3. ARCTIC CLIMATOLOGY

Knowledge of climatic characteristics of a region helps the new forecaster by providing information on the general types of weather systems to expect in day-to-day forecasting situations. Armed with such information, along with good analyses, numerical model forecast aids, and frequent high quality satellite data, the forecaster, through prolonged onsite experience, should be able to develop successful prediction techniques. Use of forecaster handbooks, such as this one, can reduce the onsite time required to become a proficient forecaster.

This chapter will first discuss the major factors that go into determining the climate of the Arctic. Second, individual weather elements will be considered in view of their regional and seasonal variations. Third, local topography and its influence on climatic conditions will be examined. In Chapter Two the physical characteristics of the Arctic regions were described and shown in various locator maps in order to relate them to the profound effect they can have on local atmospheric pressures, temperatures, winds, clouds, and precipitation. Perhaps the most conspicuous causes of these local effects include (1) the Brooks Range and North Slope of Alaska; (2) the vast, elevated ice sheet of Greenland; (3) the land/water complex of the Canadian Archipelago; (4) the fjords of Greenland and Norway; (5) the mountains of eastern U.S.S.R.; and (6) the pack ice and Arctic Ocean itself.

The four characteristics of the Arctic that are primarily responsible for its weather and climate are summarized in the following subsection. These indicators are (1) duration of daylight; (2) the circumpolar vortex; (3) surface snow and ice cover; and (4) strong temperature inversions.

3.1 Major Factors Affecting the Arctic Climate

3.1.1 Duration of Daylight

The regime of daylight and darkness in the polar regions combined with the low Sun angle ensure long periods of heat loss. North of the Arctic Circle in summer, the Sun is always above the horizon, albeit at various elevations. In winter, the Sun is always below the horizon, but there too, the degree of darkness varies throughout the period. Only at the pole itself are there 6 months of Sun below and 6 months above the horizon. Table 3-1 shows the duration of sunlight for various latitudes at the middle of the months. The figures apply to level, unobstructed locations and may differ for stations on hills and in valleys.

TABLE 3-1. TOTAL POSSIBLE DURATION OF SUNLIGHT ON THE 15TH OF EACH MONTH

Month	North Latitude											
	60°		65°		70°		75°		80°		85°	
	hr	min	hr	min	hr	min	hr	min	hr	min	hr	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	50
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	56	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	–	–	–	–	–	–	–	–

An alternate way of looking at the duration of sunlight is shown in Fig. 3-1 below.

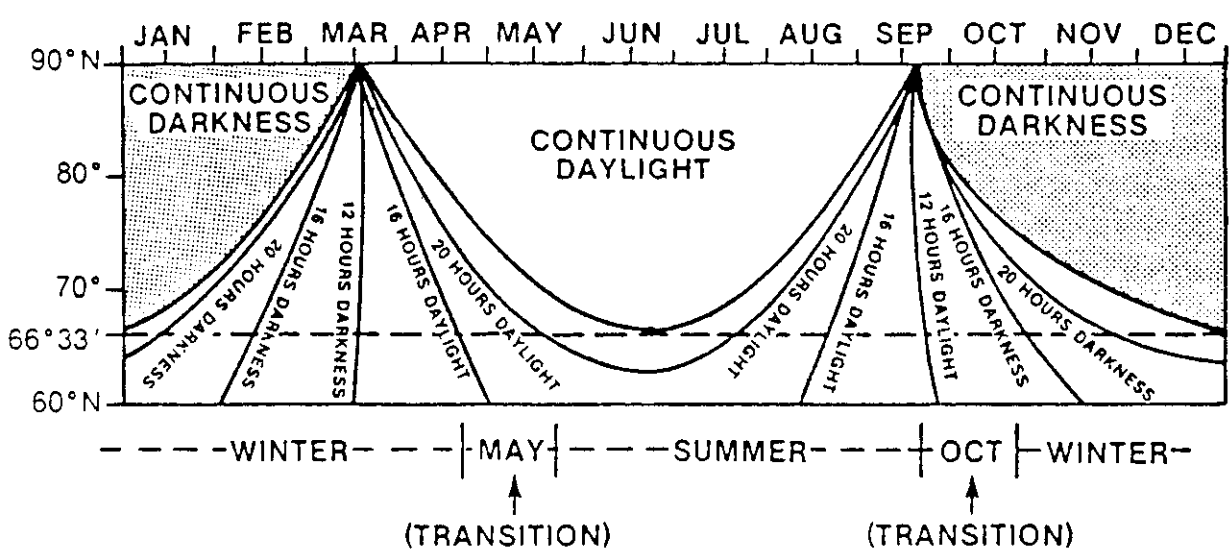

Figure 3-1. Duration of Daylight 60°N to 90°N (Welsh et al., 1986).

The notions of daylight and darkness can be misleading. For example, at the pole, only about 80 days of real night occur. For 20 days after the setting of the Sun at the pole on 24 September, a newspaper can be read by twilight if the skies are clear. Complete darkness does not occur until 12 November. The period of total darkness ends on 30 January, and the twilight grows progressively brighter until the reappearance of the Sun on 19 March.

Note the difference in the definition of civil as opposed to astronomical twilight. The former is said to occur when the Sun is less than 6 degrees below the horizon; the latter when it is less than 18 degrees below.

Another factor is that the Earth is closer to the Sun in the Southern Hemisphere winter than in the Northern Hemisphere winter. As a result, the polar day, defined as the time interval between the last sunrise in spring and the first sunset in fall, is shorter in the south polar regions. At the North Pole only one sunrise and one sunset occur per year, the duration is 189 days for the polar day and 176 days for the polar night; at the South Pole it is 183 and 182, respectively.

Figure 3-2 shows the reason for the polar region heat loss. The poles receive solar heat throughout the year, but the loss by infrared cooling results in a net heat loss poleward of 38 degrees latitude.

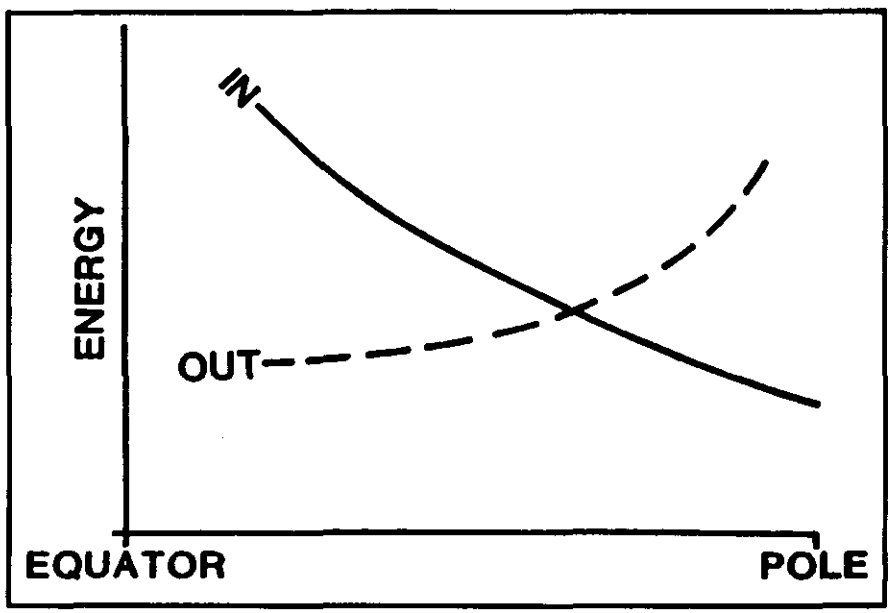

Figure 3-2. Net Radiational Heat Loss at the Poles.

3.1.2 The Circumpolar Vortex

The dominant feature responsible for the Arctic's unique weather is the cold core, circumpolar vortex in the middle and upper atmospheres. The strongest north-south temperature gradients appear in midlatitudes where the mean jet stream occurs. The westerly jet streams have their maximum wind speeds between 30,000 ft (9,144 m) and 40,000 ft (12,192 m). The jets steer surface cyclones and anticyclones and are also responsible for their formation, location, and dissipation. Weather variations at the surface are largely determined by the position and strength of the polar vortex and its associated jet streams.

The areas below the periphery of the vortex typically experience eastward moving frontal cyclones. Areas directly under the vortex are characterized by old, occluded, and stagnant lows with very cold air in them. The location of these cold air pockets aloft determines where the static stability of the underlying troposphere will be minimal and lead to convective-type cloudiness.

The circumpolar vortex displays a multiwave pattern that reflects the influence of land forms, oceans, and mountains of the Northern Hemisphere. The Southern Hemisphere with its lack of these topographic features supports a much simpler circulation pattern. In any case, the continent-ocean temperature contrasts of the Northern Hemisphere result in several major troughs in the vortex proper. Five-day mean 700-mb charts for each month are extremely useful in describing the circumpolar vortex throughout the various seasons. Such charts appear in Appendix B, but for purposes of the present discussion, one such chart, representative of mid-January, is given as Fig. 3-3. In winter, three troughs appear: one over the Canadian Archipelago, one passing from Novaya Zemlya southward, and one over Siberia curving past Kamchatka and extending southward into the Sea of Japan. A minor trough extends east-southeast from Kamchatka into the Aleutians. The main ridges lie over the eastern Pacific and the Rocky Mountains, and over the eastern Atlantic and Scandinavia, with a third weaker ridge over the Urals and western Siberia.

The north-south temperature gradient relaxes during the spring and summer causing the circumpolar vortex to contract and weaken until a single center lies over the pole. When this occurs, the jet stream retreats northward, taking with it the principal cyclone tracks. Cyclone frequency diminishes during this period of less baroclinicity as more zonal (westerly) flow replaces the largely meridional (north-south) flow of winter. Then, again in the fall the vortex expands and strengthens, and the flow around it becomes shifted to lower latitudes and is more meridional.

In the more active winter season, the circumpolar vortex is an ever-changing, dynamic center of action. As the wave number increases (more troughs and ridges appear), cyclogenesis is more frequent, and a greater variety of weather is noted. In this chapter the climatology of storm tracks will be discussed. Also, the possibility of blocking (large-scale obstruction of the normal west-to-east progress of migratory cyclones and anticyclones) action can cause a persistence of certain weather systems over a given region.

