the surface wind were to strengthen, in case A of figure 3.23, more heat would flow towards the surface, warming the lower layers at the expense of those higher up. In this way a positive lapse rate of temperature would develop near the ground, and if the wind were sufficiently strong the lapse rate would approach the dry adiabatic as a limit.

This effect of mixing by the wind is well illustrated by sounding B of figure 3.23. The lower layer, or "mixing layer" in this case, is about 1,500 feet thick and is surmounted by a well marked inversion layer. The strong winds (25 knots at the surface, 49 knots at gradient level) have established an adiabatic lapse rate next to the surface and are causing snow to blow and drift.

Although difficult to measure, vertical motions can play a significant role in forming and destroying inversions. Downward motion, or subsidence, warms the air in the free atmosphere relative to the air next to the ground, which is constrained to move parallel to the earth's surface. The relative warming aloft can strengthen an existing inversion or tend to produce inversion conditions where none existed previously. Upward motion has the opposite effect. These effects of vertical motion explain, in part, the fact that inversions are generally

more intense in anticyclones than in cyclones.

Since wind and clouds are strongly influenced by local topography, it is apparent that irregular variations in the form and strength of inversions may occur on a synoptic scale. It is often difficult, therefore, in analyzing cross sections to maintain continuity on inversion structure from one station to the next, especially if stations are far apart and subject to different geographical influences. However, at any one spot the behavior of the inversion can be explained in terms of the factors discussed above, and provided that these are forecast correctly, the local behavior can be predicted.

### 3.8 Cyclone and Anticyclone Behavior

In this section data on the frequency of occurrence of cyclones and anticyclones will be used in conjunction with preceding information on frontal frequencies and mean circulation patterns, in order to arrive at a description of the behavior of synoptic systems in winter and summer.

#### 3.8.1 Winter

From figure 3.24 it is seen that cyclones are most frequent in the region between Greenland and Iceland, giving rise to the so-called Icelandic low. The corresponding feature in the Pacific, the Aleutian low, lies just off the region of the map. The area of high frequency has two extensions, one northeastward across the Norwegian Sea and a second northward into Baffin Bay. Secondary maxima of cyclone frequency lie in each extension. Other minor areas of above average occurrence of lows are found north of Greenland and over the Queen Elizabeth Islands. These latter areas may be fictitious, reflecting a tendency for analysts to prematurely drop lows which move along the northern shore of Siberia from their analyses and then to reanalyze them when they appear near Ice Island T-3 or stations in the Queen Elizabeth Islands.

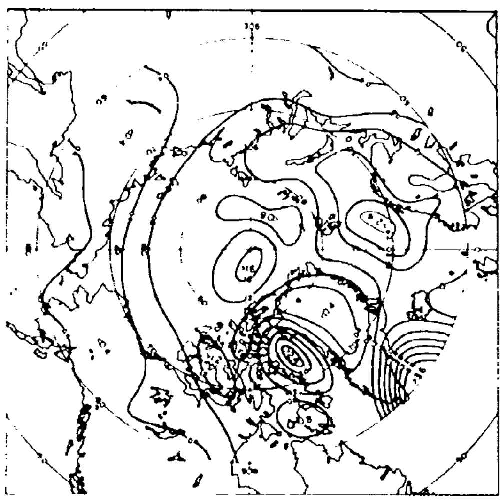


Figure 3.24. Percent Frequency of Cyclones in Winter North of 60°N, per 100,000 Square Miles.

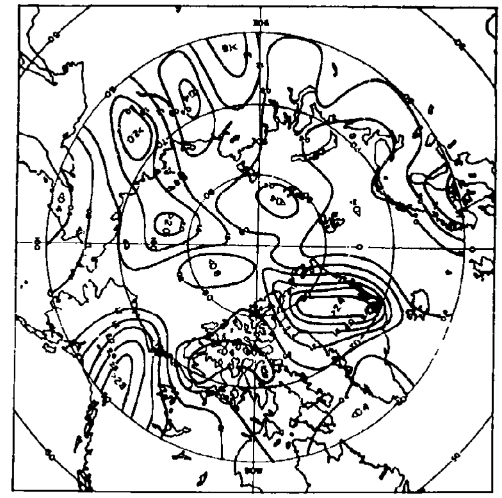


Figure 3.25. Percent Frequency of Anticyclones in Winter North of 60°N, per 100,000 Square Miles.

In general, cyclone frequencies are low in the areas where anticyclones are predominant; figure 3.25. These are seen to be the regions of the Siberian and North American highs, and Greenland. Because of the large reductions to sea level, the computed pressure values over Greenland may be questionable; therefore, some suspicion arises regarding the significance of the statistics in this vicinity.

The main features of cyclone behavior in winter may be summarized with the help of figure 3.24. Major storm tracks are found in the Pacific, carrying storms into the Aleutians and the Gulf of Alaska where they fill or weaken. Many Pacific storms regenerate east of the Rockies, and another major storm track leads from this area to the vicinity of Iceland where it converges with a major track from the east coast of the United States. Many storms with occluded or cold low structures stagnate and fill off the southern tip of Greenland. However, some

drift up the east coast and occasionally enter the Arctic Basin. Others rejuvenate between Iceland and Spitzbergen, leading to a secondary region of high cyclone frequency and to a major storm track which passes across Novaya Zemlya into northern Siberia.

Although many storms die out along the Siberian coast, others continue across the Arctic Sea and approach the Queen Elizabeth Islands from the west. When strong blocking highs are present over Alaska, storms from the Pacific enter the Arctic Basin and generally move into the islands of northern Canada.

Baffin Bay is a collecting area for many types of lows. Some arrive by way of the Queen Elizabeth Islands. Others approach from the North American Continent. Still others form as a result of lee cyclogenesis when the circulation around deep cyclones near the southern tip of Greenland brings an easterly flow across the icecap.

Minor storm tracks are present over Europe and western Siberia. These lead into the minor upper-level trough over the Kara Sea where filling generally occurs.

3.8.2 Summer

As in winter, cyclones are especially frequent off the southern tip of Greenland and over Baffin Bay (fig. 3.26). Important differences between cyclone behavior in the two seasons are

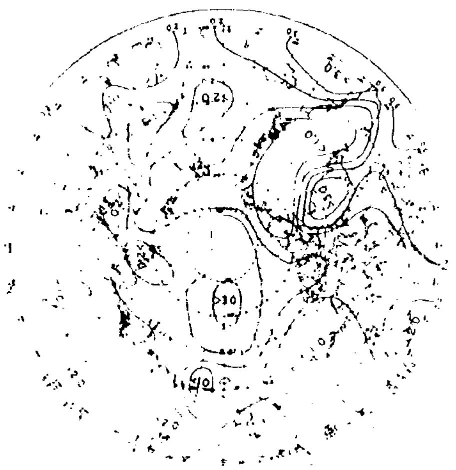


Figure 3.26. Percentage Frequency of Occurrence of Cyclone Centers in Squares of 100,000 Square Kilometers, July through September.

Figure 3.27. Percentage Frequency of Occurrence of Anticyclone Centers in Squares of 100,000 Square Kilometers, July through September.

noted, however, over Siberia and in the center of the Arctic Ocean where cyclones occur much more commonly in summer. In fact, it appears that a region comparable in cyclone frequency to the Icelandic area exists over the interior of the pack ice. Other areas of relatively large cyclone frequency appear over the Kara Sea, off the Norwegian coast, and in interior Canada.

Cyclones are infrequent over Greenland, the Beaufort Sea, the northern tip of Norway, and parts of Siberia.

According to figure 3.27 anticyclones predominate in Greenland and in the western portion of the Queen Elizabeth Islands, and to a lesser degree over the Kara Sea. If latitudinal averages are taken, anticyclones are found to be most frequent in the belt about 75° N. Cyclones, on the other hand, have latitudinal maxima near 60° N. and the pole.

The main features of cyclone behavior in summer may be summarized with the help of figure 3.26. As in winter major storm tracks converge in the Iceland area where some stagnation and filling occurs. The continuation of this storm track, however, is further south in summer, lying across southern Scandinavia and northern Russia.

The development of frontal or baroclinic waves over Siberia gives rise to major and minor storm tracks which terminate in the area of low

pressure and high cyclone frequency in the central Arctic. Some of the storms in this area continue to migrate eastward to the Queen Elizabeth Islands. Also, storms from the Pacific occasionally move through Bering Strait and along the northern shores of Alaska and Canada.

A major storm track ends near the coast of southeastern Alaska and recommences in the lee of the Rockies. Storms which follow this track generally come to rest in Baffin Bay where they are joined by storms from the United States and from the other regions mentioned previously. The result is a continuation of the high cyclone frequency in this area.

3.9 Stratospheric Disturbances

As a rule, the stratospheric flow pattern undergoes only small, slow changes. The tropospheric storms generally damp out in the lower stratosphere. Higher up in summer there exists an easterly flow which appears to behave independently of the tropospheric circulation. In winter the very large-scale disturbances of the tropospheric flow are still evident at the highest levels yet studied. However, these disturbances tend to be quasi-stationary and to vary little in intensity.

Despite the usual quiescence of the stratospheric flow patterns, it has been found in recent years that one or several times each winter an upheaval of the first magnitude may occur in the circulation at polar and subpolar latitudes. The most striking feature of the stratospheric disturbances is the drastic temperature rise which accompanies the changes in flow, so that the

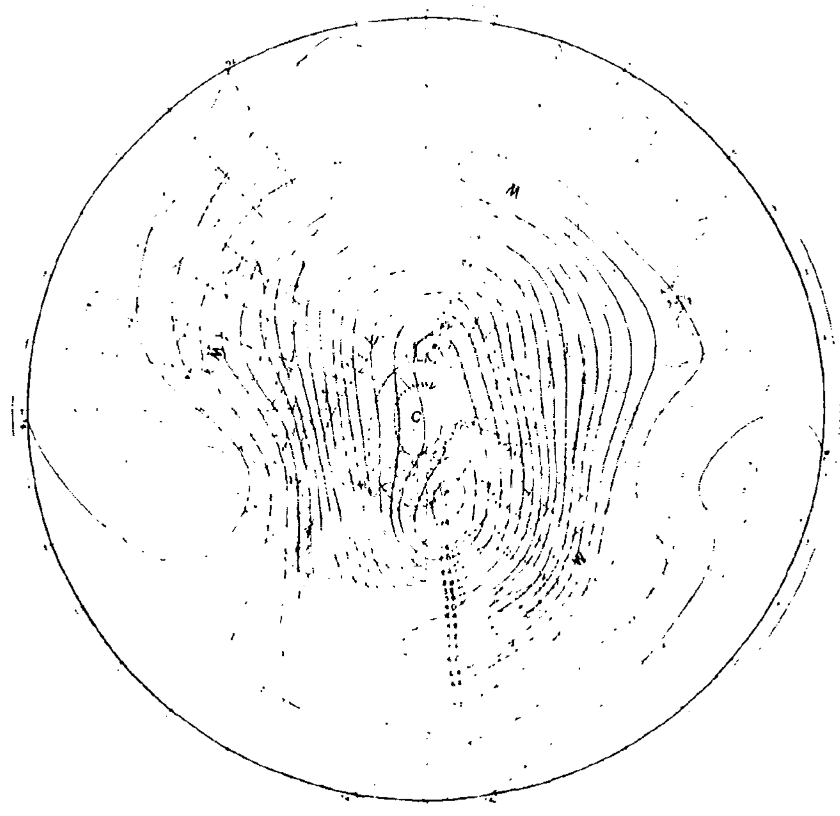



Figure 3.28. Stratospheric Circulation on 25 January 1957.

phenomenon in question is usually referred to as a sudden or "explosive warming."

An example of a sudden warming is contained in figures 3.28 through 3.30. At the beginning date, 25 January 1957, the stratospheric circulation was nearly normal for the season. Major troughs were present over North America and eastern Asia. A minor trough appeared over Europe. The ridge in the Alaskan region was also prominently developed. The temperature field displayed the typical warm pocket near Kamchatka and the pool of cold air about the pole. Other areas of warmth were present off Newfoundland and in Asia.

In the ensuing 10 days drastic changes of temperature and circulation took place. On 4 February (fig. 3.29) troughs and ridges were greatly amplified and formed a two-wave pattern. Closed lows had appeared in each trough, the pool of cold air over the pole had split and migrated to lower latitudes, and the main warm areas had intensified and moved northward to opposite sides of the pole.

The trends noted during the previous 10-day period continued during the next 5 days, and on 9 February (fig. 3.30) a truly remarkable transformation of the circulation was in evidence. Symmetrically placed pairs of highs and lows existed in the belt between 60° N. and 70° N. Warm temperatures had flooded the polar cap, values as high as -25° C. and warmer being recorded near the pole. The cold pools continued their southward migration.

At the end of the warming depicted here

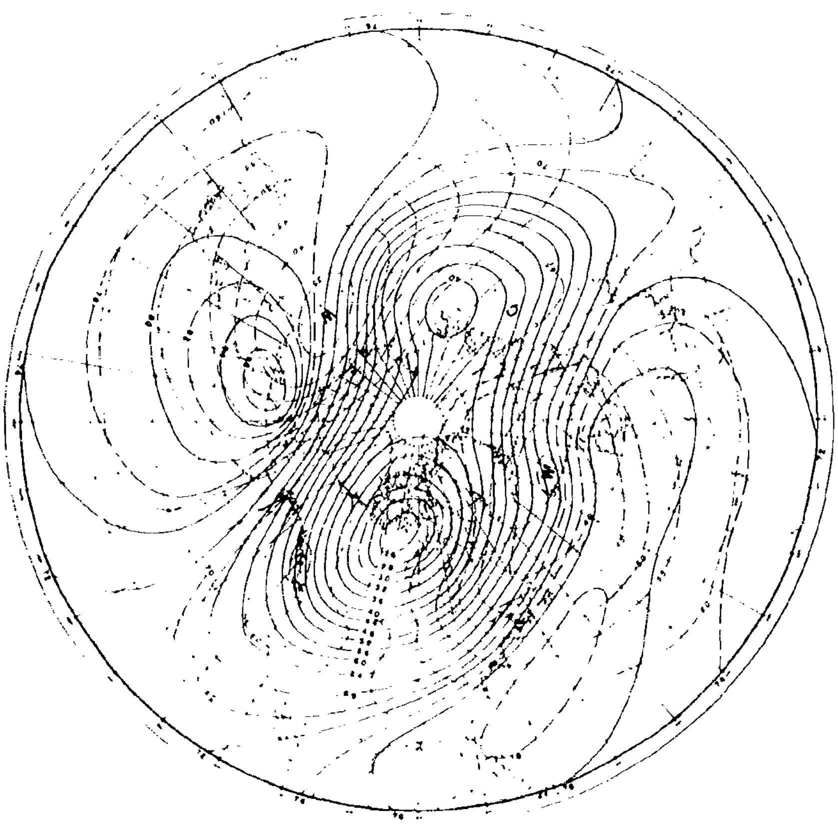


Figure 3.29. Stratospheric Circulation on 4 February 1957.

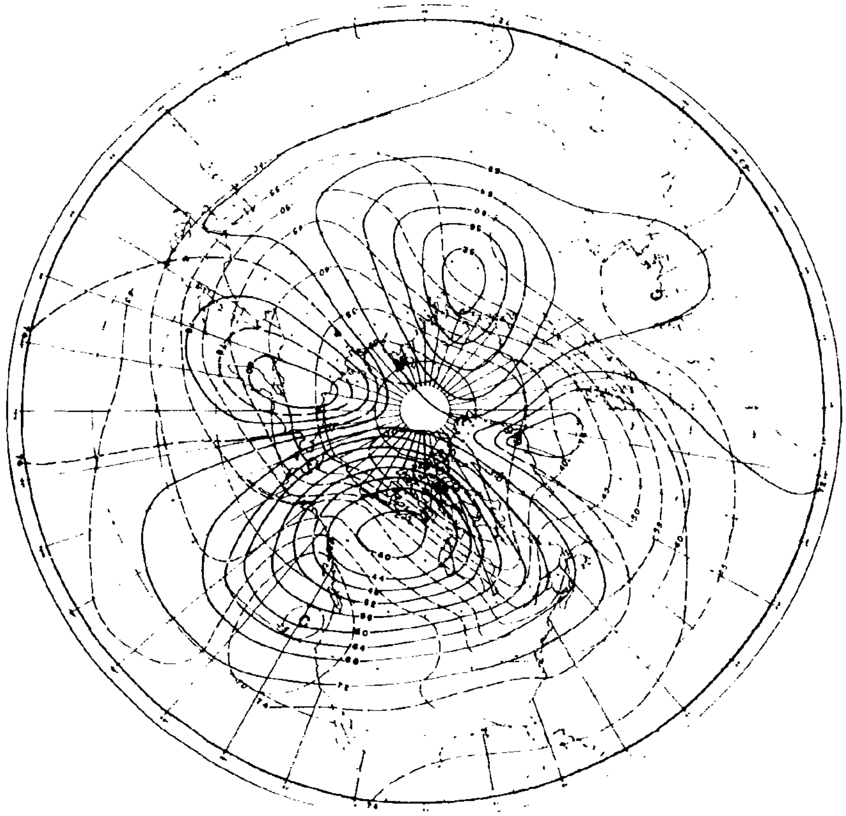


Figure 3.30. Stratospheric Circulation on 9 February 1957.

generally high pressures and warm temperatures prevailed over the pole, the strong westerlies of the winter stratosphere having been replaced by the easterlies characteristic of the summer season, in a period of less than a month. During March, however, the pattern reverted to its earlier state, though much reduced in intensity.

The behavior of the sudden warmings varies from year to year. Some years seem to be characterized by several minor warmings during the winter followed by a gradual transition to the summer pattern during a period of several months. In other years a major warming takes place as early as January and much of the change in circulation is accomplished at that time. It is doubtful, though, that a complete changeover ever occurs that early.

The origin of the sudden warming phenomenon is still a mystery. It was first believed to be caused by outbursts of solar corpuscular or ultraviolet radiation, but this view is now pretty much discredited. The fact that the warming generally begins at the highest levels of observation and is most intense at these levels is regarded by many as evidence that the disturbance begins high in the stratosphere and propagates downward.

Whether significant disturbances other than the sudden warmings occur in the stratosphere is a matter of question. Some authors believe that the intense temperature gradients associated with the polar night vortex and stratospheric jet stream generate baroclinic disturbances in the stratosphere which are distinct from those in the troposphere. Definite confirmation of these moving, amplifying waves is awaited.

# 4. ARCTIC SYNOPTIC ANALYSIS

## 4.1 Preliminary Remarks

The previous chapter dealt with the structure of arctic weather systems. This chapter will be concerned with principles and procedures for analyzing these systems.

No special methods of analysis are used in or recommended for polar regions. The arctic atmosphere is most properly thought of as an extension of the extrapolar atmosphere, subject to the same dynamic and thermodynamic controls and amenable to the same methods of analysis. In this respect polar meteorology differs from tropical meteorology, which does treat of distinctive synoptic features and does make use of special methods of analysis.

Despite the basic similarities in the circulation patterns of high and middle latitudes, the neophyte in arctic meteorology is more apt to be impressed by the apparent differences between the weather phenomena of the two zones than by their similarities. The reasons for these differences are to be sought in differences in intensities and frequencies of the various synoptic features, in the modifications in air flow produced by geographical factors, and in the difficulties imposed by a sparse observational network.

Thus, the persistent low-level inversion introduces complications into surface analysis, as has already been discussed to some extent in the previous chapter. Moreover, old occluded cyclones or cold lows are more common in polar regions than young, thermally-asymmetric depressions. These older, decaying systems do not possess the sharp features of the maturing disturbances, and, therefore, may be difficult to delineate, especially in regions of sparse data.

On the geographical side, Greenland exerts profound influences on the air flow in its vicinity. Other topographical features which affect the behavior of pressure systems, but to a lesser degree, are the mountains of Alaska, eastern Siberia, the Canadian Archipelago, and Scandinavia.

The most formidable problem of polar analysis stems from the deficiency of the observational network over the arctic seas. During the past decade, thanks to the maintenance of drifting ice stations and to regular weather reconnaissance flights, it has become possible to perform synoptic analyses on a daily basis for the entire Polar Basin. However, the analyses are often subject to considerable uncertainty, and the analyst must take more than ordinary pains to insure that his solution in a given situation is the best possible. In this respect, polar analysis has much in common with analysis over other oceanic areas where synoptic reports are scarce.

Procedures for analyzing in areas of sparse data or "silent areas" will be discussed in subsequent sections. These procedures are based on the following guiding principles:

(1) Maximum use of available data. The arctic analyst must exercise unusual care in compiling data. He must acquire a knowledge of all reports that are to be expected from remote areas and must constantly check to see that they have been received and plotted.

(2) Careful time continuity. History must be maintained on highs, lows, and fronts, and reasonable extrapolations must be made when these features move from areas where data are plentiful to areas where they are scanty.

(3) Systematic checks for vertical consistency. These may be accomplished by various methods of differential analysis. A general consideration of the subject will not be entered into here, although one "buildup" technique which has been used successfully over the Polar Basin will be explained and illustrated in the next section.

(4) Knowledge of the synoptic climatology of the area. The analyst can analyze more imaginatively if he has some knowledge of typical synoptic behavior in the region of interest. As a substitute for experience, the newcomer to an area can familiarize himself with published material on cyclone and anticyclone frequencies, storm tracks, etc. A summary of the behavior of synoptic systems in the Arctic is presented in section 3.8 of chapter 3.

In summarizing these preliminary remarks, it is noted again that arctic analysis does not pose any unique problems of a fundamental sort, but offers instead a host of lesser or secondary problems, many of them emanating from the lack of an adequate network of observing stations.