In summary, Sater (1969) has noted some general features of the circumpolar vortex and the accompanying circulation as follows:

(1) Large-scale patterns tend to persist. Regionally these stalled systems can result in extended periods of abnormally cold or mild temperatures, drought, or heavier than usual snowfall, high winds, or other extreme conditions.

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

(3) The degree of expansion or contraction of the vortex can have serious effects on Arctic operations by causing abnormal displacement of the upper westerlies.

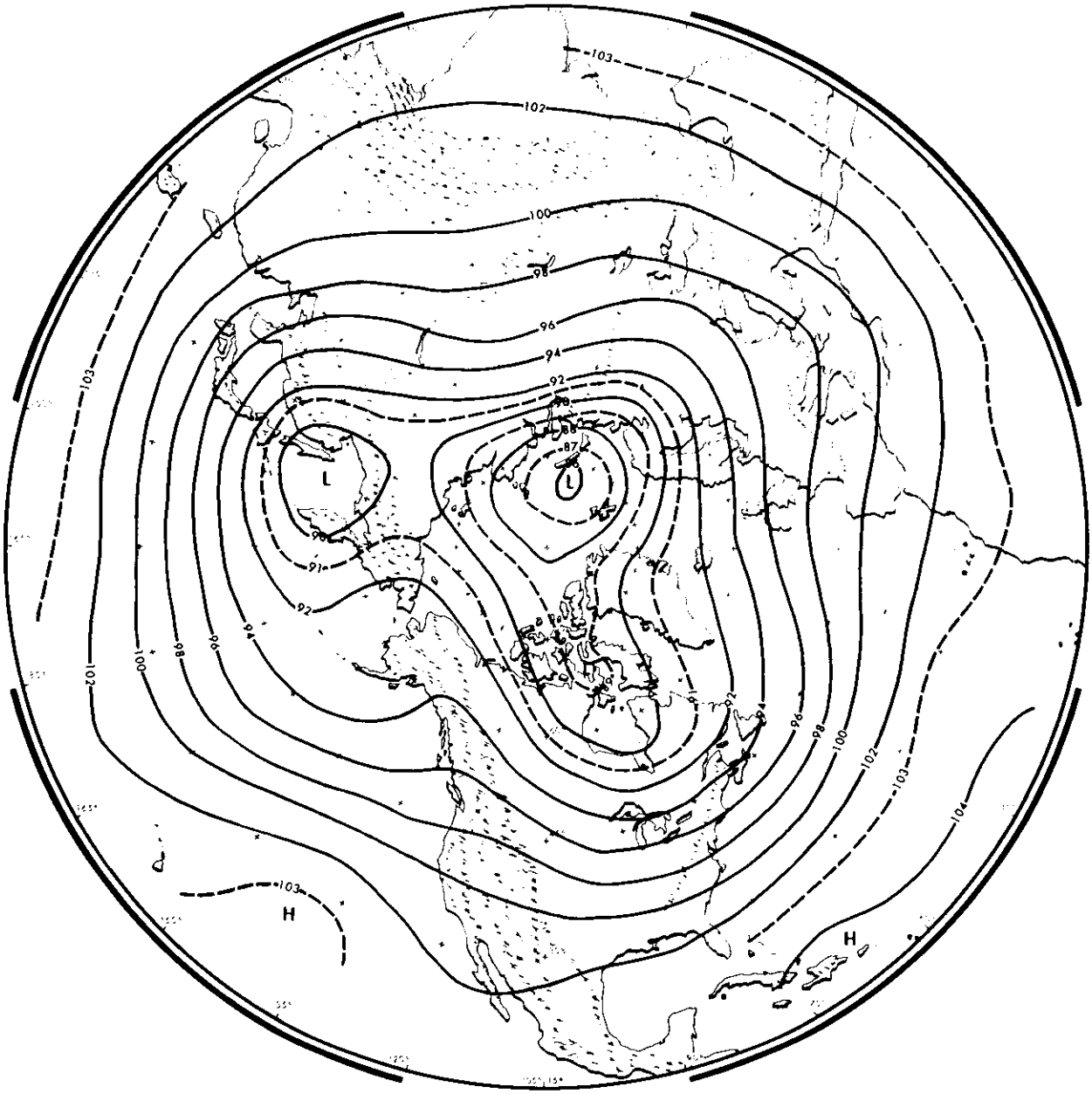

Figure 3-3. Mean 700-mb Chart for Mid-January (Wahl & Lahey, 1969).

3.1.3 Surface Snow and Ice Cover

The amount of snow and/or ice cover in any region of the Arctic determines, to a large degree, the radiation budget for that region. Solar absorption over snow-free land (low albedo) may be as high as 80 percent, and as a result, temperatures in summer may easily rise to 75 °F or 80 °F (24 °C or 27 °C). Over the ice pack the solar absorption is much less, (high albedo) perhaps 20 percent. This characteristic and the lack of enough heat input to change phases help to keep temperatures over the ice pack near 32 °F (0 °C).

At the marginal ice edges temperature and moisture gradients will often be quite large. Thus, in these areas greater than normal precipitation, cloudiness, fog, and even altered cyclone tracks may be expected. These large gradients are also present near open leads and polynyi, which are also significant sources of heat and moisture. In general, open water produces a moderating effect on local temperatures. By contrast, snow-covered land and ice tend to maintain extremely cold temperatures.

Although the actual amount of snowfall may be small due to the lack of water vapor at low temperatures, average depths of the maximum snow cover in March or April ranges from 8 to 20 in (20–50 cm) over the frozen oceans to 16 to 28 in (40–70 cm) over the subarctic.

The most spectacular changes from winter to summer occur over the interior landmasses, but they are primarily in the subarctic. For landmasses north of 60 °N, the spring melt begins first in the region of Scandinavia, which is to be expected because of the warm branches of the Gulf Stream entering the area. The melt spreads into western U.S.S.R., and, at the same time, the west coast of Greenland shows signs of reduced snow cover. By late spring practically all of the Arctic land areas show the effects of warming on the snow cover, and only the icecap of Greenland, Labrador, and north-central Siberia show significant remaining snow cover. By August the snow cover is an absolute minimum in the Arctic region. In the following weeks it will increase to its typical winter distribution.

The net effect of the variations in snow cover is reflected by the fact that the transitional periods between winter and summer are very short in the high Arctic. Spring and fall typically last only a few weeks.

3.1.4 Temperature Inversions

Unique to the Arctic and Antarctic regions are the spectacular temperature inversions that can frequently occur in certain special locations. In extreme cases temperature increases of 20 °C to 30 °C in the lowest 330 ft (100 m) have been recorded by sounding devices. The reason, of course, for these spectacular inversions lies in the nature of the polar regions themselves.

Extreme inversions occur when the lowest levels of the atmosphere experience strong cooling. This cooling is most favored under the following conditions: (1) Nighttime radiational cooling of the surface; (2) light surface winds, which preclude mixing down of warmer air aloft; (3) clear skies, which maximize the radiational cooling; and (4) snow cover surfaces, which also enhance the cooling because snow is a good radiator to space. All of these conditions can be optimized in the polar regions because of the length of the polar night and the presence of all requisite factors for long periods of time. Surface inversions are not as strong over sea ice compared to land stations due to the heat flux through the ice from the warmer ocean.

As will be discussed subsequently, these extremely strong inversions have many implications with regard to Arctic weather. To begin with, the presence of the inversion ensures that the low levels of the atmosphere have a good deal of static stability. This stability, in turn, precludes mixing and contributes to the entrapment of fog, pollutants, and/or low clouds under the inversion. Convection is also suppressed. Inversion induced refractive phenomena of the Arctic have serious impact on radio communications throughout the region. Figure 3-4 presents a typical Barrow, Alaska sounding, showing extreme surface cooling.

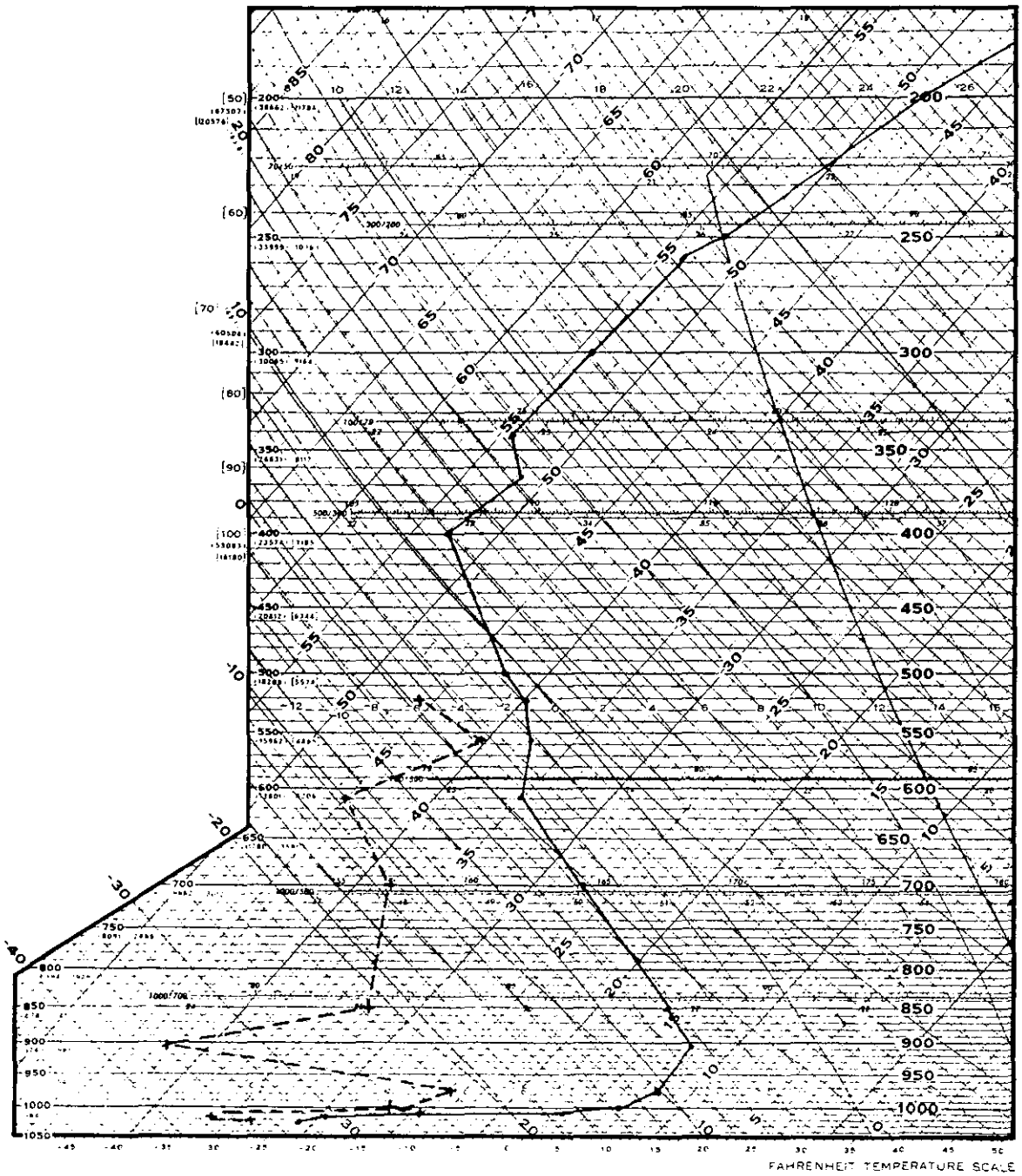

Figure 3-4. Typical Sounding from a Station in the Arctic.