## 4.2 Surface Analysis

Over areas of the Arctic where observational

data are ample, isobaric analysis presents no special difficulties. Thus, our main concern here will be with analysis over the Arctic Ocean where the amount of data is quite limited. Figure 4.1 shows the present (1961) observing network north of 65 degrees. It is seen that the main sources of data beyond the continental limits are the various offshore islands, the drifting ice stations, and the aircraft reconnaissance flights. Surface synoptic reports are sent by the islands and drifting stations every 6 hours. A number of these stations take twice-daily upper air observations as well.

Reconnaissance flights are made once a day. The usual flight plan involves a first leg at the 700-mb. level from near Point Barrow northward to about 86° N., a second leg at the 500-mb. level from there to about 79° N. 127° W., and a final leg also at 500 mb. from the latter point to Barter Island. Temperature, humidity, and wind measurements are made at intervals of 150 miles. Dropsondes are released at five predetermined spots en route.

In addition to operating two drifting ice stations (usually) from which both surface and upper-air reports are sent, Russian scientists also make occasional short-term landings on the ice pack. The analyst must maintain a constant lookout for the weather reports from these more temporary locations.

Once all available data are plotted, the analyst should mark the previous positions of highs, lows, fronts, and non-frontal troughs on his chart and should mentally, at least, estimate the probable locations of these features on the current chart. The estimates may be based on either kinematic or dynamic extrapolation techniques.

A useful rule in extrapolating surface low centers is that to a first approximation they move parallel to the flow at 500 mb. and at one-half the 500-mb. wind speed. The same rule can be applied to the segments which compose a surface cold or occluded front. When the pressure system is nearly vertical, the rule does not hold well since the steering field above the low center changes direction too rapidly. In such cases the movement of the surface system is governed largely by the movement of the low aloft. A simple method for displacing upper lows is given in the following section on upper-air analysis.

The next step in the analysis is to fit the estimated picture of the situation to the few available reports. Unless obvious errors are present, precise use should be made of reported winds and pressure in drawing isobars over the Arctic Ocean. Because of the flat and uniform surface, winds are highly representative, and, therefore, can be used to good advantage in orienting and spacing the isobars. On the average, the surface wind makes an angle of 20° to 40° with the isobars and blows at 40 to 60 percent of the gradient wind speed. Variations in wind speed and stability may alter the above figures somewhat but are not sufficiently important to be taken into account in routine analysis.

At coastal and inland locations winds must be used with discretion in isobaric analysis because of terrain effects. In mountainous regions fictitious pressure gradients appear which make isobaric analysis difficult (or meaningless). These are due in part to the pressure reduction to sea level and in part to the action of mountain barriers in building and maintaining large pressure differences between low-level stations on opposite sides. The significance of sea level isobars a mile or more below the surface of the Greenland plateau has often been questioned. In this and other areas of high topography, the analyst should put greater stress on upper-air charts in specifying the flow pattern, especially the chart for the first pressure surface above the general ridge level.

Surface frontal analysis is more difficult in the Arctic than at lower latitudes. A first reason for this is that, on the average, fronts tend to be weaker at high latitudes. The thermal contrasts are not generally so strong, and many of the fronts are old occlusions undergoing frontolysis. A second difficulty has been mentioned earlier. It has to do with the poor association between surface temperature changes and temperature changes aloft when inversion conditions are present.

Time sections or continuous records of pressure, temperature, wind, sky and weather conditions, and precipitation amount are helpful in detecting frontal passages at isolated stations. By considering all elements the analyst can compensate for the indecisive nature of the temperature record. The analyst at a weather center does not, of course, have immediate access to continuous records from ice stations, but keeping a running record of the 6-hourly synoptic observations offers a worthwhile substitute.

A log of 6-hourly reports proves helpful in other ways. When maps are analyzed at only 12- or 24-hour intervals, the information from the intermediate hours allows the analyst to determine more accurately the time of passage of

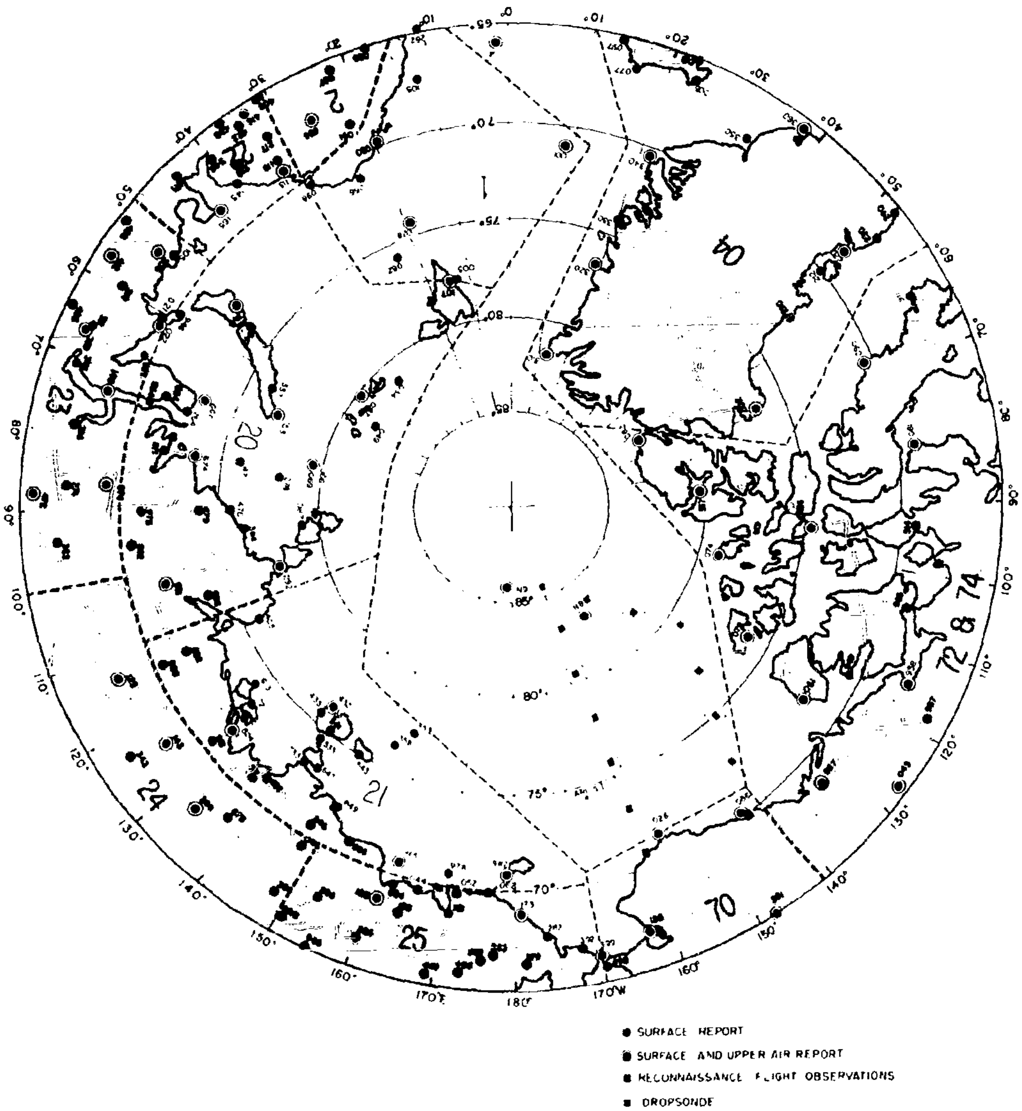



Figure 4.1. Arctic Weather Observing Network (1961).

significant features, such as troughs and ridge-lines, and the intensity at the time of passage. When reports are missing at standard map hours--as happens all too often in practice--the 6-hourly reports provide a more recent basis for extrapolation.

## 4.3 Upper-Level Analysis

As in the case of surface analysis, good upper-air analysis over the Arctic rests on maximum use of the available data, on the maintenance of careful time continuity, and, to a lesser degree, on a knowledge of the synoptic climatology of the area. In addition, some method of differential analysis is required over portions of the Polar Basin so that upper-level contours may be thermally consistent with the surface isobars.

The present (1961) sources of upper-air data are shown in figure 4.1. North of 70° N. the number of observations is quite limited, and the analyst will have little difficulty familiarizing himself with the types, times, and locations of reports.

Although the use of differential analysis will usually assure that important features of the upper-level flow pattern are delineated and placed in proper relationship to surface pressure systems, it is, nevertheless, advisable to maintain separate time continuity for the higher levels as well. Previous positions of high and low centers and troughs and ridges should be marked on the current chart, and the analyzed positions of these features should represent reasonable extrapolations from the earlier state.

A number of methods which are superior in accuracy to simple extrapolation, are available for estimating the displacement of 500-mb. troughs and ridges. A method which is quick and easy to apply and which gives useful results is the grid or box method. This is one of several methods which displaces systems according to the mean wind field, a procedure which finds theoretical justification in the Fjortoft method.

A box 20 degrees square has been employed successfully in displacing both closed lows at high latitudes, and troughs and lows at middle latitudes. An example of such a box, oriented so as to give the trough displacement at point O, is shown in figure 4.2. For a polar stereographic projection the 24-hour displacement in degrees of latitude is computed from the following formula:

$$D = \frac{1}{4}\left[(Z_2 - Z_1) + (Z_4 - Z_3) + (Z_6 - Z_5)\right]$$

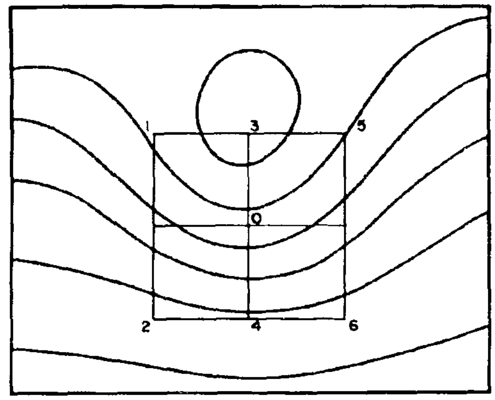



Figure 4.2. Illustration of Box Method of Displacing 500-mb. Trough.

where the Z's are heights in hundreds of feet at the appropriate grid points. In displacing lows, it is necessary to measure the component displacements in the east-west and north-south directions and to sum these vectorially.

The contours in figures 4.3 and 4.4 give some information regarding the climatology of upper-level systems. In winter it is seen that the lows tend to collect mainly in two areas--in the large troughs west of Greenland and over northeast Siberia. A third trough of lesser importance is located near Novaya Zemlya. Ridges predominate in the vicinity of Alaska and Scandinavia. Areas of mean troughs and ridges

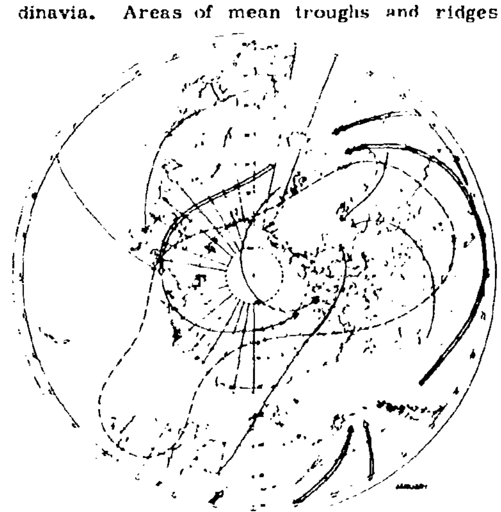


Figure 4.3. Schematic Diagram of Cyclone Behavior Over Polar Areas in January. Dashed Line Is Selected Mean Contour at 500 mb.

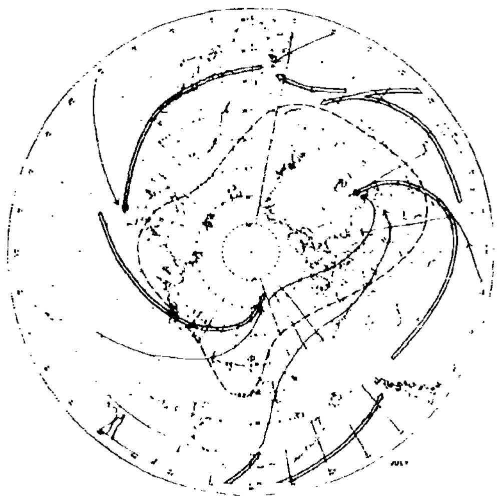


Figure 4.4. Schematic Diagram of Cyclone Behavior Over Polar Areas in July. Dashed Line Is Selected Mean Contour at 500 mb.

should be regarded as areas where daily troughs and ridges are especially intense or persistent, and it should be noted, for instance, that migratory lows may be rather common in the area where average conditions show a ridge.

From figure 4.4 it is seen that a four-wave pattern prevails in summer. The trough to the west of Greenland remains in about the winter position as does the minor trough near Novaya Zemlya. The second major trough migrates eastward to the Aleutians, and a new trough appears near Iceland. Again these troughs tend to be collecting areas for lows, though they fail to bring out the area of high cyclone frequency near the pole itself.

Vertical or thermal consistency may be achieved in a number of ways. Here we shall describe a method that has been tried and found useful in analysis over the Polar Basin. This method is based on statistical relationships between 700-mb. temperature and 1,000 to 500 mb. thickness, and provides a means of obtaining 500-mb. heights at selected points when the surface pressure and 700-mb. temperature are known.

To facilitate use of the method, nomograms have been prepared for each of the four seasons. The analyst first analyzes the surface isobars and the 700-mb. isotherms, taking care that the isotherms bear a reasonable relationship to the surface flow pattern and that they follow logically from the previous analysis. Only at the synoptic hour coinciding with the reconnaissance flight will it be possible to draw the isotherms with a high degree of accuracy. Once the analyses are completed, the analyst interpolates pressures and temperatures at various latitude and longitude intersections and enters the graphs with the interpolated values. The corresponding heights, as read from the slanting lines, are then plotted on the 500-mb. chart.

For a specified surface pressure and 700-mb. temperature, the standard error of estimate of 500-mb. height is approximately 70 feet irrespective of season. Since, however, in practical application the analyzed pressure and temperature may be somewhat in error, the analyst should allow himself a tolerance of about 100 to 200 feet in analyzing for the computed heights. The nomograms are presented in figures 4.5 through 4.8, and an example, based on the July nomogram, appears in figure 4.9.

Most of the remarks concerning upper-level analysis have dealt with the 500-mb. surface. This surface is generally regarded as the key upper level, and usually provides a good basis for interpolation downward and extrapolation upward. As a rule, the broad features of the 500-mb. pattern are still recognizable up to the 150- or 100-mb. levels. Higher in the stratosphere the flow pattern may change considerably. In winter, at 50 and 25 millibars, very pronounced large-scale disturbances develop over the polar regions which appear to be unrelated to the tropospheric flow. These are discussed further in section 3.9 of chapter 3.

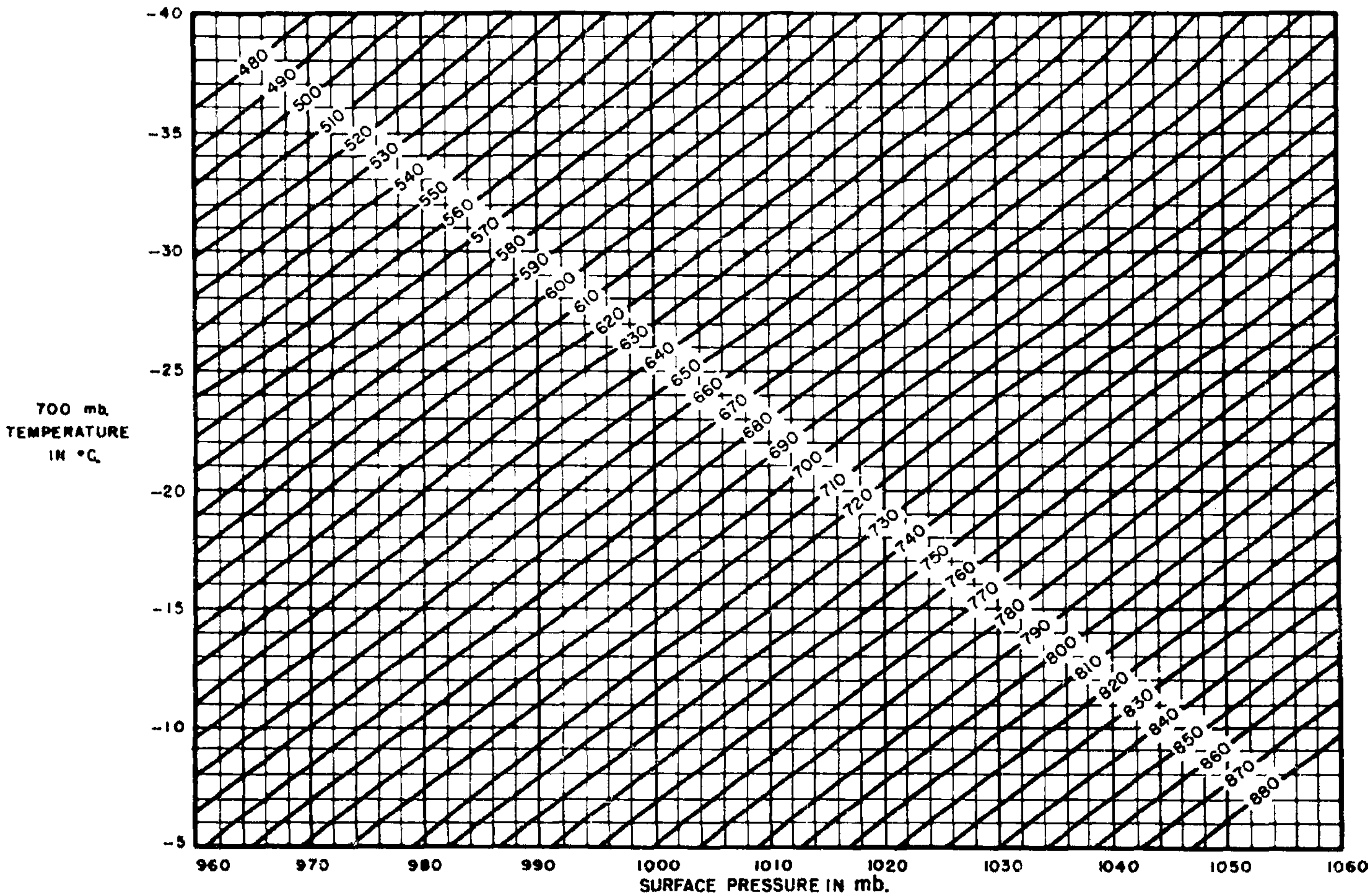



Figure 4.5. Nomogram Relating 500-mb. Height to Surface Pressure and 700-mb. Temperature -- January. (Heights in Feet with First and Last Digits Omitted.)

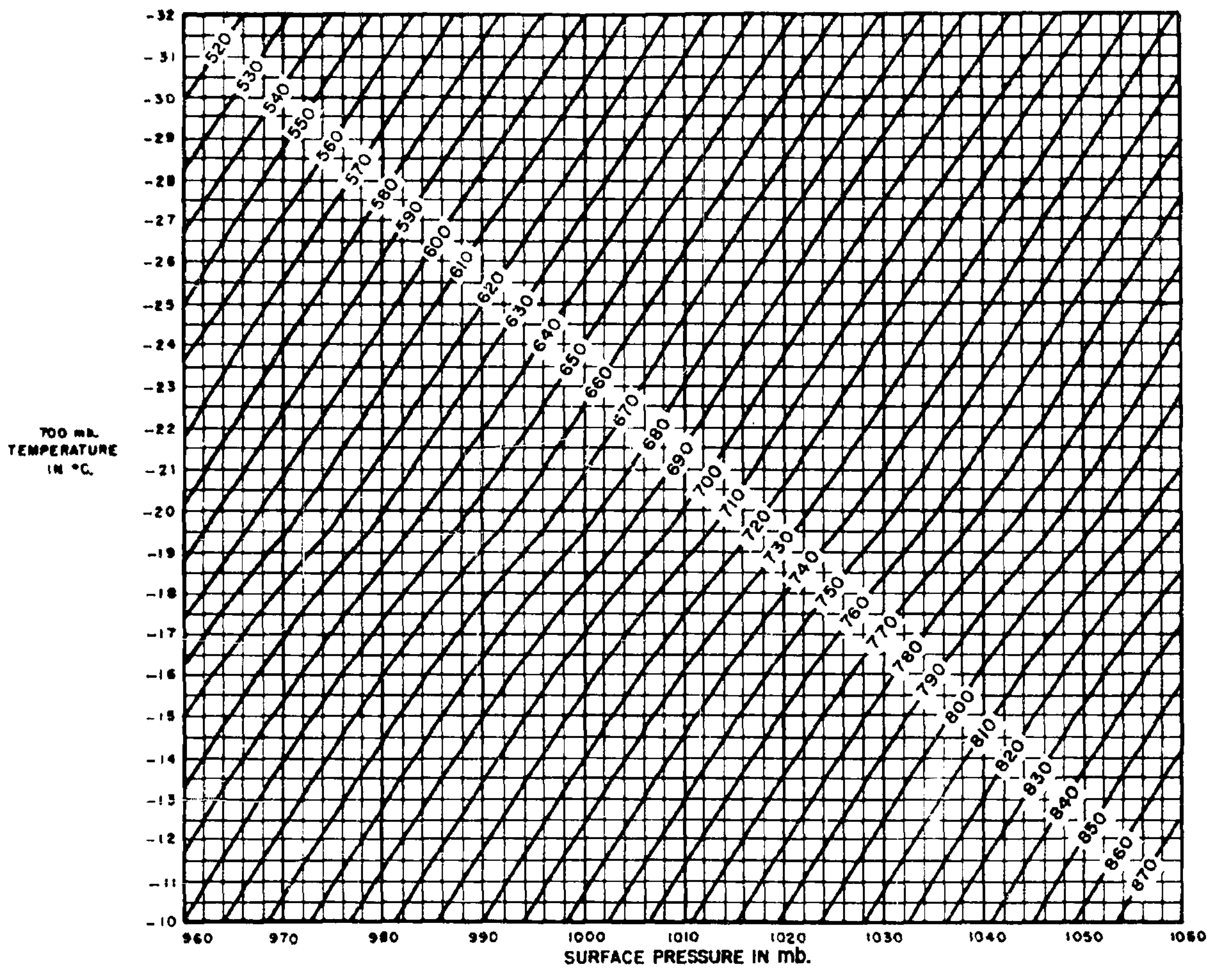



Figure 4.6. Nomogram Relating 500-mb. Height to Surface Pressure and 700-mb. Temperature -- April. (Heights in Feet with First and Last Digits Omitted.)

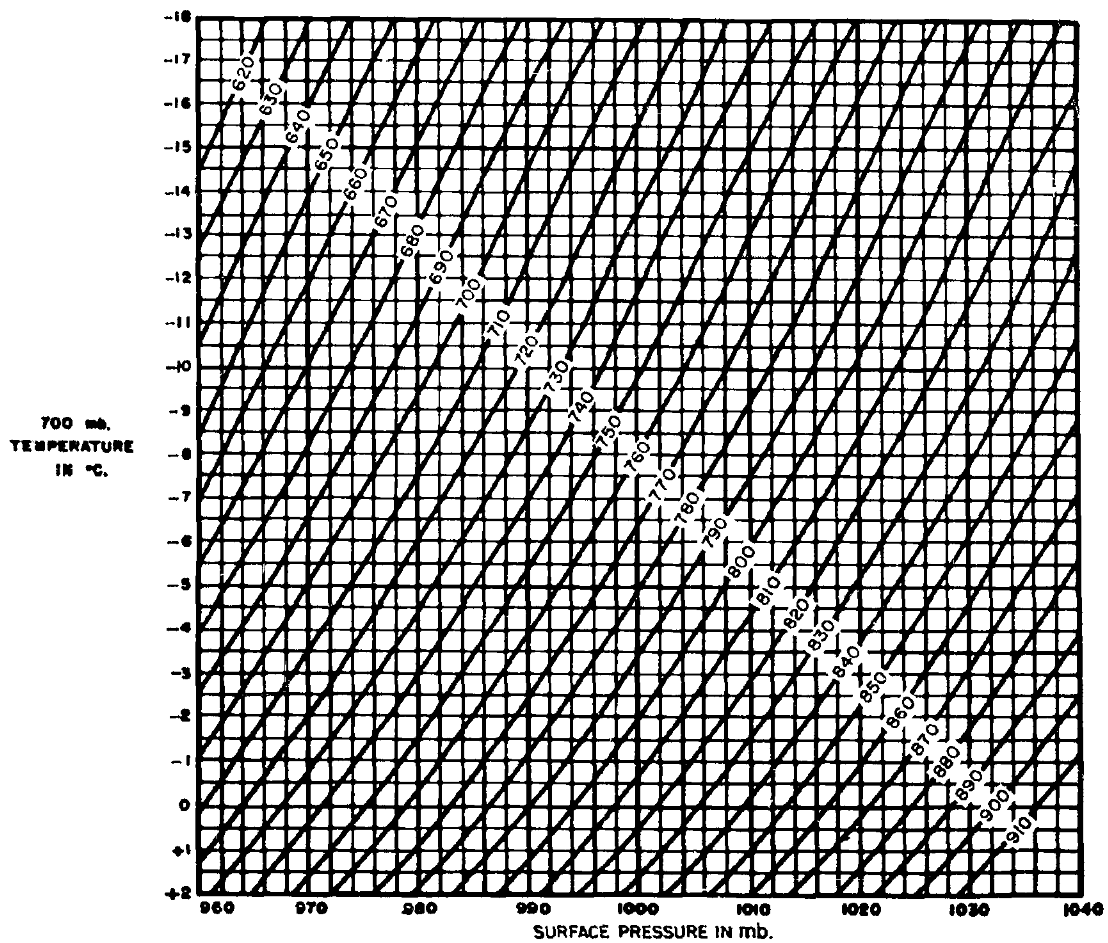



Figure 4.7. Nomogram Relating 500-mb. Height to Surface Pressure and 700-mb. Temperature -- July. (Heights in Feet with First and Last Digits Omitted.)

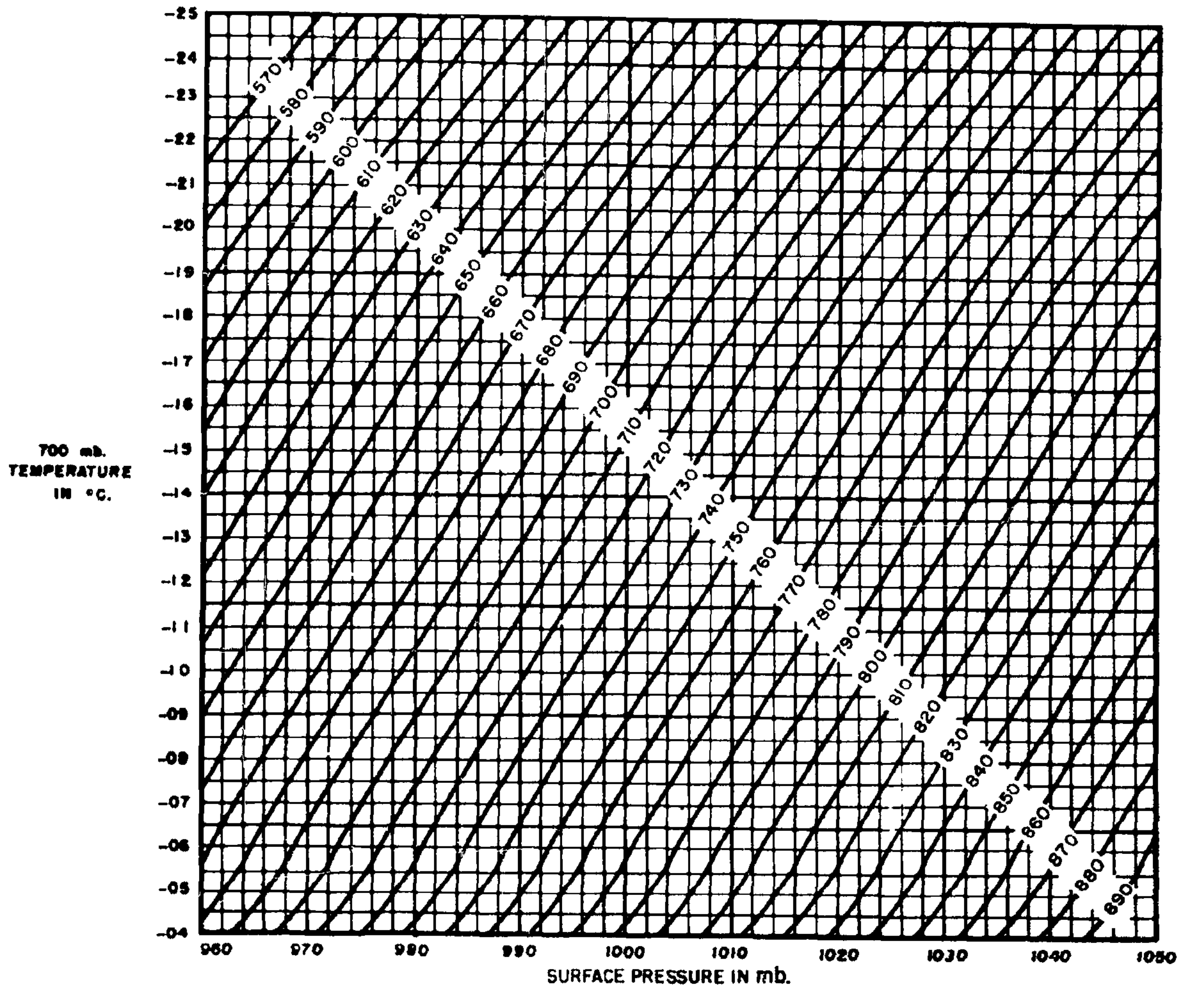



Figure 4.8. Nomogram Relating 500-mb. Height to Surface Pressure and 700-mb. Temperature -- October. (Heights in Feet with First and Last Digits Omitted.)

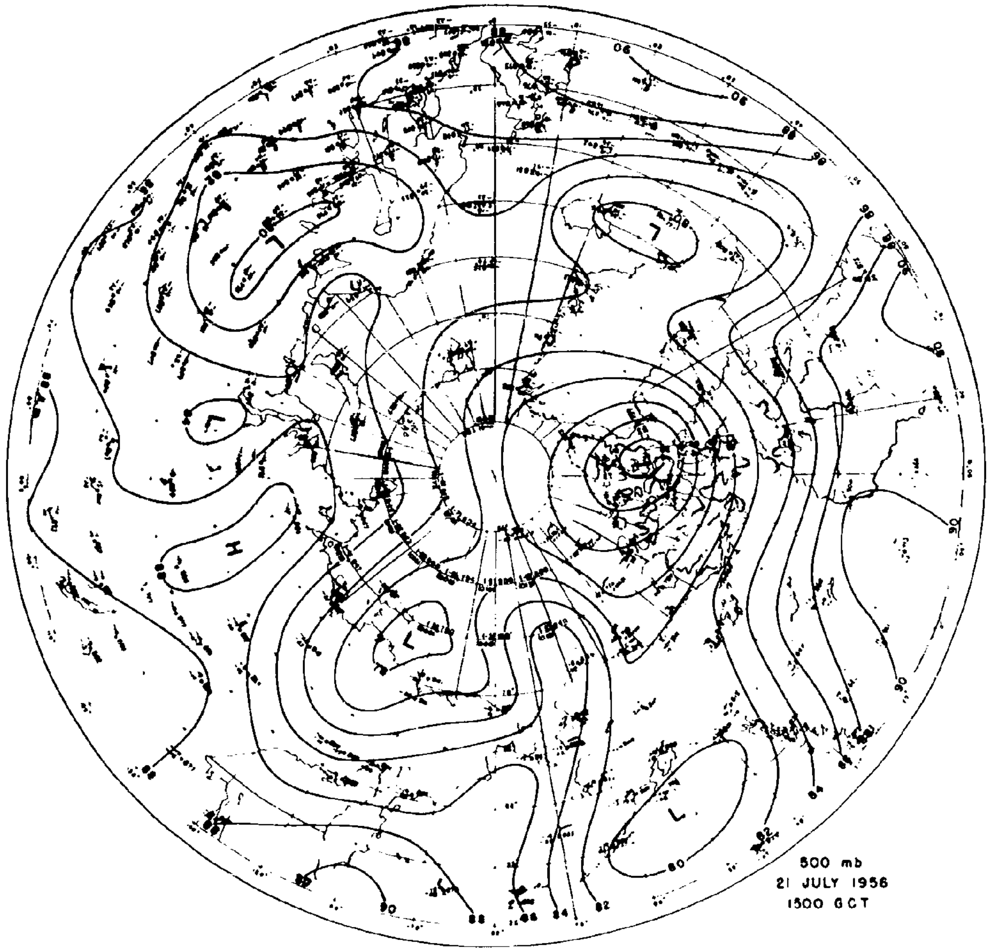



Figure 4.9. Example of Use of Nomogram. Plotting Arrangement - 700-mb. Temperature to Upper Left Point, Surface Pressure (Whole Millibars, Thousands Omitted) to Lower Left. 500-mb. Height as Determined from July Nomogram to Upper Right.

# 5. ARCTIC FORECASTING

Arctic forecasting, like arctic analysis, cannot be regarded as a fundamental branch of meteorological science, but must be viewed instead as an extension of the concepts and methods of middle latitude forecasting to the prediction problems of the Arctic. It is beyond the scope of the present work to review the wide variety of techniques that are currently in use in preparing prognostic charts and in forecasting the associated weather conditions. It is sufficient to say, that in the discussion which follows it will be assumed that the forecaster has available to him--by facsimile, teletype, or his own devices--a prognostic chart, and that this chart provides him with the information he requires regarding the behavior of the large-scale flow pattern. A review of methods for constructing prognostic charts may be found in Volumes I and II of Petterssen [8].

It will also be assumed that the forecaster is familiar with the techniques used in deriving objective forecast aids of the type used in local forecast studies. A description of these techniques is contained in Volume II of Petterssen [8].

Although the principles and methods used in the treatment of arctic forecast problems are in no way peculiar to the polar regions, the problems themselves are in many cases unique, or nearly unique. Ice fog and blowing snow may be cited as examples of phenomena which constitute important forecast problems at high latitudes but which are of only minor importance elsewhere.

Primary emphasis here will be on those phenomena which are more or less peculiar to high latitudes. Lesser attention will be devoted to other aspects of the forecast problem which are perhaps slightly altered under arctic conditions but which are otherwise the same as at middle latitudes.

An attempt has been made to include in the discussion all meteorological elements and phenomena which are of importance to the arctic forecaster. In each case the significant empirical and physical facts connected with the element are first reviewed; following this, suggestions are offered for its prediction. Where objective forecast aids are known and available to the writer, these are described in considerable detail under the appropriate headings.

One cannot read this section without realizing that much remains to be done in the field of arctic forecasting. It is hoped that the emphasis here on physical understanding of the various phenomena will provide the forecaster with the background for fresh attacks on the more pressing problems.

## 5.1 Wind

### 5.1.1 General Remarks on Wind Types

In discussing the problem of wind forecasting it is important to distinguish between winds which are part of the large-scale circulation patterns, as depicted on the synoptic chart, and winds which are of local origin. These latter winds, or local winds as they are called, are connected with features of the local topography and may be divided into two main categories according to their mode of formation.

A first type of local wind arises from differential heating or cooling of the air on a restricted scale. Examples of winds of this type are sea and land breezes, mountain and valley breezes, and glacier winds.

Such winds are much affected by the general wind in the area, in some instances being reinforced by the prevailing wind, in other cases being counteracted or completely suppressed.

The second main type of local wind may be thought of as a disturbance in the large-scale flow brought about by some feature of the local topography through mechanical action rather than through thermal action, as in the previous type. The thermally-induced wind is seen in its purest form when the pressure gradient is flat and the general wind is light or absent. The mechanically-induced wind cannot exist without a general wind and is most pronounced when the pressure gradient is large and the wind strong.

The most familiar example of a local wind of the mechanical type is the foehn, a warm, dry wind which occurs when the large-scale pressure gradient forces air to cross a mountain range and to descend on the lee side. The foehn is often marked by gustiness and may on occasions attain gale or hurricane intensity.

In areas where a mountain range or high plateau borders a coast, it is possible during the

cold season for the foehn air to arrive at the coast with temperatures lower than those previously existing despite the adiabatic warming. A foehn of this type is often called a bora, deriving its name from a violent wind that occasionally strikes the Dalmatian coast in winter. The gusty character of the bora is sometimes cited as proof that the wind has a strong katabatic component, but the true or warm foehn often displays this same feature so that it is probably more correct to regard the bora-type wind as merely a cold foehn rather than as a katabatic wind[1] reinforced by the pressure gradient.

In addition to the foehn-type winds in which air is directed downslope by the large-scale pressure gradient, there are pronounced local wind effects of the mechanical type in which the air flow is essentially horizontal. These effects are noted in the vicinity of certain topographic features such as mountain passes, water channels, and coastal bluffs and are referred to by a variety of terms--venturi, jet, funneling, channeling, and barrier effects, to mention a few.

### 5.1.2 Strong Surface Winds

By strong winds we shall mean sustained winds of 30 knots or more. This definition is admittedly arbitrary, but conforms fairly closely to the criterion set up by other writers on the subject.

Wind studies at a large number of arctic sites reveal that high surface winds are invariably associated with well-developed pressure gradients. Very often the observed speeds can be accounted for entirely by the pressure gradient, while at other times a local enhancement occurs which must be attributed to some effect of the local topography. This topographical influence is evidenced by the strong tendency for high winds to blow from a preferred direction at many sites. Thus, we may state the general rule that strong winds tend to occur when the pressure gradient is relatively pronounced and oriented in a preferred direction with respect to the local topography.

Because of the generally weak nature of the pure katabatic wind, that is, the cold downslope wind that forms during periods of flat pressure gradient, the extreme cases of local enhancement cannot be ascribed to thermal effects but must, instead, be explained dynamically. The fact that the extreme cases are characterized by abnormal warmth at most arctic stations strongly supports this view. Moreover, at some sites where unusual winds are noted there is not sufficient relief to cause a significant katabatic effect.

To give some idea of the variety of conditions under which strong winds occur, we shall review the typical circumstances at a number of specific locations.

(a) Narsarssuak, Greenland

This station lies at the head of a fjord on the south coast of Greenland. Strong winds occur with low pressure to the south or southwest of the station and high pressure to the north or northeast. The wind blows toward low pressure, descending from the icecap which rises to an elevation of 8,000 feet a short distance inland from the station. The wind speed is roughly proportional to the pressure gradient along the southeast coast and occasionally exceeds 100 knots. The wind is warm and dry and obviously belongs in the foehn catagory.

(b) Thule, Greenland

The station is located in a valley on the west coast of Greenland with the main body of the icecap lying to the east. To the southeast a tongue of ice about 2,500 feet in elevation protrudes outward from the icecap. The strong winds, almost without exception, blow from the southeasterly quadrant and are associated with relatively warm temperatures. These are due in part to turbulent mixing, in part to adiabatic descent from the ice arm, and in part to the advection of air which is of more southerly origin. A strong pressure gradient between Thule and Upernavik, 300 miles to the southeast, is a major factor in the occurrence of the "Thule Wind." From the description, it appears that this wind is at least partly foehn in character.

(c) Alert, Ellesmere Island

The station is located at the northeast tip of Ellesmere Island close to Robeson Channel, which divides this island from Greenland. Two mountain systems with permanent snow and ice fields lie well to the west and southwest of the station. The valley between these systems is oriented northeast-southwest, paralleling Robeson Channel, and is generally below 3,000 feet in elevation. The strong winds blow mostly from the southwest at times when the pressure gradient is strong and is oriented in such a way as to force the air to flow in this direction; these

[1] [illegible]

winds are relatively warm. A foehn effect is suggested, but the humidity is sometimes quite high both at the surface and aloft so that channeling and barrier effects must also be of importance.

(d) Barter Island

This small, flat island located immediately off the north coast of Alaska is occasionally visited by strong and persistent winds which blow parallel to the coast, either in an easterly or westerly direction. It has been convincingly demonstrated [1] that the abnormal wind speeds are caused by a funnelling of the air around a knob-like protrusion of the Brooks Range, located 50 miles to the south of the station.

(e) Juneau, Alaska

Juneau is located on the southeastern coast of Alaska on a narrow shelf that rises abruptly to an ice plateau of roughly 5,000 feet elevation. The strong winds that occur there are known as Takus, since they are popularly associated with the nearby Taku Glacier. These winds, however, are of the bora type, being always associated with a large pressure gradient either in conjunction with a deep low offshore or, in the more spectacular instances, with a powerful high over the inland plateau. The cold temperatures associated with the wind are due to the outflow of very cold air from the interior. In cases where the inland temperatures are less extreme, the adiabatic warming may be sufficient to produce a local warming and thus place the Taku in the foehn rather than the bora category.

(f) Big Delta, Alaska

Although strong arctic winds are characteristically associated with coastal localities, they may occur at inland stations as well, provided that the proper terrain features and meteorological conditions exist. Big Delta--located near the mouth of a long valley, oriented approximately west-northwest to east-southeast with mountains on either side--is an example of an inland site that experiences occasional outbursts of strong and gusty winds. The winds blow most commonly from the east-southeast (along the axis of the valley) and less frequently from south to southeast. They are invariably associated with a large pressure gradient in the direction of the valley and with relatively warm temperatures. The warming is most pronounced with a southerly wind from the mountains in which case the wind may be classified as a foehn. In the more common case of down-valley flow, the warming is due mostly to the destruction of the surface inversion by turbulent mixing, and the high velocities are attributed to a funnelling effect.

### 5.1.3 Two Examples of Strong Wind Occurrences

The first case which we shall use for illustration is a particularly dramatic one that took place at Alert, Ellesmere Island, on 4 February 1953. However, it is by no means an isolated case and displays many features in common with high-wind occurrences at other arctic sites.

Table 5.1 lists the successive 3-hourly surface observations surrounding the time of occurrence, and graphs of temperature, pressure, and wind speed appear in figure 5.1. From figure 5.1 one notes a steady, rapid drop in pressure and a gradual rise in temperature preceding the onset of the high winds. Following a long period of calm, the wind blows lightly at first, then rises abruptly to hurricane strength. Because of the 3-hourly interval, the exact rate of increase in not known but similar jumps are known to have occurred in periods of an hour or less.

With the onset of the gale winds the temperature shows an abrupt increase, and the pressure levels and begins to rise, at first slowly then more rapidly. A rapid decline in wind strength accompanies the sharp pressure rise and, except for occasional fitful increases, the big blow is over. The temperature remains relatively warm for many hours following the occurrence, then returns nearly to its earlier level.

From table 5.1 it is seen that only scattered high clouds were present immediately preceding the strong winds and that skies were completely clear following. At the time of occurrence the sky was obscured and the visibility was reduced to zero by blowing snow.

The synoptic pattern accompanying the strong wind situation is depicted in figures 5.2, 5.3, and 5.4. A deepening low passes to the west of the station, making its closest approach at approximately the time of greatest wind strength. Aloft the flow is from the south and quite strong. The observed peak surface wind is in excess of both the wind at 700 mb. and the gradient wind, but the latter is abnormally strong for the area and in roughly the same direction as the observed wind. The usual association between strong wind and pressure gradient is evident.