3.2 Individual Atmospheric Elements

3.2.1 Fog

Arctic weather stations experience fog conditions on the order of 100 days per year, mostly in summer. During periods of the year where some daylight is present, fog will usually occur during the early morning and nighttime hours. The regions bordering the

East Siberian and Laptev Seas present an exception. Here fog frequency is greatest around midnight. At noon the fog possibility is least. Low-level mixing in the boundary layer over the Barents and Norwegian Seas makes these areas the least foggy in summer.

Fog can be caused by a number of different mechanisms. Knowledge of these mechanisms aids the forecaster in anticipating the likely occurrence of fog at various locations. The four main fog types are

Advection Fog. Advection fog forms in summer when warm, moist air flows over the melting pack ice and cold, open water. High wind speeds will normally cause the fog to lift, resulting in low stratus. Advection fog is usually patchy and lasts only a relatively short time. June and July are the months of highest probability of advection fog. At present few studies have been made of the climatology of fog occurrence in the Arctic. The areas around southern Greenland and Iceland have received, however, considerable study because of the proximity to shipping lanes. The studies suggest that these areas are subject to considerable advection fog, and the probabilities of occurrence are highest in summer when the temperature contrast between ocean and air is greatest.

Radiation Fog. Radiation fog occurs most frequently with clear, calm conditions over snow and/or ice covers. The surface layer of air is cooled by outgoing radiation during the night, temperatures are reduced to the dewpoint, and condensation occurs. These fogs can occur over the pack ice, but they are most frequently found along river valleys where air movement is minimal and drainage renders the air quite cold. They are most common in the lower Lena and Mackenzie River Valleys. Radiation fogs may also occur in the Yukon Valley and in some of the valleys of northward flowing rivers in Siberia. Radiation fogs are generally thin and shallow and burn off in the daytime. In the lower Mackenzie 20–25 fog days may occur in a month. The radiation fog is also a good indicator of a strong low-level inversion that inhibits vertical mixing. As may be expected, radiation fogs are most common in fall and early winter.

Steam Fog. Steam fog develops when very cold, dry air comes into contact with a relatively warm water surface (leads, polynyi, etc.), causing a rapid transfer of heat and moisture from the water surface to the air. The heating from below produces an unstable lapse rate, and associated small scale convective eddies carry the warm, moist surface air aloft where it mixes with the colder, dryer air above. Condensation occurs in the form of fog, but with further mixing of the dry air the fog will dissolve. Steam fog is sometimes referred to as Arctic sea smoke.

Ice Fog. The fogs mentioned previously are usually water fogs. Ice fog occurs during the winter when large amounts of water are added to extremely cold air ($< -22\,°F$ or $-30\,°C$). The resulting light ice particles (crystals) settle slowly and therefore remain suspended in the air for long periods of time. These ice fogs are common around inhabited areas where a variety of moisture sources can be found. During the Coordinated Eastern Arctic Experiment (CEAREX) in the MIZ, a mild version of ice fog, which had little impact on visibility but which was almost always evident, was referred to as "diamond dust."

3.2.2 Clouds

As in other parts of the world, Arctic clouds are produced by one of several cooling mechanisms: (1) Dynamic lifting in low pressure systems or at convergence lines and fronts; (2) Radiational cooling, which is responsible normally for low clouds and fog; and (3) Orographic lifting, which produces clouds in the vicinity of land and sea features conducive to forced ascent and subsequent cooling of the air to condensation. The following characteristic cloudiness features are found in several different regions of the Arctic.

The Norwegian Sea. Because of the warm waters of the North Atlantic Drift current, wintertime cloudiness in this region is generally of the convective type. Cold outbreaks from the north in winter, when reaching the warm waters off the Norwegian coast, result in a destabilization of the air such that cumulonimbus clouds are possible. Despite this, cumulus cloudiness is less pervasive than stratified clouds of summer and, therefore, winter is less cloudy than summer. Cloud cover in this period reaches only up to 40 percent, whereas in summer it can reach 80 percent. Since less cyclonic activity occurs in summer, the clouds tend to be middle clouds.

East Siberian Type. In winter the central U.S.S.R. can be expected to be dominated by anticyclonic conditions. Because of the moisture supply, however, the clearest areas are east of 100 °E at around 140 °E. The cloud type is generally cirrus about 40 percent of the time. Middle clouds occur only with rare cyclonic activity, and stratocumulus under inversions occurs only infrequently. As in the Norwegian Sea area, cloudiness is greatest over the eastern Siberian area in summer. At this time anticyclonic conditions are less frequent, and open seas are a supply of abundant moisture. This situation tends to produce low and middle convective cloudiness at this time. Spring and fall are not similar in terms of cloudiness because of the rapid freezeup in the fall that ties up the moisture supply. The gradual thaw of spring assures longer lasting or persistent cloudiness.

Polar Ocean Type. Because of the extensive ice cover in winter, cloudiness over the Arctic Ocean is least in winter and greatest in summer here also. In summer a high frequency of stratus and stratocumulus occurs. It may reach 80 percent during this season, as relatively warm air condenses by cooling due to the ice cover of the ocean below. These stratus clouds often exceed 1,500 ft (457 m) in thickness. The water content of these clouds diminishes from the coast to the pole, naturally. The seasonal change in cloudiness is quite rapid during the short transition periods of spring and fall.

Figures 3-5 and 3-6 present cloud cover in percentages for winter and summer, respectively. The major differences are (1) increased cloud cover from winter to summer over the Greenland icecap; (2) a general increase in cloud cover over the East Siberian Sea from winter to summer; and (3) an increase from winter to summer over the Canadian Archipelago.

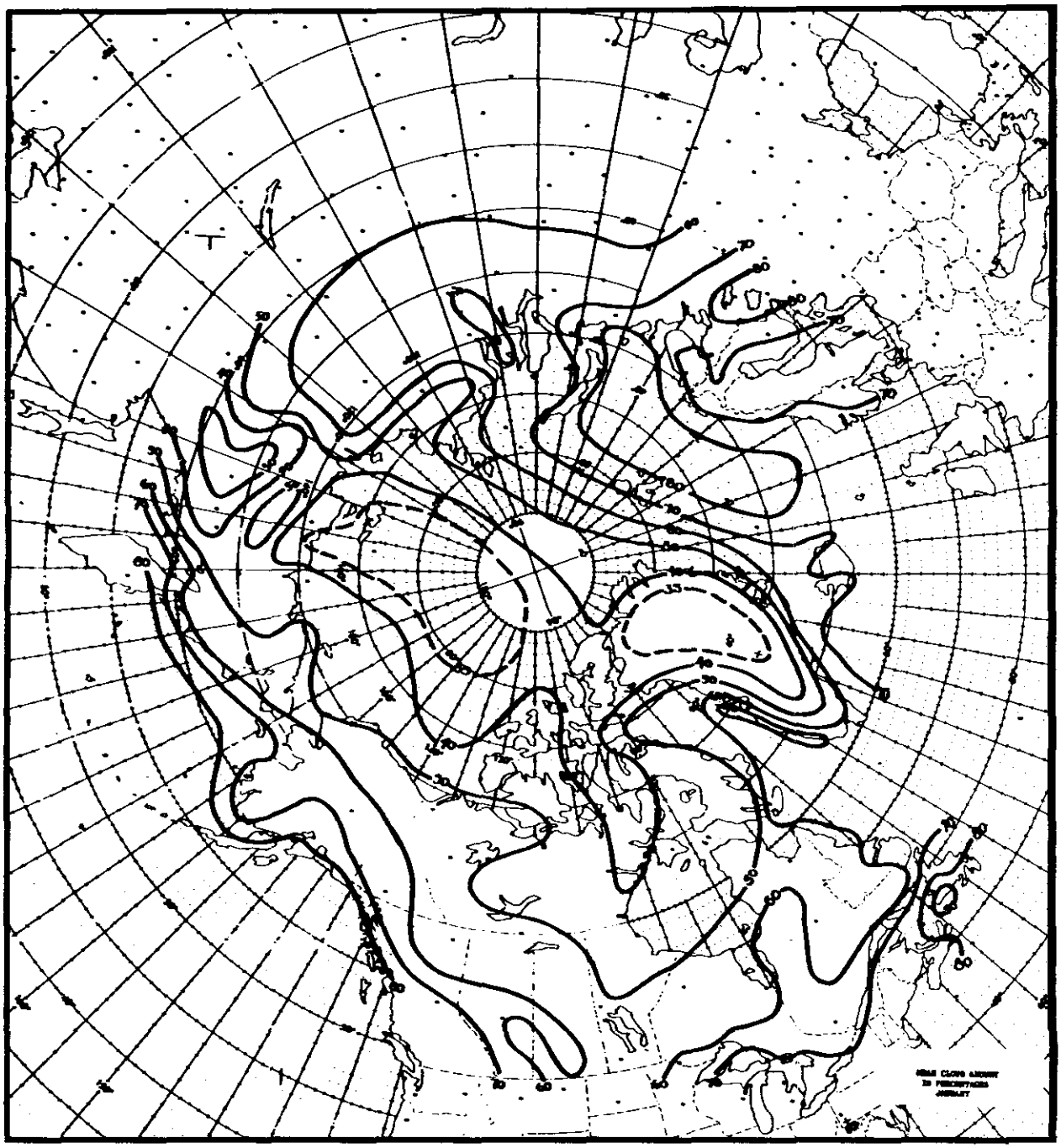

Figure 3-5. Winter Cloud Cover in Percentages (Navy Arctic Manual, 1970).