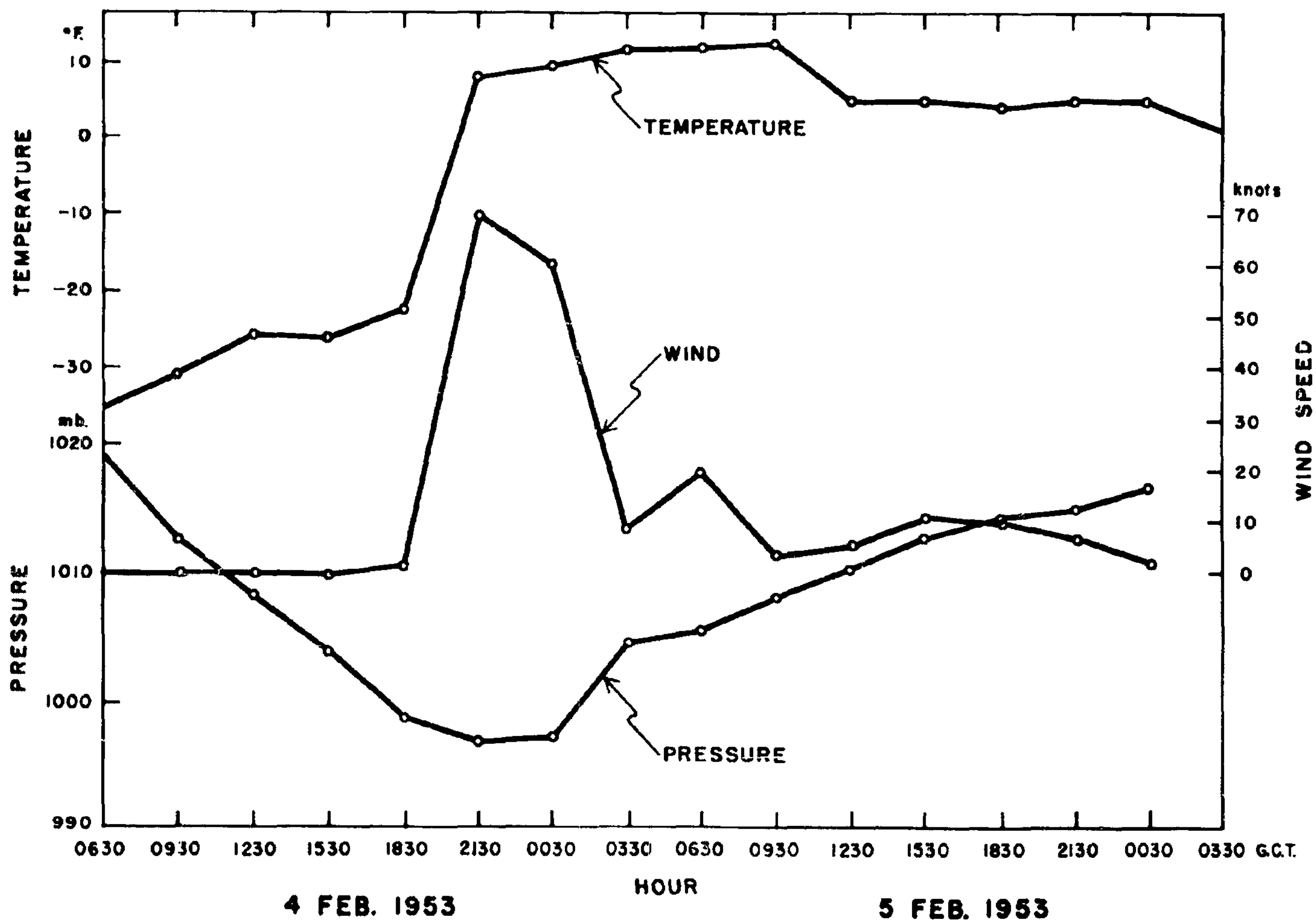



Figure 5.1. Wind, Pressure, and Temperature Traces for Alert, Ellesmere Island, 4 and 5 February 1953.

TABLE 5.1
Three-Hourly Synoptic Observations for Alert, Ellesmere Island, February 4, 5, and 6, 1953 [9].

| Date | Hour GCT | Sky Cover (Tenths) | Pressure (mb.) | Temperature (°F.) | Dew Point (°F.) | Wind Direction | Wind Speed (mph) | Ceiling (100's feet) | Visibility (Miles) | Present Weather |
|---|---|---|---|---|---|---|---|---|---|---|
| 4 | 0330 | 0 | 1020.2 | -35 | -41 | C | | UNL | 13 | 02 |
| 4 | 0630 | 10 | 1018.7 | -35 | -41 | C | | 090 | 10 | 03 |
| 4 | 0930 | 10 | 1012.6 | -31 | -37 | C | | 080 | 13 | 02 |
| 4 | 1230 | 4 | 1008.3 | -26 | -32 | C | | UNL | 10 | 76 |
| 4 | 1530 | 0 | 1004.0 | -27 | -32 | C | | UNL | 13 | 02 |
| 4 | 1830 | 3 | 998.9 | -22 | -27 | NE | 2 | UNL | 13 | 02 |
| 4 | 2130 | 10 | 997.1 | 9 | 3 | SW | 70 | 006 | 0 | 39 |
| 5 | 0030 | 10 | 997.6 | 10 | 5 | SW | 60 | 006 | 0 | 39 |
| 5 | 0330 | 0 | 1004.6 | 12 | 8 | S | 8 | UNL | 13 | 02 |
| 5 | 0630 | 0 | 1005.6 | 12 | 9 | WSW | 20 | UNL | 10 | 36 |
| 5 | 0930 | 0 | 1008.0 | 13 | 11 | S | 4 | UNL | 13 | 02 |
| 5 | 1230 | 0 | 1010.4 | 5 | -1 | SSW | 6 | UNL | 13 | 02 |
| 5 | 1530 | 6 | 1012.9 | 5 | 1 | NW | 12 | 080 | 13 | 03 |
| 5 | 1830 | 4 | 1014.2 | 4 | -1 | SW | 10 | UNL | 13 | 02 |
| 5 | 2130 | 0 | 1015.2 | 5 | 0 | N | 7 | UNL | 13 | 02 |
| 6 | 0030 | 0 | 1016.7 | 5 | -1 | S | 2 | UNL | 13 | 02 |
| 6 | 0330 | 0 | 1016.8 | 1 | -4 | SW | 8 | UNL | 13 | 02 |
| 6 | 0630 | 0 | 1016.9 | -14 | -19 | NE | 6 | UNL | 13 | 02 |

Fortunately, the high winds occurred between successive 12-hourly radiosonde observations so that a record of meteorological conditions aloft both before and after the time of occurrence is also available. The soundings in figure 5.5 show the characteristic surface inversion preceding the occurrence, in this case especially sharp because of the influx of unusually warm air aloft. Following the high winds, the inversion is greatly weakened due to a combination of cooling aloft and warming at the surface. From the surface pressure trace it may be surmised that the upper cooling began at about the time of the onset of the high winds. This cooling could be associated with an occlusion that appeared on earlier maps (not shown), but it is possible that the line of advance of the colder air was frontogenetical and therefore that it did not have logical continuity with the earlier front. Relatively high moisture values in the lower layers rule out the possibility of a significant foehn influence and suggest instead the importance of a funneling effect.

The coincidence of the high winds with the start of the surface pressure rise and the upper level cooling is not an uncommon feature of such situations and may be explained as follows. Prior to the outbreak of the winds at the surface, there is much evidence that they are generally present in the layers immediately above. Aircraft taking off or landing at this time often report extreme turbulence in the lower 2,000 to 4,000 feet, and sometimes snow from nearby hills is noted blowing overhead at high speeds despite the calm or light winds at the surface. In the case chosen here for illustration, the preceding rise in wind is suggested by the slow upward trend in surface temperature which in considerable part is probably due to an enhanced turbulent flux of heat

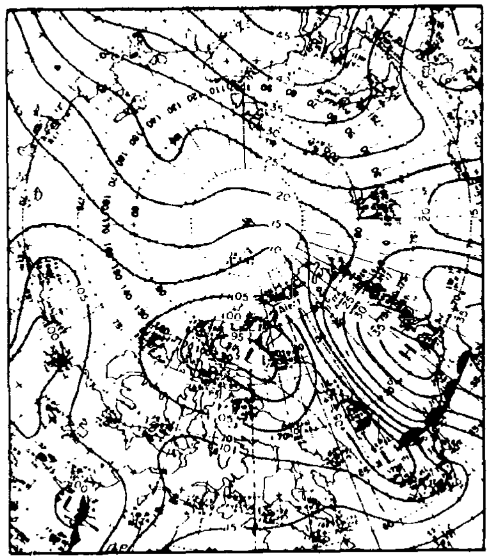


Figure 5.2. Surface Chart for 1230 G.C.T., 4 February 1953.

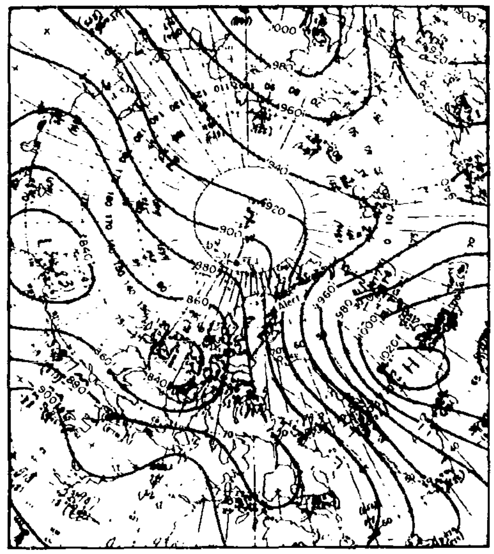


Figure 5.4. 700-mb. Chart for 1500 G.C.T., 4 February 1953.

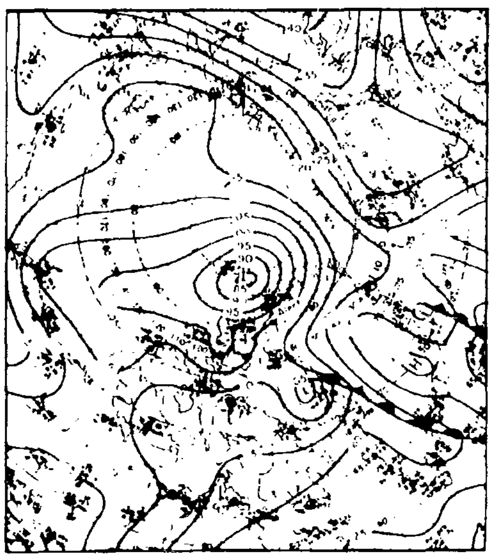


Figure 5.3. Surface Chart for 0030 G.C.T., 5 February 1953.

downward.

While warm advection continues aloft, the inversion remains strong and effectively seals off the strong winds, but with the onset of the cooling the lapse rate diminishes and the winds break through to the surface, often quite violently. Despite the overall cooling, the surface temperature rises sharply because of the downward mixing of potentially warmer air. The preceding sequence of events must not be supposed to be the only one possible. On occasions the winds break through to the surface before the upper cooling, and at other times the barometer rises for several hours before the winds strike.

The lack of a unique relationship between the onset of the winds and the upper cooling suggests that the foregoing explanation may be inadequate and that other views on the subject are to be encouraged. In hilly and mountainous regions it might prove advantageous to regard the high winds as a form of lee wave. The problem would then be treated in much the same manner as orographic turbulence, a subject which will be discussed in a later section.

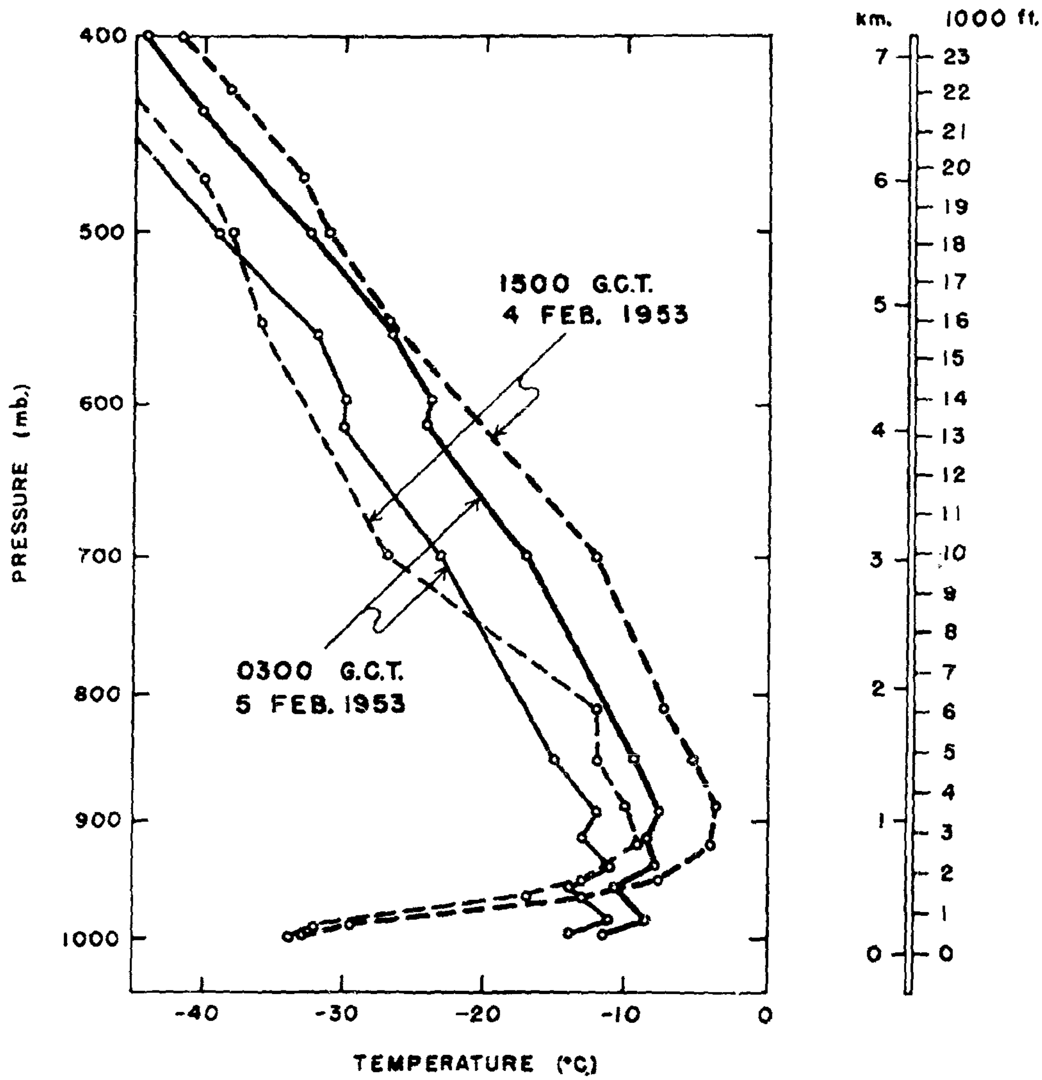



Figure 5.5. Soundings for Alert, Ellesmere Island.

The second case in illustration occurred at Barter Island, Alaska, on January 4 and 5, 1957. The wind, temperature, and pressure records for this case are shown in figure 5.6, and the surface synoptic charts shortly after commencement of the strong winds and at the time of peak velocities are presented in figures 5.7 and 5.8.

This situation occurred in conjunction with the passage of a deepening low pressure system and its associated cold front along the northern coast of Alaska. Prior to the passage of the cold front winds were light, easterly. Winds shifted to westerly behind the cold front and at first remained light. After several hours had elapsed, the wind speed increased rapidly, in part because of destabilization of the lapse rate by cooling aloft and in part because of an increase in pressure gradient which forced the air to strike the Brooks Range at a critical angle. Unlike the previous case the temperature declined even during the period of maximum winds, the cooling by advection evidently outweighing the warming by turbulent mixing.

The occurrence of abnormally strong winds at Barter Island has been successfully explained by Dickey [1] as a consequence of a funneling of the air currents about a prominence in the Brooks Range to the south. This funneling is most pronounced when the surface winds are parallel to the coast, in either direction, and causes the surface wind to be about 50 percent greater than the geostrophic wind and 60 to 70 percent greater than the undisturbed surface wind upstream. By treating the Brooks Range in the vicinity of the station as a cylindrical obstacle to the flow, Dickey has been able to re-

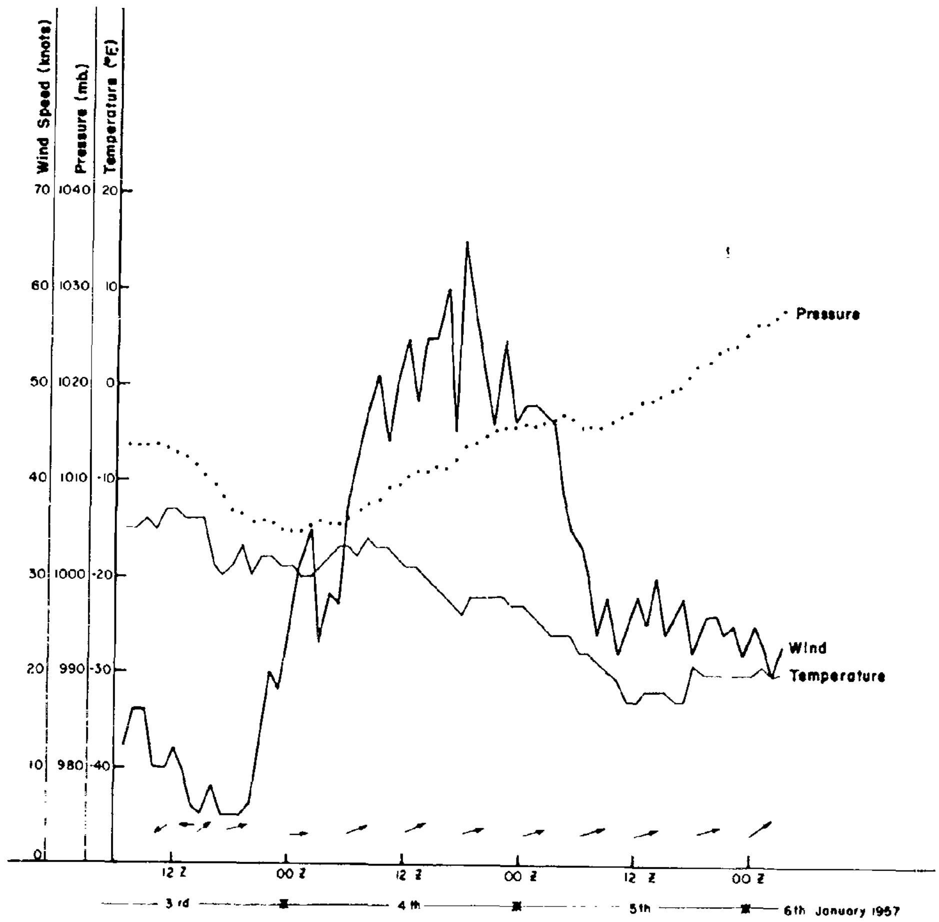



Figure 5.6. Time Graphs of Hourly Observations of Wind, Temperature, and Pressure at Barter Island During a Period of Strong Westerly Winds.

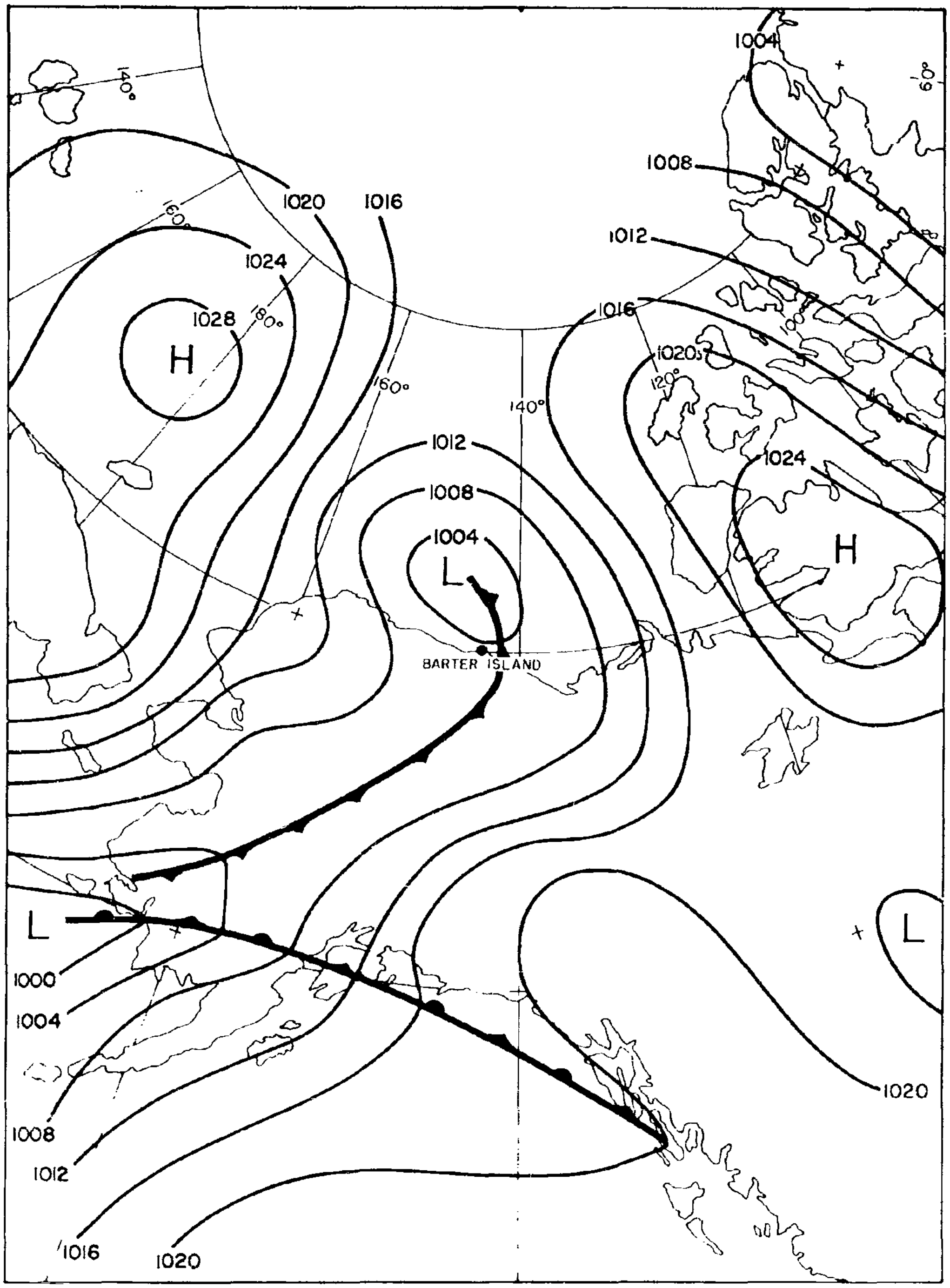



Figure 5.7. Surface Chart for 0030 G.M.T., 4 January 1957.

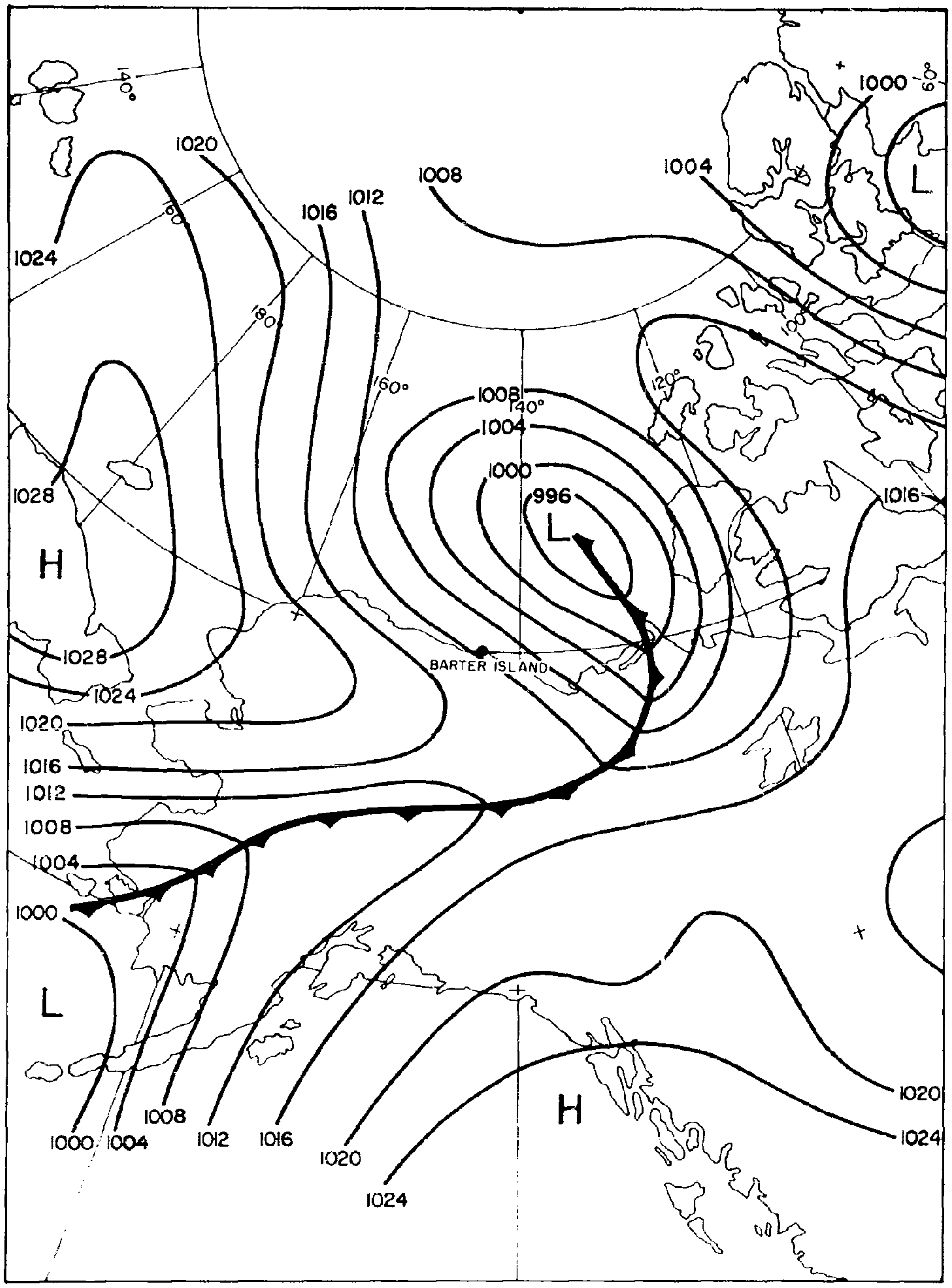



Figure 5.8. Surface Chart for 0630 G.M.T., 4 January 1957.

produce observed features of the wind distribution through application of classical hydrodynamic theory. Moreover, the accompanying observed disturbances of the pressure field have also been reproduced by the theoretical model.

The form of the appropriate equations is such that they may also be solved by means of electrical analogues. An electric current is passed through impregnated paper from which the shape of the obstacle has been cut. A voltmeter is then used to determine the voltage at different points on the paper. Lines of equal voltage are drawn, and lines orthogonal to these are sketched by eye. The latter represent the streamlines of flow, and from these the direction and relative magnitude of the current (electric or fluid) can be found. Miyake [6] has shown that this method accurately predicts the wind distribution in the vicinity of Barter Island at times of strong winds. It is possible that this method would also explain strong winds, falsely labelled as katabatic, at other arctic sites.

5.1.4 Forecasting of Strong Winds

Because of the local nature of these winds, it is not possible to set down general forecast rules that will apply at all stations. However, as has been stressed repeatedly, a well developed pressure gradient appears to be the common denominator in strong wind situations, and it is usually possible to find a pressure difference between two stations in the area that will serve as a criterion for forecasting both the onset of the winds and the probable peak velocity or peak gust. The problem therefore, reduces to one of forecasting a specific pressure gradient in the vicinity of the station in question.

Frequently the pressure prognostication is handled simply by noting characteristic pressure patterns that are known to be associated with the outbreak of high winds. Strictly speaking, such a procedure does not entail a prognostication but involves a simultaneous relationship. However, at many stations there is often sufficient delay between the establishment of the pressure gradient and the onset of the winds to derive some usefulness from this approach.

Short-range forecasts may also be based on various local indications. Lenticular clouds often foreshadow strong wind occurrences at localities subject to the foehn-type wind. Blowing snow on nearby ridges warn of the possible presence of a fast moving current a short distance above the ground. Pilot reports of severe turbulence in the near-ground layers likewise serve as a danger signal.

Perhaps the most sophisticated methods of handling the problem of strong wind forecasting have been developed by forecasters at Thule, Greenland. We conclude this section with a brief review of these techniques.

The first step in the forecast involves a classification of the pressure pattern. If the situation falls within one of four specified types, then it is a potentially dangerous one in regard to high winds, and further steps must be taken. The exact nature of these steps differs according to which of the four types is present. We shall consider only the additional measures taken with what is termed Type I -- a low, attended by a cold front, which enters Davis Strait from the south through west and moves northward toward Thule.

The second step is to determine whether or not the low will pass Thule. This is done by considering a combination of 700-mb. height differences in the vicinity of Baffin Bay. If a certain critical number is exceeded, the low is forecast to pass the station. Of the lows which pass the station only those which pass to the west bring strong winds. To judge which side the low will pass on, the 1000-to 500-mb. thermal wind is used as a predictor.

The 700-mb. height-difference index mentioned previously has been found to be well correlated with the speed of the low, and a scatter diagram has been prepared which gives the speed corresponding to a given value of the index. This diagram aids in the prediction of the onset of the winds; moreover, it provides a means of predicting the peak gusts since these are found to average 100 percent greater than the average velocity of the low center from 65° N. to Thule.

After the cold front passes Upernavik, the wind forecast is modified by means of a further scatter diagram relating the peak gusts to the maximum sea level pressure difference between Upernavik and Thule.

From the physical standpoint the foregoing procedures are not satisfying, since it appears that the predictors have been selected more on the basis of empiricism than physical understanding. However, until the origin of the strong winds is better understood, the empirical approach is the only one available, and the techniques used at Thule may prove to be valuable guides in setting up similar schemes at other stations affected by such winds.

### 5.1.5 Further Remarks on Wind Forecasting

Our remarks so far have pertained only to the forecasting of abnormal winds connected with terrain influences. The more general problem of forecasting upper-level winds and surface winds without pronounced topographical components will be considered only briefly, since this problem is handled much the same in the Arctic as in middle latitudes.

The current and prognostic surface and upper-level charts are the main tools used in the routine wind forecasts. At upper levels the velocity can usually be estimated with sufficient accuracy by use of the geostrophic wind scale. At the surface, it is often necessary to allow for the effect of the curvature of the isobars in determining the gradient wind. Moreover, the reduction in speed due to friction and the angular deviation between actual and gradient winds must be considered.

In practice these factors are often taken account of simply by extrapolation of current conditions. If this is not feasible, a gradient wind scale can be applied to the prognostic chart and the resulting wind modified to allow for friction.

The frictional influence varies with locality and season. For wind speeds greater than 4 knots the surface wind speed in winter is about 30 to 60 percent of the gradient speed, and the angle separating the two winds averages approximately 30° to 35°. In summer the corresponding figures are 60 to 70 percent and 20° to 25°.

The seasonal differences are a reflection of the fact that the frictional effect depends on the static stability, and therefore, this effect may be expected to be related to the synoptic pattern as well. For winds lighter than 4 knots, local effects become important and obscure the geostrophic control.

## 5.2 Temperature

Our concern here is with the short-term, nonperiodic changes of temperature which are often of importance in forecasting, either because of the direct effect of temperature on various operations and activities, or because of the connection of temperature with other elements of the forecast such as fog and icing. Though large, rapid fluctuations of temperature are not a dominant feature of the thermal regime of the Arctic, nevertheless, they do occur on occasions at most seasons and localities. Cold fronts moving along the shores of the Arctic Ocean in summer may bring large, sudden drops of temperature in their wakes at coastal and inland stations. In winter invasions of warm air from the Atlantic may, in extreme instances, raise temperatures by as much as 40° to 50° F. in the vicinity of the pole. When the flow of warm air ceases and skies clear, equally drastic falls in temperature may ensue. Only over the Arctic Ocean in summer, when the melting of the pack ice holds temperatures close to the freezing point, is there an absence of significant temperature fluctuation; and even then, as noted previously, it is only the layers immediately adjacent to the surface that are so affected.

### 5.2.1 General Remarks on Temperature Prediction

The change of temperature that is observed at a given spot may be regarded as the sum of two effects: (a) the transport of air of different temperature to the spot and (b) the warming or cooling of the air en route. The first process is usually referred to as temperature advection and can be evaluated simply by displacing air parcels with the wind and noting the differences between the initial and the later temperatures at the end points of the trajectories. An accurate wind forecast is obviously a prime requisite for a correct prediction of the advective change.

The heating and cooling processes may be conveniently divided into two categories, those due to adiabatic compression and expansion and those due to nonadiabatic or diabatic processes; that is, processes involving actual transfer of heat. Near the surface adiabatic heating is usually negligible since it depends mainly on the vertical component of motion, which is zero over level ground. In the free atmosphere and over sloping terrain, adiabatic effects may become a major factor in temperature change. The foehn winds of the Greenland coast furnish an outstanding example of the extreme warming that may be brought about by adiabatic compression. (Strictly speaking, at the coast itself the warming is the result of advection of the foehn air, but the basic cause of the temperature rise is clearly the adiabatic compression that occurs along the slopes.)

There are a number of nonadiabatic processes which exert an important influence on the temperature, among them short- and long-wave radiation, eddy and molecular heat conduction, evaporation and condensation, and melting and freezing. As a rule, several of these processes act simultaneously, some processes predominating under one set of conditions and other processes under a different set. Although for

the sake of convenience the processes will be discussed separately, it must be borne in mind that they rarely are independent of one another.

The direct absorption of short-wave or solar radiation by the atmosphere is exceedingly small, leading to heating rates of the order of only a fraction of a degree per day. As a consequence, any significant temperature rise connected with insolation must be attributed to indirect processes which transfer heat from the earth's surface to the atmosphere.

The conditions that prevail over the Arctic are such that in most circumstances even the indirect heating effects are slight. Foremost among those conditions is the high reflectivity or albedo of snow and ice surfaces, 50 percent to 80 percent of the incident radiation being reflected from these surfaces without ever entering into the heat budget. During the warm season when insolation is greatest, the melting of the snow and ice reduces or prevents warming of the air over a substantial portion of the Arctic. On the other hand, over continental portions of the Arctic the snow cover disappears entirely in summer, and it is then possible for pronounced insolational warming to occur, especially in air masses that migrate southward from the pack ice.

The absorption and emission of long-wave radiation by the atmosphere is greater than the absorption of short-wave radiation, thus making it possible to produce temperature changes of the order of 2° to 3° F. per day or even greater. However, as in the case of solar radiation, large rapid temperature changes connected with long-wave radiation actually involve a complex of processes in which the earth's surface plays an important role.

The surface (whether covered by snow, ice, water, vegetation, or earth and rock) may be considered a black body with respect to long-wave radiation; that is, a body which radiates the maximum possible amount of energy at these wave lengths. From Stefan's law, this amount is known to be proportional to the fourth power of the temperature. The atmosphere, on the other hand, absorbs and emits selectively, acting as a black body at certain wave lengths and being effectively transparent at others. The principal absorbing gases are water vapor and carbon dioxide.

For normal temperature and moisture stratifications the down-coming long-wave radiation from the atmosphere is less than the emission from the earth's surface, so that, as a rule, there is a net loss of long-wave radiation from the surface. This loss results in a cooling of the surface (at night) and, through heat conduction and radiative exchange, of the adjacent air strata.

So far in the discussion no mention has been made of the effects of cloudiness. Clouds of sufficient depth behave as black bodies which emit, both upward and downward, the black-body radiation appropriate to their temperature. In the case of a cloud imbedded in an inversion layer, the downward radiation from the relatively warm cloud will exceed the upward radiation from the cold ground, and the earth's surface will warm. Many of the largest nonperiodic temperature changes in the Arctic are connected with changes in cloud amount. As examples of the effect of cloudiness on temperature during the polar night, we present the sequences of 3-hourly observations in tables 5.2 and 5.3. The characteristic warming with increased cloud cover and cooling with clearing skies may be noted. The majority of cases of large temperature fluctuaion are associated with changes of wind speed as well, so that only occasionally can the effects of cloudiness be so clearly isolated.

Molecular heat conduction does not directly affect the temperature at the level of the instrument shelter but is important in heat transfer at the earth-atmosphere interface and in the layers immediately below the surface. It thus plays a

TABLE 5.2

Three-Hourly Synoptic Observations for Eureka, Ellesmere Island, November 1 and 2, 1950, Showing Warning in Connection with Increased Cloud Cover [9].

| Date | Hour (GCT) | Sky Cover (Tenths) | Temperature | Wind Direction | Wind Speed (mph) |
|---|---|---|---|---|---|
| 1 | 0230 | 0 | -27 | E | 2 |
| 1 | 0530 | 3 | -27 | E | 7 |
| 1 | 0830 | 2 | -24 | E | 5 |
| 1 | 1130 | 8 | -21 | E | 7 |
| 1 | 1430 | 8 | 18 | E | 5 |
| 1 | 1730 | 10 | -15 | E | 4 |
| 1 | 2030 | 8 | 14 | E | 3 |
| 1 | 2330 | 10 | -9 | E | 1 |
| 2 | 0230 | 8 | -8 | C | |
| 2 | 0530 | 10 | -5 | C | |
| 2 | 0830 | 10 | -2 | C | |
| 2 | 1130 | 10 | -3 | E | 5 |
| 2 | 1430 | 8 | 1 | ESE | 4 |

TABLE 5.3
Three-Hourly Synoptic Observations for Eureka, Ellesmere Island, December 20 and 21, 1949, Showing Cooling in Connection with Decreased Cloud Cover [9].

| Date | Hour (GCT) | Sky Cover (Tenths) | Temperature | Wind Direction | Wind Speed (mph) |
|---|---|---|---|---|---|
| 20 | 1130 | 10 | -16 | E | 4 |
| 20 | 1430 | 8 | -16 | SE | 12 |
| 20 | 1730 | 8 | -17 | E | 5 |
| 20 | 2030 | 8 | -17 | E | 7 |
| 20 | 2330 | 10 | -19 | E | 4 |
| 21 | 0230 | 6 | -18 | E | 5 |
| 21 | 0530 | 8 | -19 | E | 2 |
| 21 | 0830 | 0 | -20 | E | 4 |
| 21 | 1130 | 0 | -28 | E | 5 |
| 21 | 1430 | 0 | -30 | E | 4 |
| 21 | 1730 | 0 | -33 | SE | 4 |
| 21 | 2030 | 0 | -33 | E | 2 |
| 21 | 2330 | 0 | -36 | C | |

major role in determining the temperature of the surface itself and thereby indirectly influences the temperature of the overlying strata. Snow is a very poor conductor of heat, and it is this property which is in part responsible for the development of the extreme temperatures of the arctic winter.

Away from the earth's surface, turbulent eddies are far more effective than molecular motions in vertically transporting heat. Except for the comparatively few instances in which thermal convection is present, the eddy motions result from the rubbing of the wind against the ground and obstacles; therefore, the degree of eddying depends on surface roughness, wind strength, and the buoyant resistance of the air to vertical displacement. The eddy heat conduction depends on the product of these factors, as expressed by a coefficient of eddy conduction, and the vertical gradient of potential temperature.

In a stable atmosphere the eddy heat transfer is downward, accounting for the often mentioned tendency of surface temperatures to rise as the wind strengthens. The surface temperature rise depicted in figure 5.1 of the section on strong winds is no doubt due, at least in part, to the eddy flux of heat downward. As the wind dies down the eddy conduction diminishes and becomes negligible when the wind is calm.

Evaporation, condensation, melting, and freezing processes may absorb heat from the air or release heat to it. As a rule they are not important factors in short-term temperature changes. A notable exception to this statement occurs in summer when warm air masses from the continents are cooled over the melting pack ice of the Arctic Ocean.

Our remarks so far have applied to surface temperature changes. Aloft, advection and adiabatic heating or cooling (in association with vertical motions) are the major factors in promoting rapid temperature change, and only rarely will the other processes be of significant size. The adiabatic heating and cooling generally operate so as to counteract, in part, the advective temperature change. This is one reason why the local temperature change at upper levels is characteristically less than is indicated by the advection.

### 5.2.2 Forecasting Minimum Temperature

It is apparent from the foregoing discussion that the temperature change process is so complex that a general attack on the problem of temperature prediction, either from a physical or statistical standpoint, cannot be attempted at this time. However, it is sometimes possible to devise a simplified approach to a particular aspect of the problem and still achieve useful results. An approach of this limited type has been made to the problem of minimum temperature forecasting at Fairbanks, Alaska.

The number of parameters entering into the problem are restricted first of all by assuming that the extreme cases of interest occur only under conditions of light winds and clear or nearly clear skies. Consequently, the prediction diagram is based on data gathered under certain prescribed conditions and can be applied only when these conditions are met.