3.2.3 Precipitation

In winter, Arctic snowfall is light primarily due to the low moisture content of cold air. The annual precipitation over the Arctic Ocean itself is a meager 5 in (13 cm). Along the margins of the Arctic Ocean, where more measurements are taken, the winter accumulation is less than 10 in (26 cm). The coasts of Siberia and the Canadian Arctic

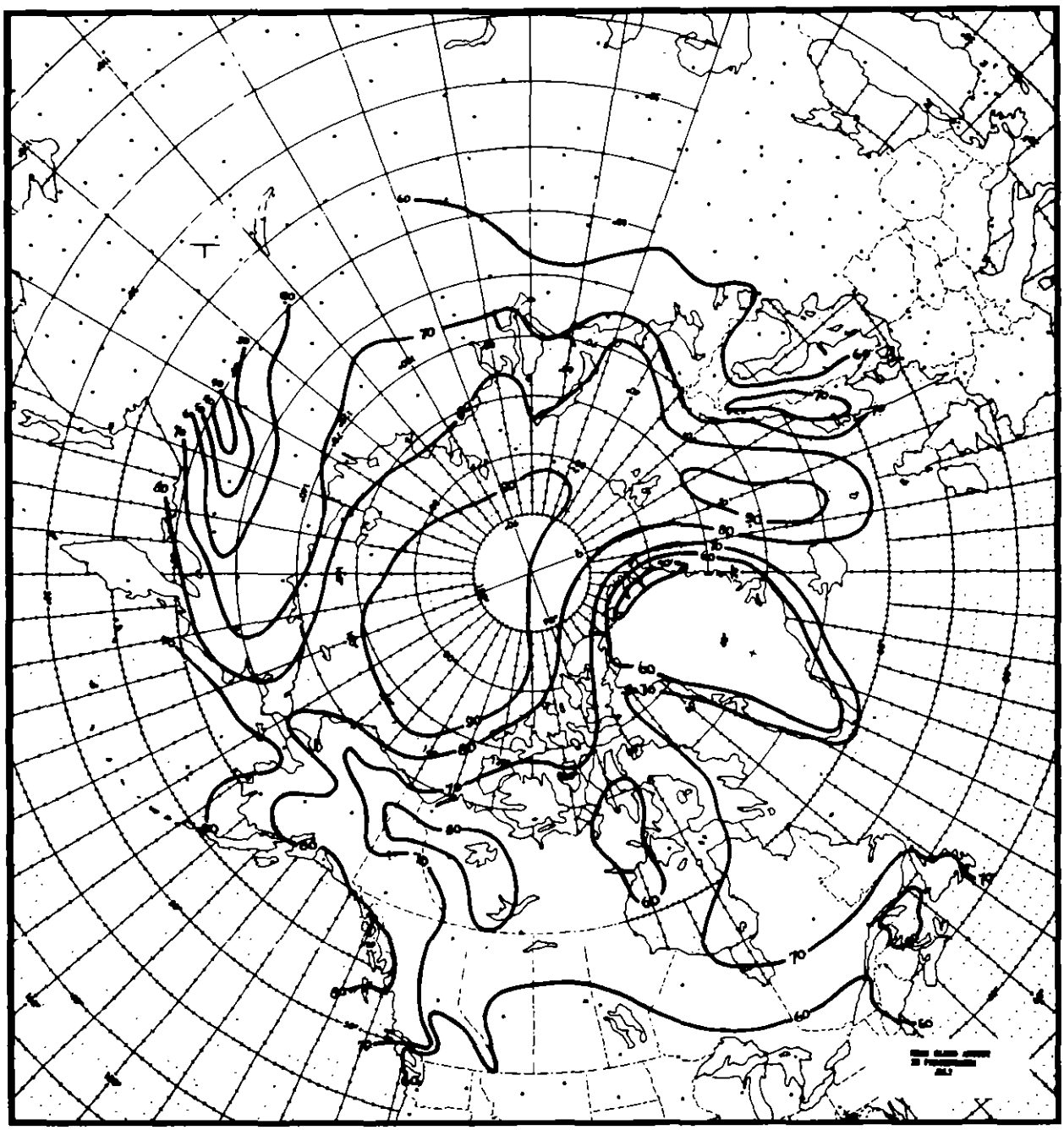

Figure 3-6. Summer Cloud Cover in Percentages (Navy Arctic Manual, 1970).

receive as little as 5.5 in (14 cm) per year, including rain and the water equivalent of snow. This precipitation is mainly snow and falls mostly in the late spring and early autumn and is due primarily to frontal systems. Frequent light snow showers will occur near leads and polynyi, and locally heavy snow showers will occur when severe cold air is aloft.

Although the amount of precipitation along the Arctic coasts is small, the ground remains saturated for long periods in summer. This condition is a result of the lack of drainage in the permafrost (permanently frozen ground). A continuous stratum of permafrost

is found where the annual mean temperature is below ≈23 °F (−5 °C). Further south at inland stations drought conditions can prevail even though the absolute amount of precipitation is greater than in the north. Drought conditions probably prevail because summer temperatures are so high in the south that evaporation exceeds precipitation.

The characteristics of the snow cover, such as thickness and duration, have important climatic effects on the heat and moisture exchange at the surface. Steady snow melt caused by solar radiation usually begins by mid-June. While marked differences occur from year to year, the ice is usually snow free by the middle of July. Over the central Arctic Ocean the snow cover becomes reestablished in late August, with a thickness up to 15 in (38 cm) by late spring.

3.2.4 Wind

Three main factors affect wind speeds in the Arctic. These components are (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.

Since most cold, interior continents contain high pressure systems in winter, stations there will experience lighter winds than will the coastal stations. Examples of stations with strong winter winds are Jan Mayen, Vardo, Barter Island, and Dikson Island. Each of these islands is exposed to steep wintertime pressure gradients. In contrast, it is interesting to compare these stations to Angmagssalik and Anchorage, which are coastal but have small wintertime wind velocities due to sheltering effects, while the unsheltered Arctic coast of Alaska has frequent gales.

Generally, continental interiors have the lightest annual winds due to the moderating effect of high pressure systems. Subsidence in the area of anticyclones produces dry air promoting inversions that inhibit vertical mixing and keep the winds light in winter. Some exceptions do occur, however: Baker Lake, in the Northwest Territories, (NWT) is situated in the flow between the semi-permanent winter high over Yukon/Alaska and lower pressure over the eastern Arctic. Baker Lake's January mean temperature is −27 °F (−33 °C) and a mean wind speed of 13 kt (7 m/s). In winter, many areas bordering on continental interiors experience persistent strong winds due to cold air drainage from the interior that flows through mountain passes and river valleys. In summer, the absence of cyclones and weak pressure gradients keep the winds to a minimum. Examples are Verkhoyansk, Fairbanks, and Watson Lake.

The dearth of extreme wind speeds is due to the high frequency of inversions and the lack of topographic effects. Wind speeds average between 8 to 10 kt (4–5 m/s) and gales (>28 kt [14 m/s]) are infrequent, occurring only a few times a year with 40-kt (21-m/s) winds being an upper limit.

In central Greenland the wind is greatest in winter and least in summer. Here the inversion persists in every season, but since the ice mass is so high the wind speed is determined by the 700-mb contour gradient, which is steepest (strongest) in winter.

Other regional characteristics are present in the Arctic wind systems over land. For example, in eastern Siberia a monsoonal change in wind direction is experienced. During winter southerly winds prevail, being strongest over the coast and light or calm in the interior. At Verkhoyansk winds average less than 4 kt (2 m/s) for the months of September through April. The frequency of storms there is greatest during the months of June, July, and August,

with an average of two or three per month. During these months, the thermal low pressure system shows greatest development over the interior. Along the coast the frequency of storms is least during the summer. Tiksi Bay has an average of one storm per month from April through September. During the colder months, when the anticyclone is well developed over the continent, no gales occur in the interior at Verkhoyansk; but during these same months at Tiksi Bay, gale frequency is greatest. From October through May an average of four storms occur each month.

The wind regime of Alaska, like that of eastern Siberia, is to some extent monsoonal in character. In the warmest months a thermal low develops over the interior of Alaska, whereas, in winter, high pressures are prevalent. As a result, winter winds at Fairbanks are regulated by the Arctic high pressure system to the north and migratory lows from the southwest. Surface winds at Fairbanks itself are light because the town is located in a sheltering valley in which the average winter surface wind speed is 4 kt (2 m/s). In summer the air pressure increases over the Gulf of Alaska, and a local thermal low is established over the interior. The prevailing winds at Fairbanks are southwesterly at about 5 kt (2.5 m/s).

Climatology records show that a north and northwesterly air flow is directed over the Canadian Archipelago throughout most of the year. The flow is strongest during winter.

The Bering Strait and the land areas immediately to the east and west of it are low in elevation and present no obstacle to the free interchange of air between the Arctic Basin and the region of the Bering Sea. During the winter the prevailing wind through this region is northerly, and this flow of cold air replenishes the air moving from the Bering Sea into the Pacific Ocean. During this season the Aleutian Low is well established, and the surface winds and upper winds at Nome, Alaska confirm this. Sixty percent of the surface winds at Nome are from the north to northeast at about 9 kt (5 m/s) and continue from that quadrant up to heights of 6,000 ft (1,828 m).