It is further assumed that under these conditions the thermal equilibrium that exists at time of minimum temperature involves a balance between the incoming and outgoing long-wave radiation, all other heat exchange processes being negligible. From this it follows that the minimum temperature is a function of the incoming radiation only. The latter quantity depends in a complicated manner on the temperature and moisture conditions in the atmosphere, but testing revealed that the warmest temperature in the free atmosphere provided a satisfactory measure of the downward radiative flux. Thus, the value of this temperature at 1700 local time was chosen

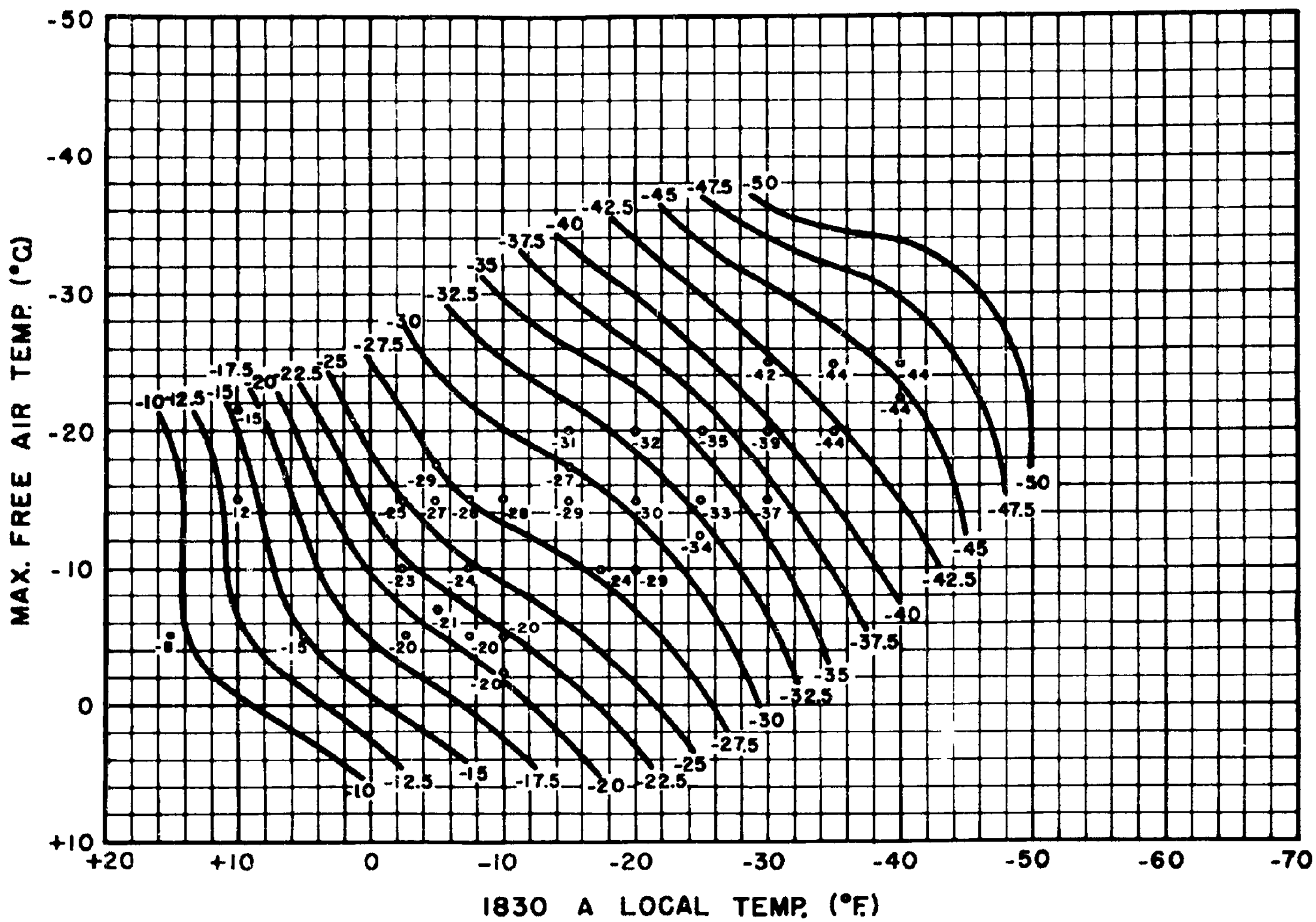



Figure 5.9. Minimum Temperature Forecasting Diagram for Ladd AFB, Fairbanks, Alaska.

as one predictor for the diagram. The 1830 local surface temperature was selected as a second predictor.

The object forecast diagram prepared from these predictors is shown in figure 5.9. The minimun temperature decreases with decreasing initial temperature and decreasing maximum free air temperature. The departure of individual values from the smoothed analysis are rather small, and preliminary application to independent data has given encouraging results.

In using a simple aid of the foregoing type one is always faced with the difficulty of deciding whether the meteorological conditions under which it may be applied will persist throughout the forecast period. Despite this shortcoming, the diagram calls attention to factors that might otherwise pass unnoticed, and insures that the prediction of the quantity in question is consistent with the overall forecast. These advantages are sufficient to justify the labor involved in the preparation of such a diagram.

## 5.3 Clouds and Ceiling

### 5.3.1 Physical Processes in Cloud Formation

We shall not concern ourselves here with the microscopic aspects of cloud formation--condensation and sublimation nuclei and the like --but will assume that, in general, cloud will form when the air becomes saturated with respect to water. At temperatures below freezing the saturation vapor pressure over supercooled water droplets is greater than over ice, so that supersaturation with respect to ice is not uncommon. At very cold temperatures (-40° F. and less) water will generally pass directly from the vapor to solid phase.

In order to saturate air, either the air temperature must be lowered to the dew point by some cooling process, or the water vapor content must be increased by the addition of moisture. In the case of the cooling process, further temperature decrease will lead to slight supersaturation, and cloud droplets will condense. The evaporation process, however, cannot raise the the relative humidity above 100 percent so that although it may contribute to cloud formation, it cannot in itself produce cloud.

Of the several cooling processes mentioned in the section on temperature change, only two are of importance in the actual generation of cloud: adiabatic cooling and cooling due to vertical mixing. Other processes such as long-wave radiation may increase the relative humidity of moist air, thereby hastening condensation, or may thicken clouds once they have formed, but as a rule they will not be critical factors in cloud formation. Next to the ground these processes are of primary importance in initiating condensation; but here the result is fog, not cloud.

It is convenient to consider the various cloud types and patterns in relation to the scale or type of motion involved in their formation. In the discussion which follows we shall distinguish five categories: (a) large-scale upglide, (b) convective updraft, (c) turbulent mixing, (d) a combination of convection and mixing, and (e) orographic uplift.

The large-scale upglide motions occur most commonly on the forward side of the baroclinic or wave disturbances, the familiar cyclonic storms of middle latitudes. The vertical motions are of the order of 1 to 10 centimeters per second, and the adiabatic cooling of the rising air masses causes cloud to condense in the classical sequence of cirrus and cirrostratus, lowering to altostratus and altocumulus, and then to nimbostratus. The arctic disturbances described in section 3.1 of chapter 3 appear to have this characteristic cloud distribution, though there is evidence that the different cloud types often lie in discrete overlapping strata rather in a single, sloping layer.

Convective updrafts occur on the scale of cumulus cloud or thunderstorm cells and are about one hundred times greater in magnitude than the gentle upglide motions discussed previously. Generally speaking, convection is poorly developed in the Arctic, so that cumuliform clouds are uncommon--except in summer over heated land masses and in winter over open bodies of water. Occasionally convection can develop aloft, presumably as the result of the release of convective instability in lifted air masses. On such rare occasions it is not impossible for the convective activity to occur over the arctic seas. In one instance, the Ptarmigan Weather Reconnaissance Flight encountered a thunderstorm over the Beaufort Sea at a position nearly 300 miles north of the Alaskan coast. This case occurred in July in a current of warm, moist air of Siberian origin.

Under stable conditions vertical mixing leads to a downward transport of heat, and thereby to cooling aloft and also condensation, provided that the moisture content of the mixed layer is sufficiently high. The cloud formed in this way is the dull, shapeless stratus which is the

characteristic cloud-form of the Arctic Basin in summer and early autumn. The ultimate cause of the stratus is the advection of warm, moist air from continental portions of the Arctic. The wind speed largely determines whether the stratus will be elevated, and hence a true stratus, or whether it will stay close to the surface as fog.

Over the Arctic Ocean the stratus is at a maximum in late August and September. At coastal locations a double maximum occurs, one in May and a second in October. In the intervening months the frequency diminishes slightly because of the solar heating.

When a shallow convection exists in conjunction with the mixing, the cloud layer assumes the sharper and more rounded outline of the stratocumulus cloud. This cloud occurs much less frequently than the pure stratus.

When air is forced to rise over a hill or mountain, orographic clouds may develop as a result of the adiabatic cooling. Under certain conditions the orographic lifting gives rise to very distinctive cloud forms, as will be discussed in the section on turbulence. Lifting and heating by mountain slopes has a considerable influence on the summertime cumulus activity in certain areas, the shower-bearing clouds appear to form mainly over high ground.

#### 5.3.2 Ceiling

For aviation purposes the cloud base or ceiling is usually the most important part of the cloud forecast. In theory, it is a relatively simple matter to estimate where the cloud base will form under specified conditions. However, in practice the basic parameters are rarely known in advance with sufficient accuracy nor can they be forecast easily.

The bases of clouds connected with large-scale upglide can be determined from a knowledge of the moisture distribution by displacing parcels upward following the dry adiabatics on a thermodynamic diagram and noting the condensation levels (LCL). The same approach is also used in determining the height of orographic clouds.

Convective cloud bases lie near the convective condensation level (CCL) if the convection originates close to the ground. The idea of the mixing condensation level (MCL) may be used to estimate the probable height at which stratus will appear. For details concerning the various levels, the reader is referred to a standard text such as Volume II of Petterssen [8].

#### 5.3.3 Forecasting of Clouds and Ceiling

It would be nice if the foregoing physical principles could be used in predicting cloud distribution and height, but to apply them it is necessary to know the moisture distribution, the field of vertical motion, and the degree of mixing. Of these factors only the moisture is known with any precision at the beginning of the forecast period and none are accurately known at the end, though a promising start on vertical velocity prediction has recently been made by the Joint Numerical Weather Prediction Unit. Consequently, in forecasting cloud type and ceiling, it usually is necessary for the forecaster to fall back on his two old standbys--extrapolation and statistical methods.

Extrapolation usually involves displacing the cloud patterns with the pressure system, making allowances for deepening and filling and for local effects. Statistical methods range all the way from empirical rules, handed down by experienced forecasters, to prediction diagrams in which past data are analyzed either graphically or statistically to obtain relationships between selected variables.

Where observations are scarce, the forecaster often resorts to models of cloud distribution in filling in the cloud cover. The Bjerknes or Norwegian cyclone model is the best known of these models and probably provides a useful estimate of the cloud distribution in arctic systems that possess distinct fronts of appreciable depth.

Up to the present, to the writer's knowledge, no objective diagrams for predicting ceiling height have been prepared for arctic stations. However, various rules of thumb dealing with important local problems, such as the forecasting of summer stratus at Narsarssuak, are contained in station manuals.

### 5.4 Precipitation

#### 5.4.1 General Remarks on Precipitation

In order for clouds to precipitate it is necessary for many small cloud droplets to collect together into larger drops. Two processes are recognized as being of importance in the formation of precipitation elements: (a) the collection of moisture by ice nuclei in supercooled clouds and (b) the coalescence of droplets of different size due to the different rates of fall.

The details of the precipitation process are of special interest to the cloud physicist, but are usually not regarded as vital to the central forecast problem of when and where precipitation will fall and in what amount. Just as it is reasonable to assume that cloud will form when the relative humidity reaches 100 percent, whatever the micro-processes involved, so it is probably also safe to assume that precipitation will fall from the cloud in rough proportion to the rate at which water vapor condenses in the cloud. From this point of view the precipitation problem may be regarded as an extension of the problem of cloud formation and, therefore, may be approached in much the same manner.

In regions of large-scale upglide, precipitation is normally light and continuous. Snow, of course, is the most common precipitation form at high latitudes. However, rain may fall in any part of the Arctic in mid-summer--even at the pole--and may fall in outer portions of the Arctic, which border on the open seas, in any month. During the cold season the border areas are also susceptible to infrequent occurrences of freezing rain.

The area encompassed by precipitation is, of course, smaller than that covered by cloud. Part of the cloud shield is too high and thin to precipitate to the earth. Moreover, the clouds may spread out horizontally beyond the region of upglide, while the area of precipitation must remain close to the region of most active upward motion.

Showers will nearly always fall from the deeper, more vigorous convective cells. These develop primarily in summer over continental areas, especially where aided by the topography. Snow showers occur with relatively flat cumulus or stratocumulus in cold air masses that move across open water. Along the eastern shore of Hudson Bay a substantial snowfall results from such showers in November before the bay freezes over. Snow flurries may occur during the cold season downwind from open leads in the pack ice of the arctic seas.

Convection is occasionally strong enough in summer to initiate thunderstorm activity in areas south of 75° latitude, except over and adjacent to Greenland. Over interior lowlands it appears that these thunderstorms may occasionally be of the airmass type. However, many thunderstorms in both Alaska and Siberia develop in the vicinity of topographic features so that it is likely that orographic uplift is also a factor in their information.

As pointed out in chapter 3, and elsewhere, cyclonic disturbances with strong cold fronts frequent the belt near the Arctic Circle in summer. A certain percentage of the arctic thunderstorms are connected with these synoptic features, and it is safe to say that the rare thunderstorms that occur over the Arctic Ocean in summer must all be attributed to the presence of storms or fronts.

Mixing processes usually are not strong enough to cause precipitation from the characteristic stratus of the arctic summer. On the occasions when precipitation is observed from the stratus, it is in the form of drizzle.

The orographic influence is most pronounced along the fringes of the Arctic where mountain or ice barriers intercept maritime air currents. Thus, the windward slopes of the coastal ranges of Alaska receive hugh yearly precipitation amounts, and large amounts also are measured along the south coast of Greenland. In areas sheltered by mountains, such as the region about Anchorage, Alaska, the precipitation can be surprisingly small when compared to nearby exposed areas.

Orographic features can also cause marked abnormalities in the precipitation distribution with respect to the synoptic pattern. At Thule, Greenland, the northeast quadrant of a low is generally free of precipitation, and sometimes of cloud, evidently because of downslope flow.

The importance of elevated terrain in summertime shower and thunderstorm activity has already been mentioned.

5.4.2 Snowfall

In some ways the prediction of snowfall is a lesser headache to the arctic forecaster than to the forecaster in more southerly latitudes. The decision as to whether expected precipitation will be in the form of rain or snow is seldom as difficult at high latitudes, and the depth of snow that can fall in any one storm is usually much less. The foregoing remarks apply to the Arctic as a whole. In certain areas of the subarctic--southern Alaska, the Aleutians, Kamchatka, southern Greenland, and Labrador, to mention a few--the problem can at times be acute.

Within the inner Arctic, yearly precipitation (rainfall plus water equivalent of snow) amounts to about 4 inches, the major part of which comes in summer. From this it is apparent that snowfall is usually very light in winter, almost al-

ways less than 1 inch in a single fall. At the end of the accumulation season the snow cover usually averages only about 1 foot. Much of the summer precipitation comes as rain, but well to the north snowstorms are possible in any month and during the warmer months may bring heavy falls. The greatest snowfall on record appears to be a depth of 20 inches in 24 hours measured close to the North Pole at the Russian drifting station, North Pole 1.

5.4.3 Forecasting Precipitation

As with cloud forecasting, the use of models and extrapolation procedures are standard techniques in precipitation forecasting and require no elaboration here. Physical and statistical approaches to the precipitation prediction problem have also been devised but as yet have not been applied to the arctic region.

Under certain simplifying assumptions, a quantitative expression may be derived relating the precipitation rate to the moisture distribution and vertical velocity. Difficulties in measuring and predicting vertical velocities have prevented the use of the physical approach up to now, but with the development of numerical weather prediction these difficulties are being overcome, and it is possible that quantitative estimates will be available in the near future.

Statistical techniques for forecasting the occurrence or nonoccurrence of precipitation and the probable amount have been developed for a number of middle-latitude stations. No doubt similar techniques could be used in the Arctic. A more detailed discussion of methods of quantitative precipitation forecasting is found in Volume II of Petterssen [8].

In predicting snowfall it is advisable to estimate the water equivalent and convert to snow depth. A ratio of 10 inches of snow to 1 inch of water is a suitable average figure. Of course, in many instances, the main problem is to decide whether the precipitation will be in the form of rain or snow. The 0° C. isotherm is the critical factor in this respect and, because of the possibility of shallow surface inversions, it is better to base the decision on the predicted temperatures at the 850-mb. level rather than on surface temperatures.

When above-freezing temperatures are indicated aloft but the cold film is predicted to persist at the surface, a forecast of freezing rain is in order.

The subject of ice crystal haze will be discussed in the section on visibility, though strictly speaking it should be included under precipitation.

5.5 Visibility

5.5.1 General Remarks

The arctic atmosphere is relatively free of impurities so, in the absence of falling or suspended condensation products, visibilities are characteristically high in polar regions. Moreover, unusual optical effects caused by the bending of light rays under the arctic inversion may, on occasion, lead to abnormal extensions of the visual range. Our concern here, however, will be with conditions which reduce or restrict visibility, of which fog, ice fog, and falling and blowing snow are generally regarded as the most significant in the regions to be considered.

In addition to conditions of low visibility, the arctic forecaster must occasionally contend with a strange phenomenon known as the "arctic whiteout", when the visibility is in a sense indeterminate. This condition occurs when the sky and landscape become uniformly white. The lack of contrast between objects both near and far effectively reduces the visibility to near zero, though a black object introduced on the scene would be discernible at a great distance.

5.5.2 Fog

Under this topic we shall consider only water fogs, leaving for separate discussion the subject of ice fog. At least four of the recognized types of fog are observed in the Arctic--advection, radiation, upslope, and steam fogs. Of these, advection fog is of greatest importance since it affects large areas and often leads to very low visibilities. It is primarily a summertime phenomenon and occurs wherever warm, moist air blows over a cold surface and becomes cooled below its dew point. Winds must also be light or else the fog is lifted and transformed into a stratus layer.

Advection fog is particularly prevalent over the arctic seas during the months of July, August, and September. It forms when warm, moist air of more southerly origin blows into the Polar Basin. The fog dissolves rapidly in return currents that are heated over land.

The characteristics of the summer advection fog over the pack ice have been studied extensively for the summers of 1957, 1958, and 1959 by

Hansen [2], using data from drifting stations Alpha, Bravo (T-3), and Charlie. For these stations and years fog occurred on 3 percent of all observations in May, 10 percent in June, 15 percent in July, 25 percent in August, and 7 percent in September. As a result of variations in circulation patterns and station locations, large year-to-year variations in fog frequency were observed in any month. During the months of July and August no evidence of a diurnal variation in fog frequency was noted.

Fog was most frequent over the pack ice at temperatures close to freezing and at pressures that were in the range of 1000 to 1030 mb. At pressures lower than 1000 mb, the typical low-level temperature inversion was generally absent, and fog was rare. At extremely high pressures the inversion was well developed, but due to strong subsidence relatively dry air was present very close to the surface. Because of the homogeneity of the surface the frequency of occurrence of fog was little affected by wind direction, showing only a slight decrease in frequency for northwesterly winds. As would be expected fog was most frequent at light wind speeds. Not a single case was recorded at speeds of 20 knots or greater.

At coastal and insular stations the summer advection fogs show a distinct diurnal cycle, with maximum occurrence in early morning and minimum in early afternoon. The fogs at these stations should, therefore, be classified as combined radiation-advection fogs.

Radiation fog is local in nature, being restricted mainly to inland valleys. It is generally of short duration and is most likely to occur during the months with above freezing temperatures. Because of the difference in vapor pressure over ice and over water a snow cover tends to dissipate or inhibit the formation of radiation and other water fogs, so that below the freezing point the frequency of fog diminishes with decreasing temperature until the critical temperatures for the formation of ice fog are reached [8].

Upslope fog occurs where the air is lifted over sloping terrain and may be regarded as either a cloud or fog, depending on the point of view of the observer. Expeditions to the interior of Greenland frequently report upslope fog, usually in conjunction with precipitation [11-Vol. 1].

Steam fog, or arctic sea-smoke, is often regarded more as a curiosity than as a serious impairment to visibility. However, it may, under certain circumstances, become quite dense and persistent and present a hazard to both shipping and aircraft operations. Steam fog develops when very cold air blows over open water, thereby promoting a rapid transfer of heat and moisture from the water surface to the air. The heating from below produces a unstable lapse rate and associated small-scale convective eddies which carry the warm, moist surface strata aloft where they mix with the cold, dry air above. The first result of the mixing is to produce condensation in the form of fog, but with further mixing of the dry air the fog dissolves.

As a rule, then, the fog appears as wisps of smoke emanating from the sea surface. However, if a strong inversion is present the upward mixing is confined to a relatively shallow layer, within which the fog collects and assumes a more uniform density. In these circumstances the visibility may be reduced to 300 yards or less.

Steam fogs occur most frequently along the shores of open bodies of water in fjords or inlets where the water is kept open by wind or tidal action. Small patches of steam fog are also observed during the cold season over fresh leads in the pack ice of the arctic seas.

### 5.5.3 Forecasting Water Fogs

Although the problem of fog prediction is often amenable to objective forecast methods of the statistical type, there have been few successful prediction diagrams prepared for arctic sites. This may reflect a lack of sufficient effort in this respect, or in cases where attempts have been made, may mean that the problem is often too complex to yield to crude or simple methods. For instance, the occurrence of advection fog at a particular location, such as Thule, may depend more on the vagaries of the local wind than on the large-scale flow pattern responsible for transporting the fog to the general area.

Hansen [2] in his study of summer fog in the central Arctic devised a fog prediction diagram (fig. 5.10), applicable to the region of the pack ice, which promises to give useful forecasts. So far the diagram has been tested only on dependent data. In order to use the diagram upper-air soundings are required. The forecast is for the 18 hours following the time of observation and is in the form of a yes or no answer to the question of whether fog (visibility $< 3/4$ mile) will occur on any 3-hourly observation during the period.

The steps in using the diagram are described by Hansen as follows:

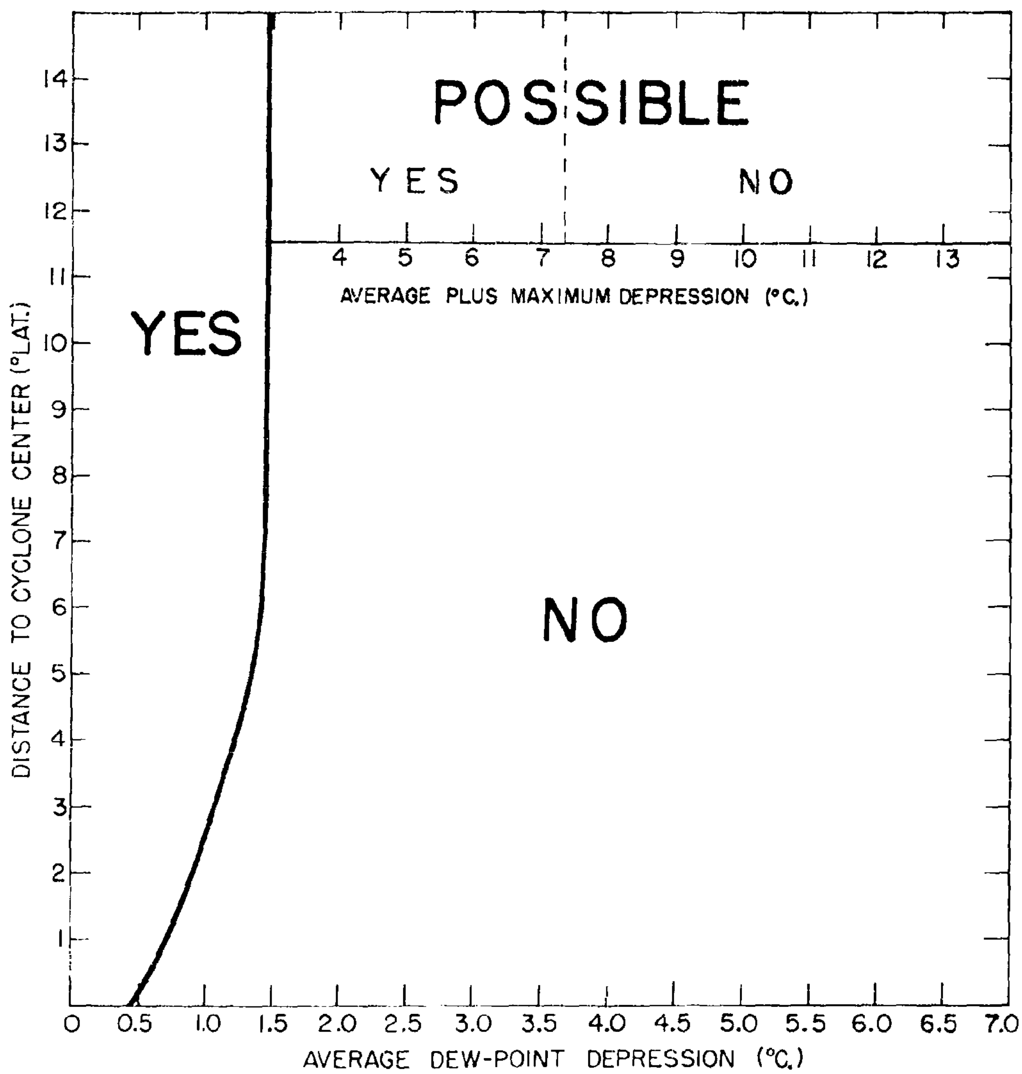



Figure 5.10. Final Fog Prediction Diagram.

(1) "Compare the temperature at the surface and 900 millibars. If the 900-mb. level is colder, forecast no fog. When the surface is colder, proceed to the next step.

(2) "Measure the distance, in degrees of latitude, from the station to the center of the nearest cyclone. If the station is definitely under an anticyclonic influence, the distance is automatically taken to be 14 degrees.

(3) "Compute the average dew-point depression in the lowest 50 millibars or to the top of the inversion, whichever is lowest. (The average temperature and dew point can each be found by simple areal averaging - the distance between the two being the desired depression.)

(4) "With the values from (2) and (3) enter the diagram. If the plot lies in a Yes or No region, the forecast is completed. If it falls in the Possible area, proceed to step (5).

(5) "Add the greatest dew-point depression to the average depression of (3) and use the abscissa of the Possible area to make the final forecast."

The few cases of failure of the method appear to result from changes in temperature and moisture stratifications during the forecast interval. Its general success would seem to imply that the stratifications do not, as a rule, change very rapidly and that the summer fog is an erratic phenomenon whose subsequent probability of occurrence is better related to conditions on the current sounding than to the actual occurrence

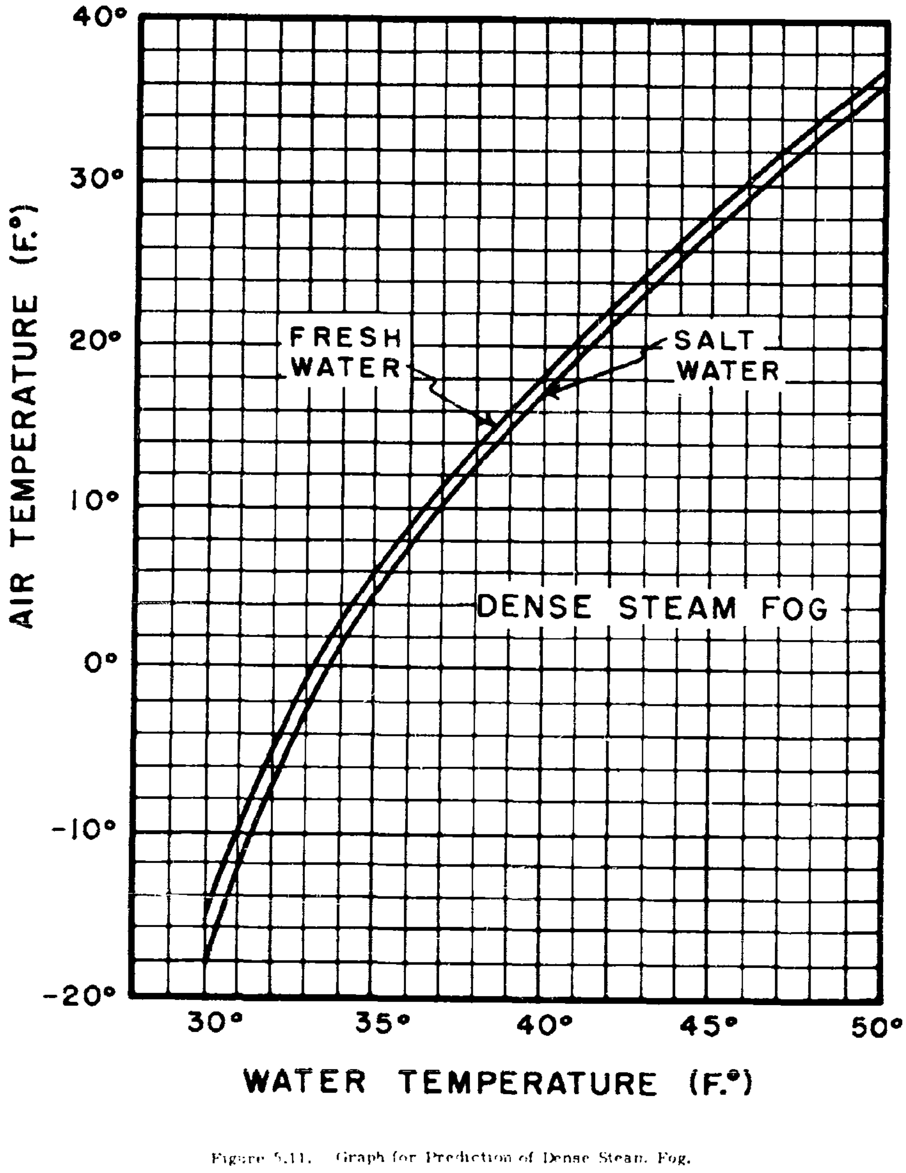



Figure 5.11. Graph for Prediction of Dense Steam Fog.

or nonoccurrence of fog at observation time.

The main parameters in the prediction of visibility reduction due to steam fog would ap- to be the difference between the saturation vapor pressure of the air and the vapor pressure corresponding to the temperature of the water surface, and the initial strength of the inversion. A vapor pressure difference of 5.2 mbs. has been found to be the lower limit for thick or dense fogs.

A graph for the prediction of dense steam fog, based on the foregoing criterion,[1] is presented in figure 5.11. A knowledge of the water temperature and an estimate of the air temperature are required for use of the graph. The wind speed has been found to have little, if any, effect on the formation of steam fog. When blown inland by the wind, the fog normally dissipates rapidly.

### 5.5.4 Ice Fog

Ice fog is largely a man-made addition to the arctic scene and perhaps for this reason has been the object of more intensive study than the more familiar water fogs. From this study much is known concerning the composition of ice fog and its mode of formation, and at least moderate success has been achieved in developing methods for its prediction.

Ice fog of significant density is found only in the vicinity of human habitation where large quantities of water vapor are added to the air through the burning of hydrocarbon fuels. Steam vents and motor vehicle and jet aircraft exhausts are among the important sources of moisture that can produce sharp reductions in visibility in restricted areas.

Meteorological conditions favoring the formation of ice fog are low temperatures (usually below -20° F. and preferably below -30° F.) and a low-level inversion that traps and concentrates the moisture in a shallow layer. At the colder temperatures the ice particles are found to be roughly spherical in shape and are believed to be formed from the rapid freezing of a cloud initially composed of water droplets.

### 5.5.5 The Prediction of Ice Fog

The combustion of hydrocarbon fuels adds heat and moisture to the air in known amounts. Appleman [9] has shown that -- depending upon the initial environmental temperature, pressure, and relative humidity -- the effect of this addition may be either to raise or lower the relative humidity of the environment; and for various environmental conditions he has computed the critical temperatures below which combustion will lead to supersaturation and condensation. The computations are based on the known ratio of moisture emission to heat emission for hydrocarbon fuels. For locations close to sea level, a constant pressure of 1000 mb. may be assumed so that the critical temperature will be a function of relative humidity (or dew point) only.

The critical temperature, however, merely serves as a criterion for determining whether condensation will appear sometime during the period of mixing of exhaust gas and air and does not, in itself, provide an indication of the persistence of the fog. This latter characteristic is determined by the degree of mixing, being greater when the mixing is small (but in excess of a certain minimum), and also by the initial relative humidity--which must be equal to or greater than the saturation value with respect to ice in order for the fog to endure.

On the basis of the foregoing considerations, Appleman has constructed the ice-fog prediction diagram shown in figure 5.12. The diagram is divided into four regions defined as follows:

I. Region of persistent ice fog. This region lies below the critical temperature curve and within it the atmosphere is supersaturated with respect to ice.

II. Region of nonpersistent ice fog. This region also lies below the critical temperature curve, but is nonsaturated with respect to ice.

III. Region of no ice fog. Combustion causes drying within this region.

IV. Region of no ice fog. Combustion causes drying and the temperature is too high for droplets to freeze.

In order to make a prediction, the forecaster simply enters the diagram with the predicted values of temperature and dew point, and determines the appropriate region. The success of the diagram when applied to data for the period December 1946 through February 1947 at Fairbanks, Alaska, is indicated by the dots and crosses in the figures. A striking relationship between the predicted and actual occurrences is noted.

[1] In accordance with the 6th revised edition of the "Smithsonian Meteorological Tables," a value of 4.93 has been used instead of the earlier value of 5.2 mbs.

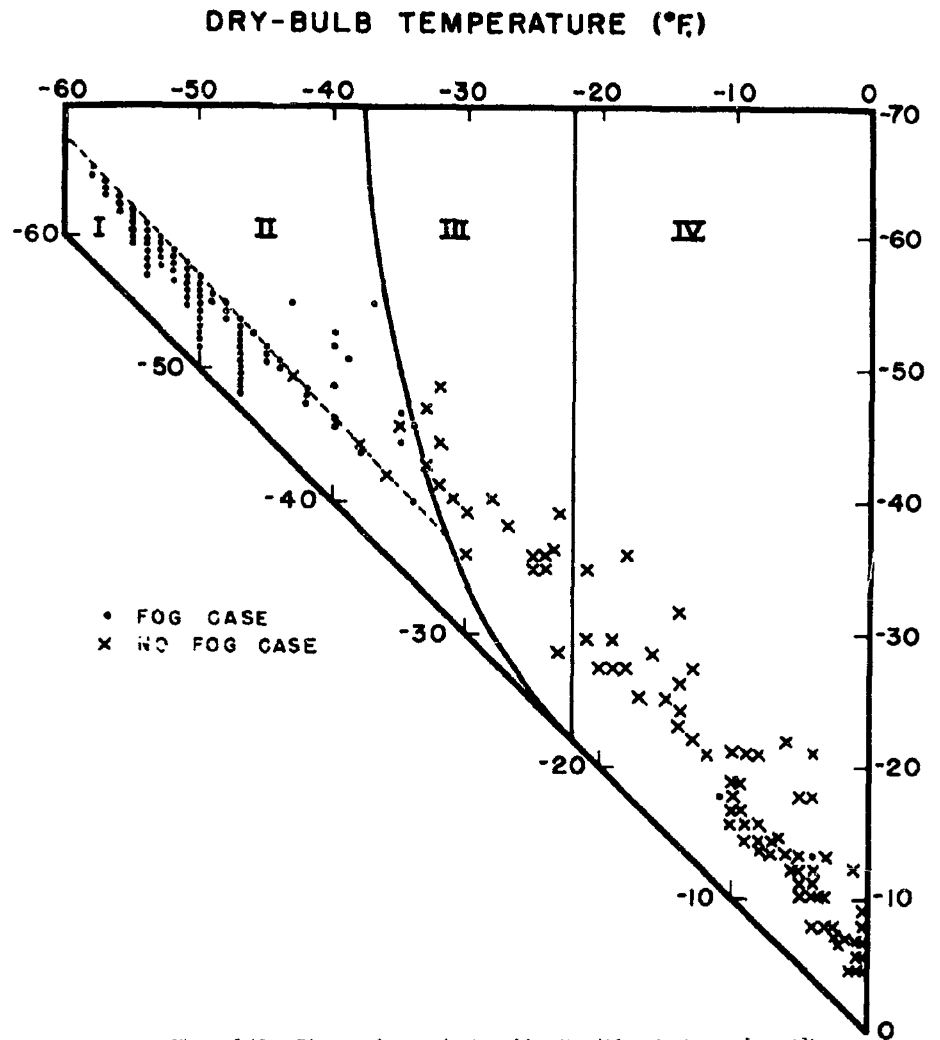



Figure 5.12. Diagram for Prediction of Ice Fog (After Appleman [See 9]).

Later tests conducted at Ladd Air Force Base, on the outskirts of Fairbanks, showed marked deviations from the earlier results. In particular, a large number of ice fog occurrences were observed in Region III. Since Appleman's graph is based on the ratio of heat to moisture content in the exhaust of jet engines, such a discrepancy can be explained by the presence of some moisture source with a lower heat to moisture ratio. The discharge from steam plants is one possibility in this respect.

Because of the foregoing difficulty and the additional fact that it does not take any account of the degree of mixing, Appleman's diagram is perhaps best regarded as a useful guide in setting up objective studies of ice fog. At Ladd, the available moisture depends so highly on the degree of human activity and, therefore, on the hour of the day, that the temperature and hour have been used as the sole predictors in deriving an objective forecast diagram. Appleman's theory would suggest, however, that the mixing, as determined by wind speed and inversion strength, and the relative humidity of the environmental air are also important to the problem.

The frequency with which ice fog occurs at

a specific arctic station is thus seen to be a complex function of temperature, relative humidity, and type and amount of moisture addition. On the basis of the climatic records alone it is possible, with the help of Appleman's diagram, to make a reasonable estimate as to the maximum number of days of ice fog that can be expected at a given location. How closely this figure conforms to the observed depends mainly on the present level of human activity at the location.

### 5.5.6 Blowing Snow

At many arctic locations blowing snow is the most frequent cause of low visibility. This is particularly true in winter when the ground is often covered with fresh, loose snow and other factors in visibility reduction, such as fog, are at a minimum.

At wind speeds as low as 5 knots fresh-fallen snow begins to creep or drift. At speeds of 10 to 15 knots snow is swept into the air in a sufficient quantity to reduce the visibility close to the ground. As the wind increases beyond 15 knots the snow is carried aloft to greater heights and in greater amounts so that rapid deterioration of visibility results. Table 5.4 shows the average relationship between wind speed and visibility at Barrow, Alaska, Resolute, N.W.T., Northice, Greenland (elevation, 7,700 feet, 78° N. 38° W.), Maudheim, Antarctica, Site I, Greenland (elevation 3,800 feet, 120 miles north of Thule Air Force Base), and Site II, Greenland (elevation 6,800 feet, 200 miles east of Thule Air Force Base). Fairly close agreement will be observed between values at Barrow and Resolute. The stations in Greenland and Antarctica all show somewhat greater visibility reductions for the same wind speeds.

The density of blowing snow decreases rapidly with height in a more or less exponential manner; so that unless there is an inversion present which forms a sharp upper boundary to the turbulent layer, it is difficult to define a depth of blowing snow. Significant concentrations probably do not occur at heights of greater than a few hundred feet, except under very strong winds.

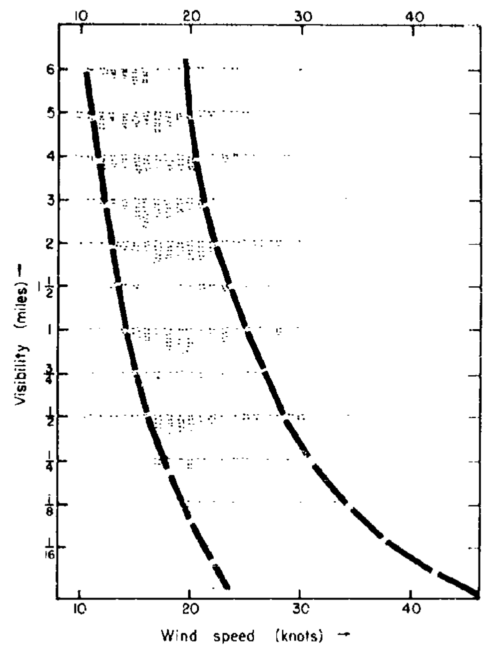



Figure 5.13. Plot of Visibility Versus Wind Speed for Barrow, Alaska.

Although there is a good average relationship between wind speed and visibility in blowing snow, wide deviations from the relationship occur in individual cases, as can be seen from figure 5.13. To understand why these deviations occur it is necessary to consider the mechanism of blowing snow in greater detail. The amount of

TABLE 5.4
Comparison of Mean Visibilities for 5-Knot Intervals at Polar Locations.

| Wind | (Visibility in miles) | | | | | | |
|---|---|---|---|---|---|---|---|
| (knots) | Barrow | Resolute | Northice | Maudheim | Site I | Site II | Thule |
| 10 | 7+ | 6+ | 7+ | --- | 2.3 | 2.8 | 6+ |
| 15 | 5.93 | 4.5 | 1.85 | --- | 1.45 | 1.9 | 4.5 |
| 20 | 2.47 | 2.0 | 0.73 | 0.81 | 1.0 | 1.15 | 2.0 |
| 25 | 1.25 | 0.75 | 0.22 | 0.25 | 0.85 | 0.65 | 0.75 |
| 30 | 0.76 | 0.4 | 0.09 | 0.11 | 0.7 | 0.25 | 0.4 |
| 35 | 0.21 | --- | 0.03 | 0.05 | 0.4 | 0.1 | --- |
| 40 | 0.07 | --- | --- | --- | --- | --- | --- |
| 45 | 0.02 | 0.01 | --- | --- | --- | --- | 0.0 |

snow carried aloft and held in suspension depends on three factors: (1) the supply or availability of loose snow at the surface, (2) the upward transport of snow by turbulence, and (3) the downward flux of snow by gravitational settling. Each of these factors in turn depends upon a number of subsidiary factors.

The supply of loose snow is determined by the nature of the snow surface (the size and shape of the particles, the bonding between them), its density, and even its configuration. Dry, fresh-fallen snow can readily be carried aloft, but as the snow ages the surface structure changes. Wind action and gravitational settling compact the snow and increase its density. The crystal structure also is altered in a manner and at a rate determined by wind, temperature, and humidity conditions. In general a snow surface hardens with time, thereby acquiring the ability to support amazing loads. Snow tractors weighing 14 tons have been known to pass over drifts 3 to 4 feet deep and leave a depression of only 1 to 2 inches.

The shape of the surface also undergoes an evolution with time, and this may have an effect on the supply of loose snow. When sustained winds blow from a single direction, the denser particles form ridges normal to the wind direction. The lighter particles settle into the troughs between the ridges where they are sheltered from the wind and consequently are no longer borne aloft.

The amount of snow transported upward from the surface depends on the strength of the turbulence, and this in turn is a function of the wind speed, stability, and surface roughness. From the well-defined relationship which exists, on the average, between wind speed and the density of blowing snow, it would appear that the speed is by far the dominant factor in determining the turbulent flux.

The gravitational fallout depends on particle size and shape, large particles falling faster than small. It is probable that the fall speed or terminal velocity of snow in the Arctic is in the range 20 to 50 cm./sec. When the snow particles are large, they are not lifted very far above the surface and consequently bounce or drift along in a process known as saltation. Hence blowing snow; that is, snow which reduces visibility at eye level; is favored by the presence of small particles.

In view of the many factors which enter into the problem, it is not surprising that the average relationship between wind speed and visibility during blowing snow is badly violated in individual cases. Until more of these factors are measured routinely and their roles defined more precisely, there is little hope of making accurate predictions of visibility reduction in blowing snow.

### 5.5.7 Forecasting Blowing Snow

Lord [5] has prepared an objective diagram for predicting visibility during blowing snow at Barrow in which wind speed forms one coordinate and the wind shift, since the last occurrence of blowing snow, the other. This diagram appears in figure 5.14. Several other variables were tested, but none improved the forecast significantly. Of necessity these were meteorological variables which, it was hoped, would provide a crude measure of some of the physical factors mentioned in the foregoing section.

From the diagram it will be noted that for the same wind speed the visibility tends to be less, the greater the wind shift since the last occurrence of blowing snow. This condition is believed to result from the sheltering effect of snow ridges. When the wind shifts the lighter snow in the troughs is scoured out at relatively low velocities. Without a shift a much stronger wind is required to dislodge the particles.

### 5.5.8 Ice-Crystal Haze

Ice-crystal haze is the name given to a phenomenon in which fine ice crystals are observed to settle earthward from a seemingly cloudless sky. This condition is distinguished from ordinary snowfall by the lack of cloud and from ice fog by the great horizontal extent and vertical depths involved. Ordinarily ice-crystal haze does not reduce visibilities below 1 to 2 miles, and the sky is usually clearly discernible through it. At night the stars are visible, and unless a beam of light causes the ice needles to scintillate the phenomenon may pass unnoticed.