High surface wind speeds at many Arctic stations are often enhanced by local topographic effects. Many stations show locally preferred wind directions due to topography, and surface wind data from Arctic land stations must not be taken as typical of the wind over larger areas. The following examples indicate the wide variety of conditions under which strong winds occur:

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 that rises to an elevation of 8,000 ft (2,440 m) 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 kt (51 m/s). The wind is warm and dry and obviously belongs in the foehn category.

Thule, Greenland. Thule 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 ft (762 m) 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 winds are due in part to turbulent mixing, in part to adiabatic descent from the ice arm, and in part to

the advection of air that is of more southerly origin. A strong pressure gradient between Thule and Upernavik, 300 mi (483 km) to the southeast, is a major factor in the occurrence of the "Thule Wind." Strong katabatic winds at Thule are triggered by northward-moving occlusions and/or their parent lows.

Alert, Ellesmere Island. Alert is located at the northeastern 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 ft (914 m) in elevation. The strong, relatively warm 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. A foehn effect is suggested, but the humidity is sometimes quite high both at the surface and aloft so that channeling and barrier effects are of importance.

Barter Island. This small, flat island located immediately off the north coast of Alaska is occasionally visited by strong and persistent winds that blow parallel to the coast, either in an easterly or westerly direction. The abnormal wind speeds are caused by a funneling of the air around the semicircular-shaped protrusion of the Brooks Range, located 50 mi (80 km) to the south of the station.

Juneau, Alaska. Juneau is located on the southeast coast of Alaska on a narrow shelf that rises abruptly to an ice plateau of roughly 5,000 ft (1,525 m) elevation. The strong winds that occur there are known as *takus,* since they are popularly associated with the nearby Taku Glacier. These winds are of the bora type, however, 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.

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 are met. 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 frequently 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 funneling effect.

3.3 Some Regional Coastal Climatology

Three main factors control climate, in the Arctic as elsewhere. These components are latitude, distribution of land and water, and pressure.

Latitude is important in the Arctic because it causes the marked variation in amount of solar radiation throughout the year. The long hours of sunlight in summer and the absence of sunlight in winter result in a wide range of temperatures, while the oblique angle at which the Sun's rays reach the Earth greatly reduces their effectiveness.

The effect of latitude on weather is modified by *land and water distribution* because landmasses heat and cool rapidly, whereas water heats more slowly and retains its heat longer. It is well known that continental climates are characterized by extreme temperatures and maritime climates by moderate temperatures. In the Arctic, however, the water areas are completely ice covered for more than half the year, and the insulating power of the snow and ice reduces the modifying influence of the water temperatures. A tendency exists, therefore, for the climate of the Arctic Ocean and Canadian Archipelago to be characteristically continental in winter and maritime in summer.

The third factor, *pressure,* is closely linked with the differential heating of land, water, and ice. Thus, in winter the weather of the Canadian Arctic is dominated by a high pressure system over the land and ice of the western Arctic and a trough of low pressure in the area of Davis Strait and Baffin Bay, where open water is commonly present. As a result of this pressure pattern, cold dry air covers the western and central parts much of the time, bringing clear skies and good weather, while in the eastern Arctic cyclonic activity is present with frequent storms moving northward into the Baffin Bay area. Consequently, it is not unusual for Thule, Greenland to have temperatures considerably higher than Edmonton and Saskatoon.

3.3.1 Brooks Range and North Slope of Alaska

The climate of Alaska depends primarily on three factors: (1) latitude; (2) proximity to oceans; and (3) influence of mountains. Since this discussion is confined to the North Slope and Brooks Range, consideration will be given to the impact of each of these factors on Alaska's Far North.

At the latitude of Barrow, which is 71 °N, during a 2-month period in winter (December–January) "day" is only a few hours of twilight (no sunlight). Contrast this with Anchorage at 61 °N where, for the same 2 months, sunlight duration is about 5.5 hours.

The Arctic Ocean to the north has year round sea ice. While the ice may be driven well offshore in the summer months, the water temperatures are usually not much above freezing. In winter, the blanket of sea ice over the water can chill to well below freezing. In this respect, it becomes almost an extension of the continent.

A number of mountain chains in the state act as barriers to the free flow of oceanic air (Fig. 3-7). Southward from the Arctic Ocean is the Brooks Range, the northernmost extremity of the Rockies, with a lateral extent of nearly 500 mi (800 km) and elevations over 9,000 ft (2,750 m). Next is the great Alaskan Interior, a jumble of unglaciated uplands and broad river valleys. South of that lies the majestic Alaska Range, dominated by Mt. McKinley, elevation 20,320 ft (6,194 m). This chain is followed by another complex

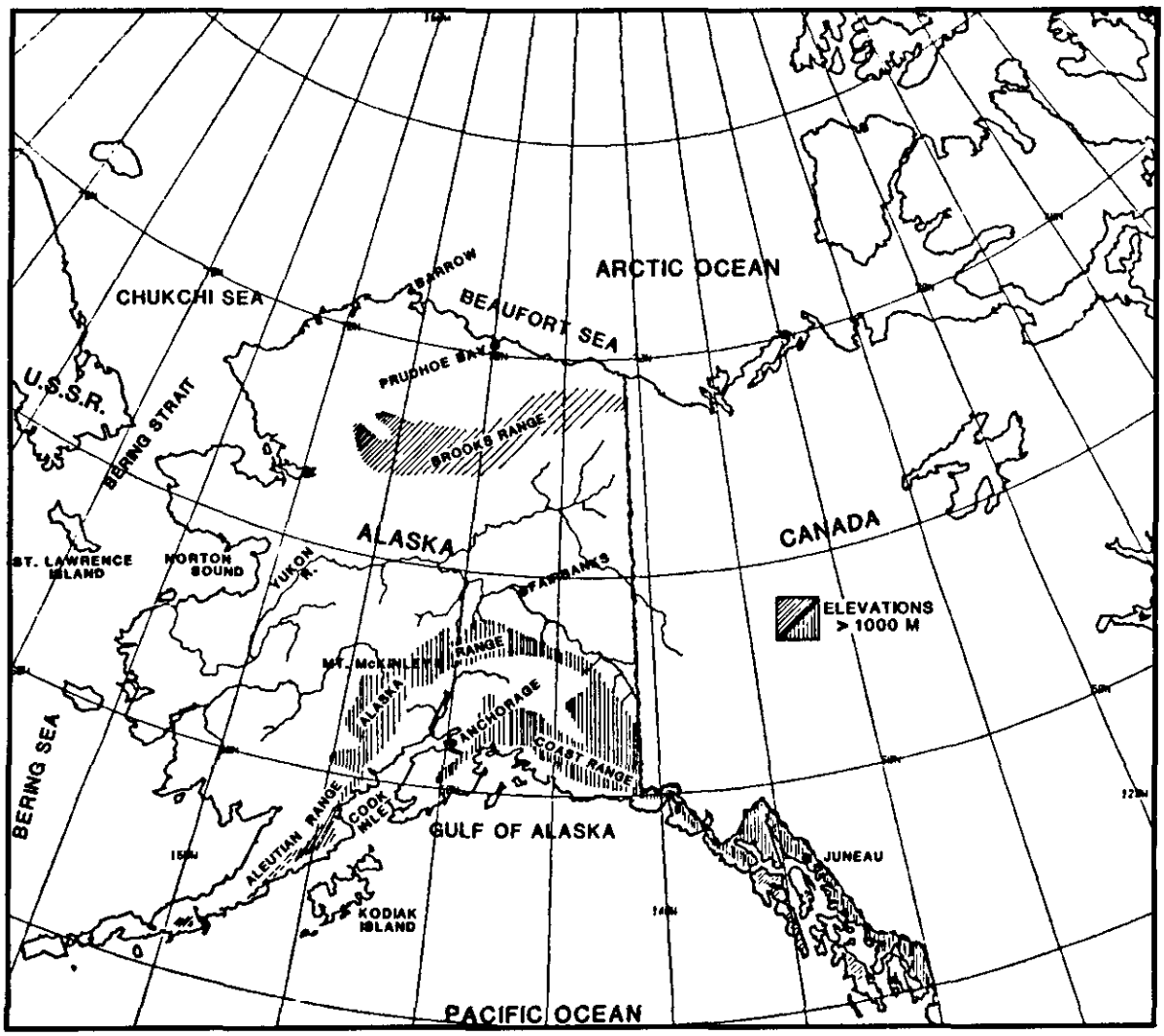

Figure 3-7. Land Features of Alaska.

of lowlands and mountains and finally the Coast Range complex, which rings the north and east sides of the Gulf of Alaska. The overall pattern is one that allows unmodified oceanic air to reach the Alaskan interior only from the west. The climate of the Arctic Slope is strongly affected by air masses from the Arctic Ocean, and that of the Coast Range by the Pacific Ocean. Consequently, the climate of the state is powerfully zoned by the mountains.

Clearly a truly Arctic climate exists on the North Slope of the Brooks Range. Precipitation is quite light, less than 8 in (20 cm) per year. Trees are nearly nonexistent. Mean summer temperatures along the coast are somewhat lower than in the western Aleutians; for example, Barrow, at the northernmost tip of Alaska, has a mean July temperature of 39 °F (4 °C). Cloud overcast is infrequent in spring but almost constant in summer. Midwinter mean temperatures are marginally lower than those of the interior, but the spring warming is far slower. February is usually colder than January. Absolute minimum temperatures are not as extreme as in the interior, but combinations of low temperatures

and high winds are frequent, resulting in extreme wind chill conditions. Along the entire North Slope, permafrost is continuous and extends to considerable depths (over 1,500 ft [457 m]). In the Brooks Range itself, as expected, colder temperatures at higher elevations and more snow and higher winds are likely except in strong inversions.

The following regional features influence the climate of Alaska's Far North:

Mountain Wave Clouds. The mountains of Alaska frequently create spectacular cloud formations. These overcasts range from local wave clouds to much more extensive cloud streets and vortex structures. An example of a wave cloud is shown in Fig. 3-8. Undulating waves occur when moist air is forced to rise over mountain ranges or mountain peaks. If humidity is high and the layer is relatively stable, the air waves become clearly visible by the clouds that form on the mountain crests. Photographs from polar orbiting satellites have captured some of these fascinating shapes and have revealed how surprisingly far they extend downwind.

The types of waves and vortices are determined by the speed of the wind and the size of the obstacle they encounter. They may be accompanied by turbulence and may be hazardous to aircraft.

Sea Breezes. Sea breezes are common mesoscale features in coastal areas around the world. The intensity, duration, and extent of the sea breeze circulation are mainly

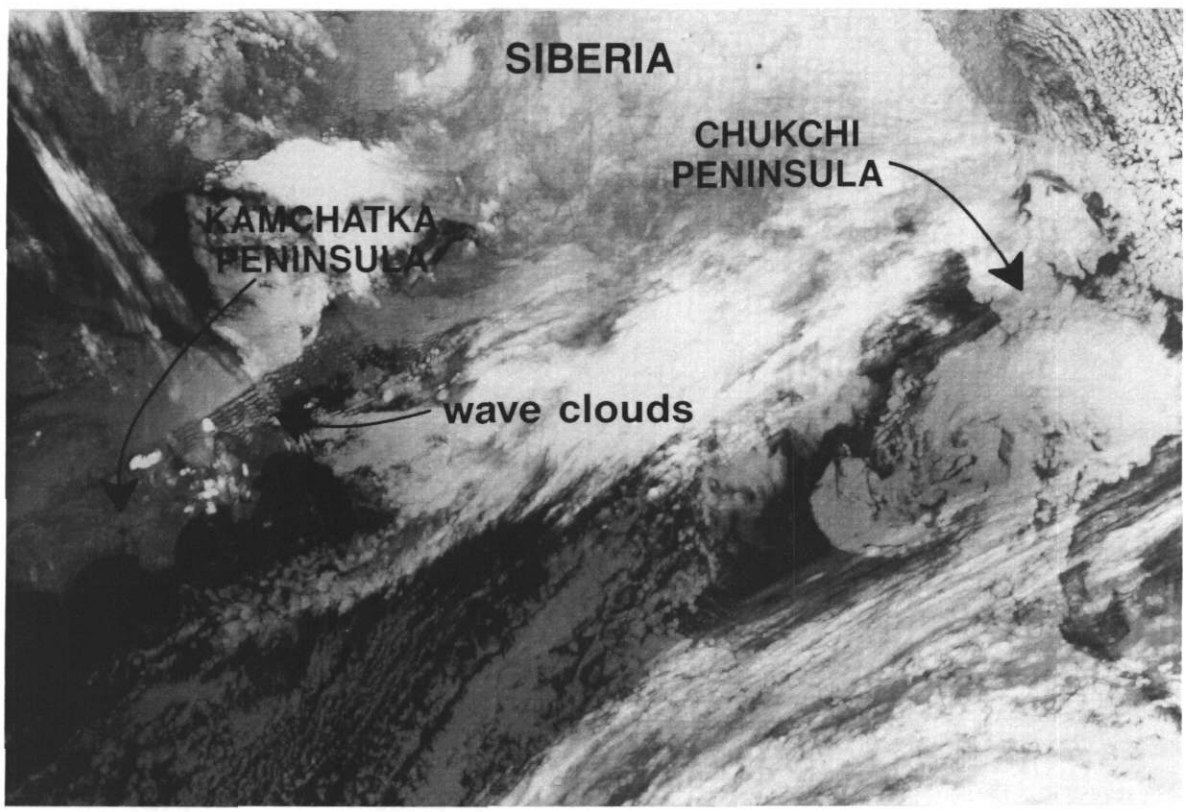

Figure 3-8. Mountain Wave Clouds over Kamchatka Peninsula (11 October 1986).

determined by horizontal gradients of temperature between land and water. The horizontal gradient in surface heating is caused by a difference in the responses of land and water to solar radiation. For equal amounts of incoming radiation, the surface temperature of the water will be less than the land because more energy is used for evaporation, and the rest of the energy reaches greater depths due to water turbulence and water transparency.

Coastal lands warming in the morning cause an increased thickness of isobaric layers over land and a downward sloping upper pressure surface from land to sea. The resultant horizontal pressure gradient force accelerates air from land to sea aloft. This air movement causes an increase in sea-level pressure offshore (mass excess) and a decrease onshore (mass defect). The result is a sea-level pressure gradient force that accelerates air from sea to land. This circulation is normally reversed at night when the land-sea temperature differential reverses, and the land becomes cooler than the ocean. In the Arctic such a circulation is not expected, however, because (1) the great stability of the air here inhibits vertical circulation of any kind, which would preclude a sea breeze, and (2) since the land-sea temperature does not reverse at night in the Arctic, no land breeze would be expected. From data collected in August of 1976, 1977, and 1979 (along the Beaufort Sea Coast of Alaska), Kozo (1982) has shown, however, both circumstantial and direct evidence of the existence of sea breezes as far north as 70 °N latitude. Evidently the sea breeze can occur even though the stability is great. The counteracting effect of a low-level temperature difference between land and sea can be as much as two times as great as the contrast in temperatures in low and middle latitudes. The land breeze, however, does not appear because the temperature difference between land and sea does not reverse at night along the Beaufort Sea coast.

The foregoing theory is borne out by the following observations:

(1) The Arctic sea breeze is largely responsible for the increased persistence of surface onshore (northeasterly and easterly) winds as confirmed by historical data for August along the Beaufort coast.

(2) Normal planetary boundary layer turning, usually the result of a three-way balance between the Coriolis force, the large-scale pressure gradient force, and the viscous force should show winds shifted about 20° counterclockwise from their free stream direction to their 10-m direction. Along the Beaufort coast a more extreme turning occurs. The apparent "turning" is more likely to be in excess of 120 degrees from the free-stream level to the surface.

(3) Since the Coriolis parameter for 71 °N is 37 percent greater than that for midlatitudes and because the summer sea breeze along the Alaskan Arctic coast is not followed by a land breeze, temporal clockwise rotation (northerly to easterly) of the sea breeze stops at the coast, hence the increase in frequency of easterly flow in August.

The Arctic sea breeze initially faces a strong ground-based inversion that implies extreme stability and small eddy thermal diffusivity. This factor limits the vertical extent of the circulation since the depth of the landward flow is often seen to be less than 1,300 ft (400 m). The initial conditions faced by the Arctic sea breeze are similar to those encountered by midlatitude land breezes.

This stability typically occurs under calm synoptic conditions. Whenever actual synoptic offshore winds exist, however, they retard the shoreward movement of the breeze and cause a cold front type system, which has been referred to as a sea breeze of the second kind. Note that early morning offshore flow advects warm air from land out over the water to start the sea breeze offshore rather than near the coast. It has also been noted that when the large-scale wind is from land to sea, the sea breeze develops at sea and advances more slowly landward, reaching the coast much later in the afternoon, acting like a front with a characteristic wind shift.

The database for Arctic sea breeze studies has been enhanced by the discovery and development of the Prudhoe Bay oil fields in the mid-1960s. Since the nearshore ocean current flow along the Alaskan Beaufort Sea coast is primarily wind driven, a direct application to oil spill trajectories in lagoons and embayments exists. This possibility has given rise to still other studies of wind and ocean circulations in this critical area of the Arctic. A discussion of these studies follows, and Fig. 3-9 shows the locations of most of the oil rigs in the Beaufort Sea.

Corner Effect of the Brooks Range. In a study of topographic effects on the wind in Alaska, Dickey (1961) concluded that strong ageostrophic winds occurred at Barter Island due to the "knob" of the Brooks Range. Dickey observed that during periods of high surface winds, Barter Island has a wintertime abundance of west to southwest winds, while Barrow, less than 300 mi (483 km) to the northwest, has an east to northeast bias in its wintertime weather observations.

Schwerdtfeger (1975) took issue with the knob effect and showed rather elegantly that when a stable air mass is moving toward a mountain barrier without being heated from

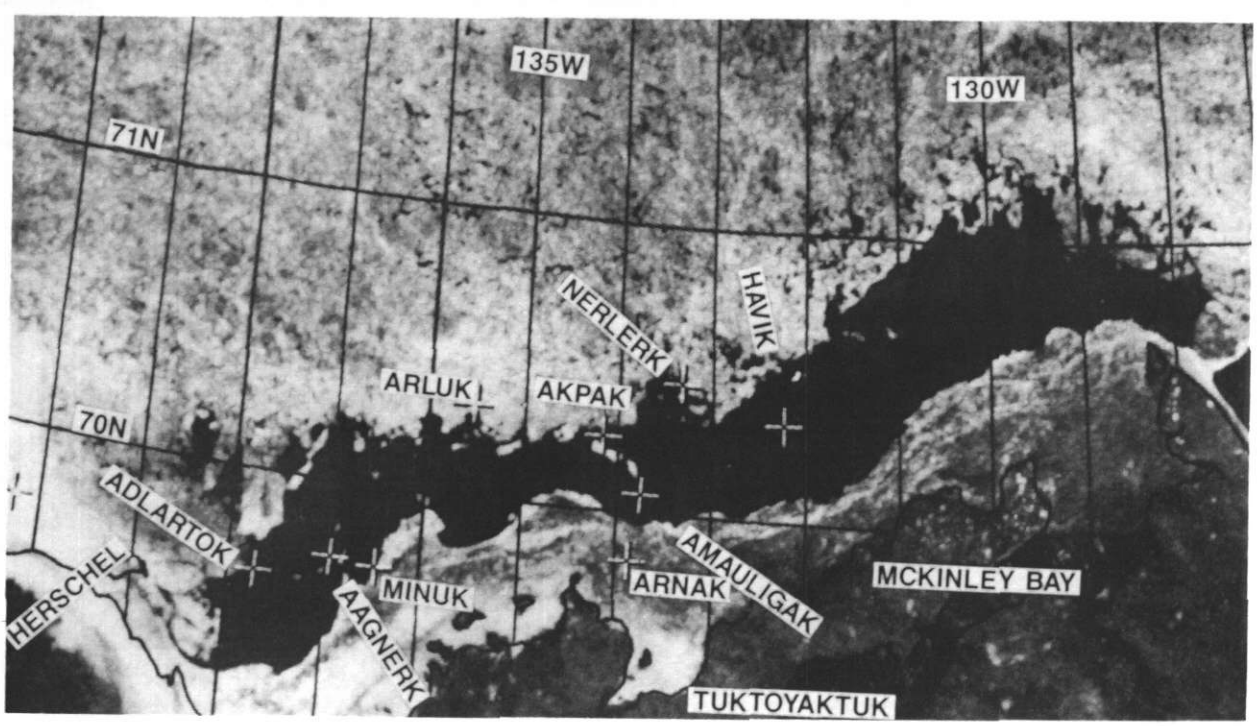

Figure 3-9. Oil Rig Locations in the Beaufort Sea (Atmospheric Environment Services, 1985).

below, the increased elevation will alter the isothermal surfaces, thus changing the horizontal surface pressure gradient. This pressure change results in a horizontal temperature gradient directed toward the barrier and hence, a thermal wind parallel to the barrier. The flow then becomes deflected toward the east, and its strength increases markedly along the north slope. Schwerdtfeger also found that the remarkable difference between the resultant wind vectors of Barrow and Barter Island is not restricted to the surface layer; it is found up to about 6,500 ft (2,000 m).