Ice-crystal haze is often associated with a deep, cold <u>low</u> at high latitudes in winter, suggesting that a direct sublimation from the vapor to the ice phase is occurring in the rising current connected with the <u>low</u>. The absence of water cloud can be explained by the very cold temperatures that prevail in such cases. Occurrences of the ice haze in Labrador have been attributed to the seeding of air, supersaturated with respect to ice, by snow crystals blown aloft from the surface or by ice particles precipitated from above. In this area, the haze has been observed to extend to heights as great as 9,000 feet, al-

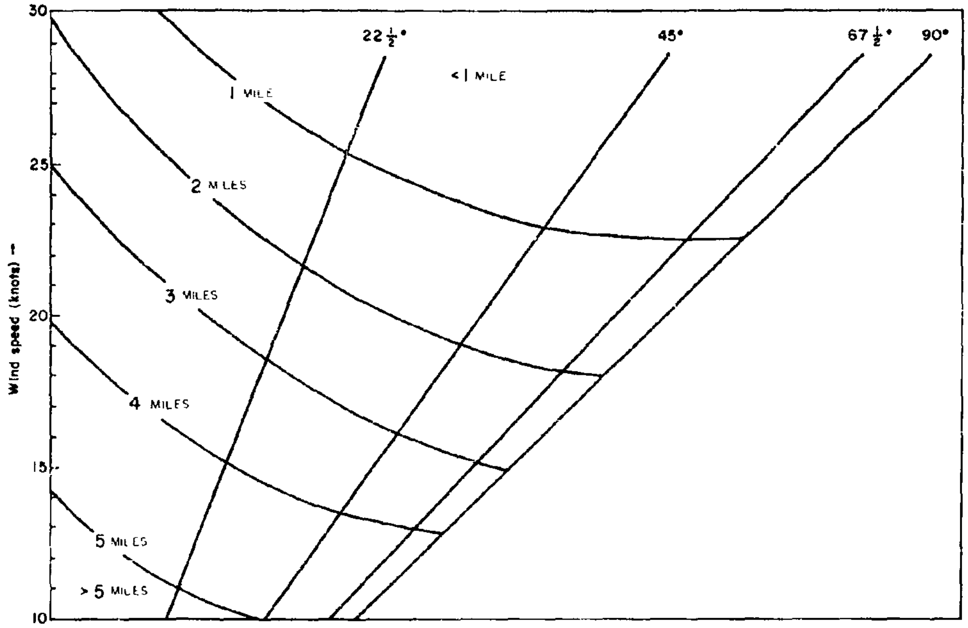



Figure 5.14. Prediction Diagram for Visibility During Blowing Snow. Sloping Straight Lines Represent Wind Shift Since Last Occurrence of Blowing Snow (in Degrees). Visibility Is Given by Curved Lines (in Miles).

though as a rule the depth is considerably less.

Until the phenomenon is better understood it will remain a difficult forecast problem. Fortunately, however, it is not an urgent problem since rarely, if ever, will the haze be sufficiently dense to reduce visibilities below flight minimums.

5.5.9 Whiteout

This condition results when sky and snow assume a uniform whiteness, making the horizon indistinguishable and eliminating the contrast between visible objects both near and far. Under such a condition the observer loses all sense of perspective, and aircraft and other operations become extremely hazardous.

The whiteout occurs most frequently in spring and fall, when the sun is near the horizon, and requires the presence of a uniform snow cover and a cirrostratus, altostratus, or stratus overcast. The sky cover is the most important factor in the prediction of whiteout, and therefore the prediction of this phenomenon is essentially a corollary of the cloud forecast.

5.5.10 Falling Snow

At some arctic locations falling snow is the most frequent cause of low visibility. However, the discussion of this problem is more properly included in the section on precipitation forecasting.

5.6 Turbulence and Icing

5.6.1 General Remarks on Turbulence

Four types of turbulence may be distinguished which are of importance to the safety of aircraft. These will be referred to here as (a) ground turbulence, (b) convective turbulence, (c) orographic turbulence, and (d) clear-air turbulence.

Ground turbulence develops in the lower layers of the atmosphere as a result of the disturbing effect of the earth's surface on the wind. The gustiness of the wind trace is a manifestation of the eddying motions which comprise the turbulence. Characteristically, the gusts are close together, indicating the presence of eddies of small dimension, but in the vicinity of certain topographical features the eddies may be of larger size and associated with substantial updrafts and downdrafts. At Thule a number of cases have been reported in which winds of 20

to 30 knots were sufficient to cause fluctuations of 100 to 150 feet in the height of aircraft on the final GCA approach. Fluctuations of this magnitude represent an extreme hazard, and on at least one occasion it was necessary for the pilot to employ full take-off power in order to land safely.

Since a high degree of static stability tends to inhibit the development of vertical motions, it is apparent that turbulence forms more readily with unstable than with stable lapse rates. Once the turbulence has formed the strong mixing will generally produce a near-adiabatic lapse rate, whatever the initial conditions.

In summary, high wind speed, low static stability and large roughness are the major factors in the development of ground turbulence. Because of the prevalence of light to moderate wind speeds and very stable lapse rates, ground turbulence is not normally an important problem in the Arctic. However, during the infrequent occurrences of high winds, described in section 5.12, the turbulence may become severe.

Convective turbulence is encountered in connection with the often violent updrafts and downdrafts of convective cloud cells. The thunderstorm is the most frequent seat of the more violent occurrences. In view of the relatively weak nature of convection in the Arctic and the relative scarcity of thunderstorms, it follows that convective turbulence is a problem of much less importance in polar regions than in the middle latitudes and the Tropics. However, occasional instances of severe turbulence have been reported in connection with the summertime thunderstorms that develop in the mountainous sections of Alaska and the Yukon, so that it would be foolhardy to ignore completely the dangers from this source.

Data collected on the Ptarmigan Weather Reconnaissance Flights reveal that turbulence is almost entirely lacking over the Arctic Ocean, as might be expected from the stabilizing effect of the underlying surface. The few reports of light turbulence that are reported appear, almost always to be associated with clouds and, hence, must presumably be ascribed to convection aloft. In the case mentioned previously, in which the flight encountered a thunderstorm over the Beaufort Sea, moderate turbulence was reported.

Orographic turbulence occurs in the vicinity of mountainous terrain when certain critical values of wind speed, vertical wind shear, and static stability are met. The motions which produce the turbulence are frequently in the form of organized waves over and to the lee of peaks and ridges and, therefore, are not turbulent in the strict sense of the word. However, as used here the term relates to the effect of the motions on aircraft, and in this respect the mountain waves are turbulent. Moreover, a truly chaotic turbulence usually appears in association with the waves, sometimes as a result of a breakdown of the wave motions themselves.

Under proper humidity conditions the waves are made visible by a cap cloud over an isolated peak, or by a foehn wall along a ridge, and by lenticular or lens-shaped clouds in the lee of the obstacle. The presence of a rotorcloud--a roll-shaped cloud with a horizontal axis of rotation--warns of the danger of extreme turbulence.

All mountainous areas of the Arctic are potential sources of orographic turbulence, and, indeed, pilots have reported instances of moderate to strong turbulence in the neighborhood of a number of arctic ranges. Through experience, the forecaster can acquire valuable knowledge concerning areas of maximum danger and the wind conditions connected with the outbreak of turbulence in these areas. For example, forecasters in the Whitehorse region of Canada have noted the tendency for strong downdrafts to develop near the Wolf Range when southwest to west winds of greater than 50 knots blow across the St. Elias Range.

Clear-air turbulence occurs most commonly in the zones of large vertical wind shear, above or below strong jet streams, and in the zone of large horizontal shear on the cyclonic side of strong jets. The turbulent regions are often found in corridors about 30 miles across and 1,000 to 2,000 feet deep. As yet, the cause of clear-air turbulence is not fully understood. Few, if any, cases of significant clear-air turbulence at high latitudes are reported in the literature. However, the possibility definitely exists since strong jet streams occasionally invade the Arctic as described in section 3.3 of chapter 3.

### 5.6.2 Forecasting of Turbulence

Ground turbulence is so dependent on local surface and terrain features that one cannot hope to obtain universal forecast rules. In constructing an objective forecast diagram for a specific location, wind direction and speed should be chosen as principal predictors. Stability may be an additional parameter of importance.

The prediction of convective turbulence is part of the convective cloud and thunderstorm problem which has already been discussed in a previous section.

Orographic turbulence, like ground turbulence, is a highly local phenomenon and usually is found to be most highly related to the wind component normal to the pertinent mountain range. According to both theory and observation, the static stability and vertical wind shear are also important factors in the development of mountain waves.

From the evidence presently available, it appears that clear-air turbulence is associated with large vertical wind shear of either sign, low static stability, and large positive or cyclonic shear in the horizontal. The Richardson number, which is a function of both stability and vertical wind shear, has often been suggested as a more fundamental criterion for the development of the turbulence. In view of the foregoing relationships, clear-air turbulence should be forecast for the regions near and to the cyclonic side of pronounced jet-stream maxima.

5.6.3 Icing

Icing occurs when supercooled cloud droplets freeze and adhere to aircraft surfaces. The ice coating changes the aerodynamic properties of the airfoils and may lead to a dangerous loss of lift.

The fact that icing receives little mention in discussions of arctic flying conditions may be regarded as an indication that it is not a serious problem in arctic flight operations. This relative unimportance of icing may be explained by the comparatively small moisture content of arctic air and, hence, the low liquid water content of arctic clouds. The liquid water content, of course, is the principal factor in determining the severity of the icing. Furthermore, in winter, temperatures are often so cold that clouds cannot exist in liquid form.

Data compiled from the records of the Ptarmigan Weather Reconnaissance Flights show that at 10,000 feet icing occurs only 2 percent of the time over the arctic seas. The corresponding figure for flights over the North Atlantic is 19 percent. Because of the smallness of the figure for the Arctic, it should not be assumed that icing is never a hazard in polar regions. Heavy glaze ice may be encountered, occasionally, in maritime currents which invade the Arctic from the Atlantic and Pacific sectors. Moderate to severe rime icing also occurs on occasions. At Fairbanks a mixing ratio, at flight level, of 2 grams per kilogram has been found to mark the beginning of moderate rime icing.

When cold air passes over open water, a low cloud with large concentrations of supercooled droplets may form. Moderate to heavy icing has been observed in clouds of this type, but fortunately the cloud deck seldom extends above 4,000 feet so that it may be easily topped.

The icing forecast hinges on the cloud and temperature forecasts. Where cloudiness is predicted within the temperature range -22° F. to 32° F. (-30° C. to 0° C.) the possibility of icing must be considered.

5.7 Optical Phenomena

Besides the phenomenon of whiteout, described in section 5.5.9, a number of other optical phenomena are of importance and interest to the arctic traveller. Two types will be discussed here: distortions of distant objects caused by abnormal bending or refraction of light rays and reflections on the underside of clouds known as iceblink and water sky.

5.7.1 Looming

It is well known that the speed of light is slightly less in the atmosphere than in outer space and that the reduction in velocity increases with the air density. As a consequence of this and the normal decrease of density with height, a wave front emanating from an object will travel faster in its upper portion than in its lower portion, and the rays of light will be bent or curved in the downward direction. This arching of the light rays causes objects to appear slightly elevated above their true positions and allows the observer to see somewhat beyond the line-of-sight horizon.

When the density decreases more rapidly than normal with height, as occurs under conditions of temperature inversion, distant objects appear to be unusually elevated, and objects normally below the horizon may come into view. The term looming is applied in this circumstance. No doubt many erroneous estimates of distances by early explorers can be ascribed to this phenomenon.

Often when looming is present, the rays from the upper portion of an object are bent more than those from the lower. In this case the object appears stretched, as well as elevated, and the

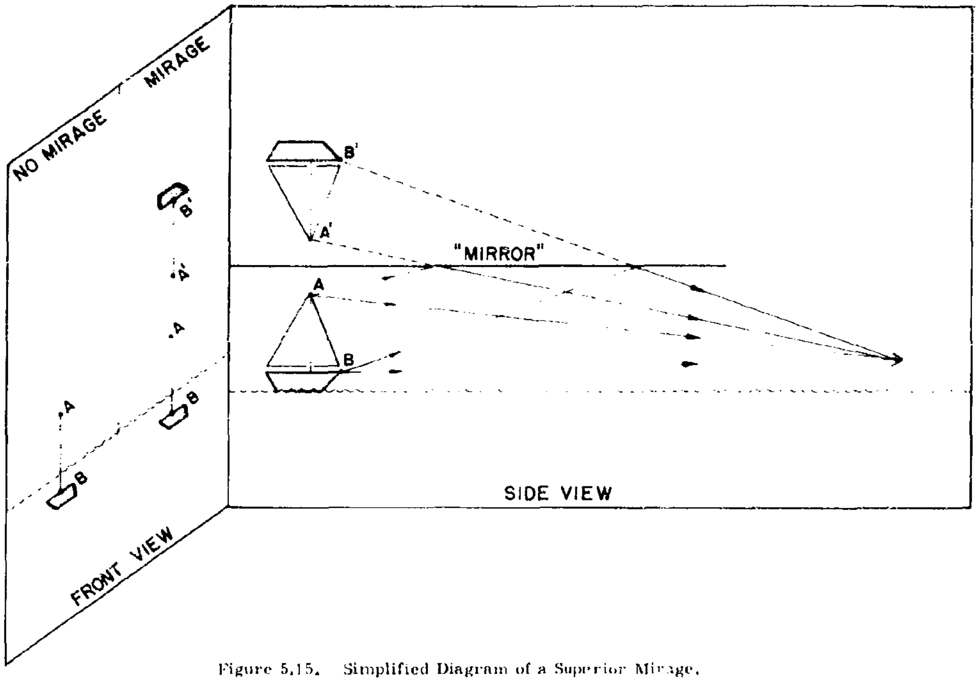



Figure 5.15. Simplified Diagram of a Superior Mirage.

term towering is used to describe the situation. It is also possible for the rays from the lower portion to be more curved, in which case shrinking or stooping of the object is observed.

5.7.2 Sinking

When the air density increases with height, light rays curve upwards, causing objects to appear below their normal elevations and even to sink from view below the horizon. This condition is of less frequent occurrence in the Arctic than its inverse condition, looming, since it requires an abnormally steep lapse rate. However, when very cold air passes over open water, sinking may be observed, provided steam fog does not develop. It may also occur over strongly heated interior areas in summer.

5.7.3 Superior Mirage

An elevated inversion, if sufficiently pronounced, may cause such a sharp downward bending of light rays that the inversion layer may be likened to a mirror, suspended in the sky, which reflects light from objects located beneath it. Under such a condition an observer may see an inverted image above the object. With complex inversions the wave front may suffer multiple distortions and a farther, upright image may appear above the inverted image.

The origin of the superior mirage is illustrated in figure 5.15. For the sake of illustration it has been necessary to exaggerate greatly the bending of the rays. In reality the refraction is much smaller than shown so that only distant objects can be reflected from the "mirror."

5.7.4 Inferior Mirage

Under superadiabatic conditions the upward bending of light rays near the ground may be so extreme that one may replace the ground at some distance from the observer by an imaginary mirror. Thus the observer will see an inverted image beneath a distant object. Since the "mirror" will reflect the sky above and beside the object, the observer will also have the illusion that the object is surrounded by a lake or body of water. This condition is illustrated in figure 5.16.

It will be noted that the inferior mirage is accompanied by sinking, the horizon appearing closer than normal to the observer. As in the previous illustration the curvature of the earth is neglected. The effect of curvature is to cause the base of the object to disappear from view so that only the upper portion of it, and the inverted

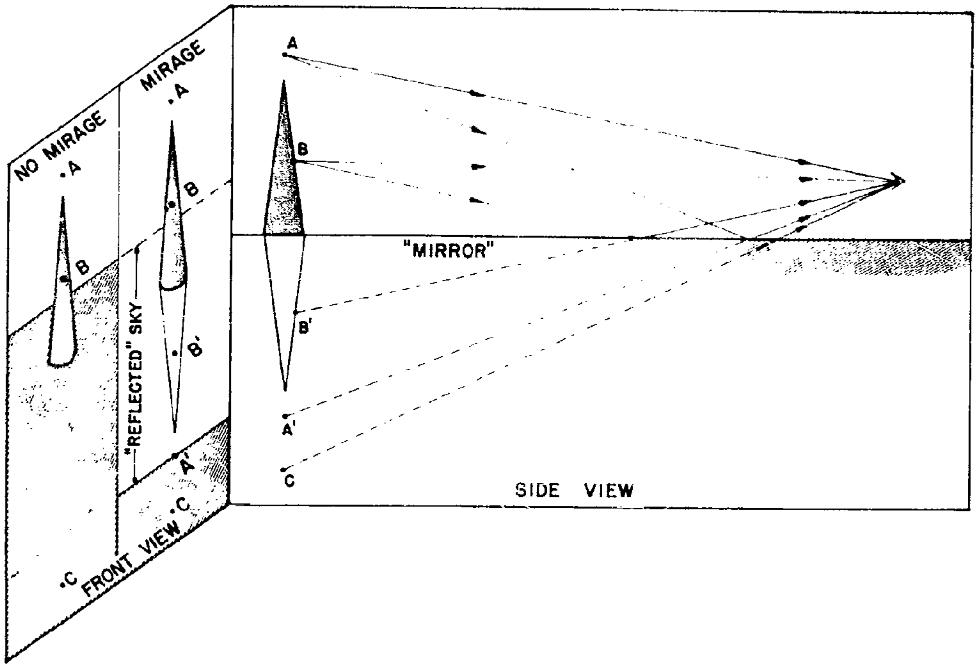



Figure 5.16. Simplified Diagram of an Inferior Mirage.

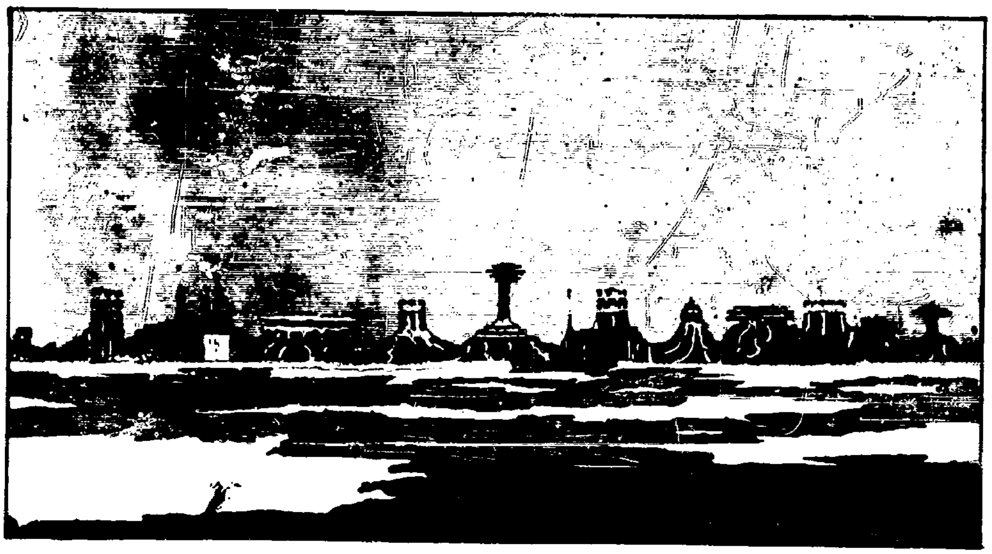


Figure 5.17. An Example of the Fata Morgana. East Coast of Greenland at a Distance of 35 Miles, 18 July 1820. (From: Petterssen, Jacobs, and Haynes [11].)

image of that portion, are seen. Because of the difference in curvature of rays emanating from the top and bottom of the object the image is ordinarily shortened or stooped.

Generally the inferior mirage is blurred by a phenomenon known as optical haze or shimmer. This arises from the convective activity that accompanies the superadiabatic lapse rates. Bubbles of warm, less dense air rise and are replaced by cooler, denser air from aloft. Consequently, rapid variations of density and, hence, refractive index occur.

It also should be mentioned that the appearance of mirages, both superior and inferior, is much affected by the distance of the object and the relative positions of observer, object, and layer of abnormal density variation. A slight change in any of these factors may cause pronounced changes in appearance.

5.7.5 Fata Morgana

This is the name given to a complex mirage in which distant objects become greatly elongated in the vertical direction and take on a bizarre aspect. A shoreline may be drawn out into tall cliffs and columns, and houses near the shore may assume the appearance of wondrous castles.

The phenomenon occurs under much the same meteorological conditions as the superior mirage and indeed often contains many of its features, though in more distorted form. The unusual stretching takes place in the layer where the normal upward lapse of temperature ceases and the elevated inversion begins. At the cold point in the sounding the density is at a relative maximum, and the degree of bending of the light rays changes sharply. The result is a vertical magnification of objects in the layer.

The fata morgana may be regarded as an intermediate stage between towering and the superior mirage or a mixture of the two conditions. This may be seen in the example in figure 5.17, where both the stretching and the tendency for formation of inverted images are observed.

5.7.6 Iceblink and Water Sky

In summer a white or yellowish-white glare may be seen on the underside of clouds, as a result of reflection of sunlight from snow or ice fields. When these reflections are intense they are referred to as iceblink. Conversely, under the same circumstances dark patches or streaks may appear on the cloud base above areas of open water. This condition is known as water sky. When other means of reconnaissance are not available these phenomena are of assistance in navigating through the ice of the polar seas, since they give at least a rough idea of ice conditions at a distance.

5.7.7 Forecasting of Optical Phenomena

Rarely is there any call to predict these phenomena. It is obvious that the likelihood of mirages and other refractive phenomena is determined by the lapse rate conditions expected to prevail in the lower layers of the atmosphere. Iceblink and water sky require a low sun and a deck of low clouds for their formation.

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