Kozo (1980) has verified Schwerdtfeger's findings with observations from the buoys deployed by the Polar Science Center of the University of Washington. His results indicate that the average turning of the wind from above the boundary layer to the surface was 140 degrees during periods of what he calls "the mountain barrier baroclinity." The horizontal extent of this phenomenon ($\approx$75 mi [120 km]) indicates that surface wind measurements in winter months for a coastal zone from east of Barter Island to Prudhoe Bay may be poor indicators of the stress exerted on nearshore ice. The implications for wind and ice movement forecasting is clear.

Strong Valley Winds. Contrary to popular thought, most authorities on Alaska's weather discount the presence of katabatic winds in the mountain valleys. Rather, in Alaska the increases in temperature of the valley surface during wind episodes always occur during the cold season. The surface temperature inversions that occur under clear sky conditions are broken by these winds as the warmer air aloft is brought to the surface by mixing, and the valley surface temperature rises nearly to that of the surrounding mountain top temperatures.

To locate the areas of strong valley winds with accompanying blowing snow and severe turbulence, use can be made of the infrared sensor on the NOAA polar-orbiting satellites. Marvill and Jayaweera (1975) discussed this possibility while reporting on two valley wind episodes in early 1975. The ability of the satellite to view these areas of warming and associated turbulence will alert forecasters to valley wind episodes for which no conventional surface observations or pilot reports are available. When strong valley winds occur during the winter months, the infrared (IR) imagery from the Defense Meteorological Satellite Program (DMSP) satellites shows a dark band (relatively warm surface) as illustrated in Fig. 3-10.

Unlike clouds, this band on the IR imagery appears with a very smooth texture and is coincidental with the valley. Since most valleys have very small slopes, drainage effects are not usually responsible for the high winds. Consequently, synoptic-scale surface pressure gradient seems to be the most important parameter to initiate and maintain strong valley winds. Marvill and Jayaweera observed the presence of a large amplitude, 500-mb ridge to the west of the area of interest at the beginning of both windy periods. This ridge location places the interior valleys under the region of the flow pattern that typically produces subsidence, clear skies, and strong radiational cooling at the surface. The surface high center was in the southern half of the Yukon Territory, creating a surface pressure gradient nearly parallel to the valleys. This situation appeared to be ideal for the production of strong valley winds and for good satellite detection of the area affected by these winds. Once the upper air ridge had moved east of Alaska, satellite viewing was hampered by cloudiness.

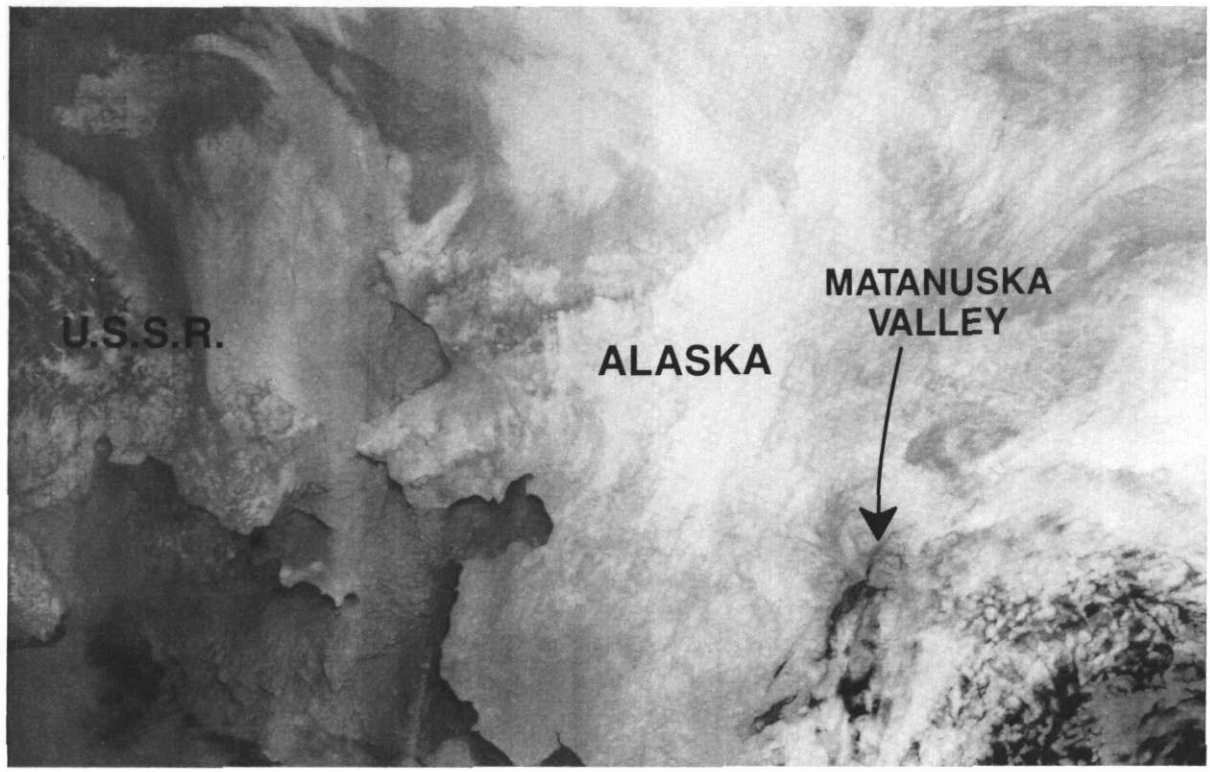

Figure 3-10. IR Imagery Showing Dark Band in Warm Valley.

3.3.2 Canadian Archipelago

In summer a flat pressure gradient is characteristic of the Arctic Ocean and the Canadian Arctic. The maritime influence asserts itself, so that differences between temperatures at 75 °N and the pole are negligible. A frontal zone is generally found along the north coast of the continent, although its position varies from day to day. Disturbances developing along this front are a feature of the weather over the Arctic during this season. The following review of the characteristics of Arctic weather patterns shows that a difference of about 2 months occurs in the length of the summer between the far north and the mainland, and a corresponding difference in the other seasons.

Airmasses. The main properties that distinguish the continental airmass from the maritime airmass in winter are the large temperature inversion at the surface, stability, low temperatures, and dryness. In summer the continental airmass is drier, warmer, and less stable than the maritime airmass.

Pressure Patterns Unique to the Area. The weather of Arctic North America is dominated by the Arctic High, the Icelandic Low with its westward extension, the Baffin Bay Trough, and the Aleutian Low. (Refer to monthly mean sea-level pressure maps, Appendix B).

Arctic High. In the fall pressures build up over eastern Siberia and the Arctic Ocean, and the two systems are often shown on weather maps as a single ridge. This condition persists until spring, when the Siberian High dissipates and the center of the Arctic High moves to a position northwest of the Canadian Archipelago, with a ridge extending south toward Churchill. It reaches peak intensity in April and May. In summer the gradient weakens to such an extent that the Arctic High practically loses its identity and is frequently displaced by migratory lows.

Yukon High. The Yukon High can be described as a ridge extending from the Arctic High over the Mackenzie Valley. It commences to build up in November, reaches its peak in February, and disappears in the spring.

Baffin Bay Trough. The Baffin Bay Trough is the name generally applied to the westward extension of the Icelandic Low. The Baffin Bay area is one of almost continuous cyclonic activity, and migratory lows enter it in all seasons. Many come from the south, others from the Gulf of Alaska across Canada, while some come from the Bering Sea via the Beaufort Sea and the Canadian Archipelago. Occasionally a storm arrives by way of the Arctic Ocean. Most have passed their peak intensity by the time they reach the area.

This trough, like the Icelandic Low itself, is most intense in winter, and it frequently extends westward over the islands from the head of Baffin Bay. In spring the trough begins to fill and during the summer is usually centered off northern Labrador. During this season it frequently loses identity because of the weak pressure gradient. Occasionally the Baffin Bay Trough becomes part of a weak trough, associated with the Arctic Front, that lies in an east-west direction along the north mainland coast.

The extent and permanent nature of the Baffin Bay Trough has a strong influence on the weather of the eastern Canadian Arctic, and the circulation over the archipelago is influenced more by the Icelandic-Baffin Bay Low than by systems originating in the Pacific.

Thule Low. During late summer and fall a "closed" low appears frequently at the northern end of Baffin Bay, which is generally called the Thule Low. Usually little weather is associated with it, and the cloud cover is thin and diffused.

While the preceding paragraphs described the prevalent weather patterns of the Canadian Archipelago, the following paragraphs will address the various wind, cloud, precipitation, and temperature regimes of the Canadian Archipelago.

Winds. In winter the general pressure distribution over the Canadian Archipelago consists of a ridge extending from the Arctic Ocean southeast to the central provinces, and a trough over Baffin Bay and Davis Strait. The prevailing winds near the surface are therefore generally north-northeast over the Canadian sector of the Arctic Ocean; northeasterly over Ellesmere Island, backing to the north and northwest over the central and western part of the archipelago; and generally northwesterly over the mainland. Storms moving into Davis Strait may alter the pattern in the eastern part of the archipelago and similarly in the western part when storms arrive from Alaska. Aloft the pattern is much the same, with northwest winds usually dominating except over northern Ellesmere and the North Pole area, where they may be easterly.

In summer, owing to the weak pressure gradient, winds are variable and subject to frequent changes as migratory lows drift across the region. At this time of year the frequency of winds increase from the east and south. In general terms, the mean circulation aloft is similar to the winter pattern.

Wind speeds over the Canadian Arctic average from a low of about 4 kt (2 m/s) at Arctic Bay, which is sheltered, to a high of 16 kt (8 m/s) at the Resolution Island site (eastern Hudson Strait), which is located on the top of a small rocky island at an elevation of 1,207 ft (368 m) and experiences many storms.

Clouds. Throughout the Canadian Arctic and the Arctic Ocean the maximum cloudiness is in autumn and the minimum in winter. This effect is in direct contrast to more temperate latitudes, where winter is usually the cloudiest season. Cloudiness increases in the spring and reaches a secondary maximum in late May or June. Land areas, away from coastal regions, may have a secondary minimum in July. In August cloud coverage increases in all areas, building to the fall maximum. Over the Arctic Ocean and some coastal areas the spring maximum is delayed and the fall maximum advanced so that they merge into one.

Precipitation. Throughout most of the Canadian Arctic and the polar basin precipitation is light, and some areas of the Archipelago are among the driest in the world. A large proportion of the precipitation falls in summer; in the Arctic 90 percent occurs between May and October. Locator maps in Chapter 2 can be used as references for the following discussion. At Eureka 70 percent of the precipitation falls in summer; in north and east Baffin Island (NWT) 75 percent; in Victoria Island (NWT) and the Beaufort Sea region 65 percent; on the mainland 70 percent; while in southeast Baffin Island and Hudson Strait summer, autumn, and early winter are the seasons of greatest precipitation. South of Hudson Strait, however, considerable precipitation occurs throughout the winter.

The annual precipitation over the Arctic Ocean and the low lying western island of the archipelago averages about 6 in (15 cm). In Axel Heiberg and Ellesmere Islands (off northwest Greenland) the average is less than 5 in (13 cm), and only 1.74 in (4.4 cm) per year at Eureka, which is comparable to the desert areas of the world. Generally speaking precipitation increases southward. For the area between 75 °N and the Arctic Circle, the annual range is about 5 to 10 in (13–25 cm). North Baffin Island has about 6 to 7 in (16–18 cm). On the east coast of Baffin Island, the Clyde River area has about 7 in (18 cm), and the Home Bay region about 9 in (23 cm). In southeastern Baffin Island precipitation is higher, from 10 to about 18 in (25–46 cm), and in Hudson Strait the yearly average is from 12 to 18 in (30–46 cm), the highest in the Canadian Arctic. Cyclonic activity and open water in Davis Strait are largely responsible for this high figure.

The Arctic Ocean and the low-lying western islands of the archipelago receive about 25 in (64 cm) of snow per year. In the high elevations of the interiors of Ellesmere and Axel Heiberg Islands the snowfall is very light, averaging 10 to 15 in (25–38 cm) annually at Eureka, the least in the Canadian Arctic. In coastal areas the average is much higher; about 70 in (178 cm) at Alert and 65 in (165 cm) at Craig Harbor. In the Baffin Bay area the snowfall is about 60 in (152 cm), but at Pond Inlet and Arctic Bay in northern Baffin Island it is only about half as much. In southeastern Baffin Island the snowfall is much

heavier, the inland areas averaging approximately 40 in (102 cm) and the coastal regions 100 in (254 cm), the heaviest snowfall in the archipelago. In the coastal areas of the Beaufort Sea and Amundsen Gulf the total annual snowfall is between 50 and 60 in (127–152 cm).

In the Arctic Ocean and the archipelago north of latitude 75 °N, the greatest average monthly snowfall is usually in September. Between 75 °N and the Arctic Circle the maximum generally occurs in October; in Davis Strait and southern Baffin Island in November and early December. Generally snow falls in small particles, particularly in winter, so that the number of days on which snowfall is reported is out of proportion to the total accumulation.

Temperature. Three distinct types of climatic conditions are found in the Canadian Arctic: continental, maritime, and coastal, each having their own characteristic temperature curve.

The continental climate, which is typical of the mainland west of Hudson Bay and possibly Victoria Island, has a high summer temperature with a distinct maximum, and a very low winter temperature with a pronounced minimum. The annual range of mean temperatures is about 75 to 95 Fahrenheit degrees (42–53 Celsius degrees). The change from summer to winter, and vice versa, is rapid.

The climate of the maritime areas produces flat temperature curves for both summer and winter, and the mean annual range seldom exceeds 35F° (20C°). The most striking example of an Arctic maritime climate is found at Resolution Island. In the Arctic Ocean the climate is maritime in summer, but because of the almost solid ice pack it is more continental in winter; the mean range of about 60F° (34C°) is found around the coasts of many of the Arctic islands, particularly along the south and east coasts of Baffin Island, in the Beaufort Sea region, and around Labrador.

3.3.3 Greenland Icecap

Scorer (1988), in his presidential address delivered to the Royal Meteorological Society, credited Greenland with a major role in affecting the weather of the Northern Hemisphere. He contrasted the roles of the Tibetan Plateau and the Greenland Icecap. The former serves as an atmospheric heat source while the latter is just the opposite, a cold source. The implications of this fact are rather far reaching. Fig. 3-11 presents Greenland's topography measurements.

To begin with, Scorer pointed out a general slope downward to the north and west from a high point near east-central Greenland. On the plateau a permanent katabatic flow occurs toward the west coast, which can extend upward to heights of 1,640 ft (500 m) or more. It follows that the northwest coast of Greenland will be subject to significant katabatic winds. Also, in southeast Greenland at the fjord of Kangerlugssuaq, which has a very steeply sloping hinterland, the adiabatic warming is usually perceptible on IR satellite imagery (Fig. 3-12).

Of course, the even steeper slope (noted in Fig. 3-11) on the east-central coast can lead to significant katabatic flow in the Kangerlugssuaq Fjord (as seen in Fig. 3-12). Because of the strength of the katabatic flow, weather-producing land breezes can be expected to be associated with these katabatic outflows. The land breeze on the west coast is frequently made visible by a line of cloud lying up to 54 n mi (100 km) out to sea. Scorer also pointed

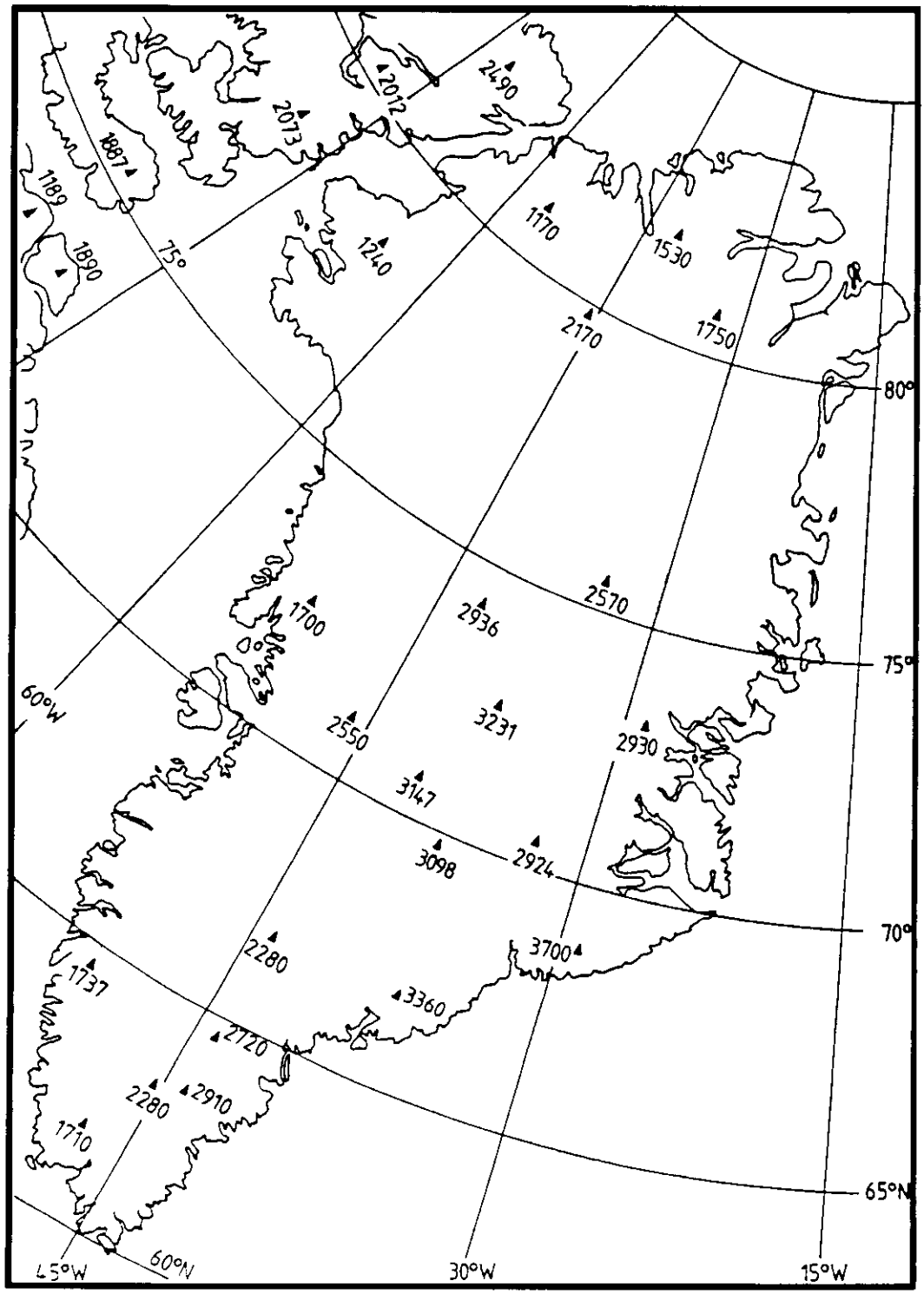

Figure 3-11. Greenland Topography with Measured Altitudes (Scorer, 1988).

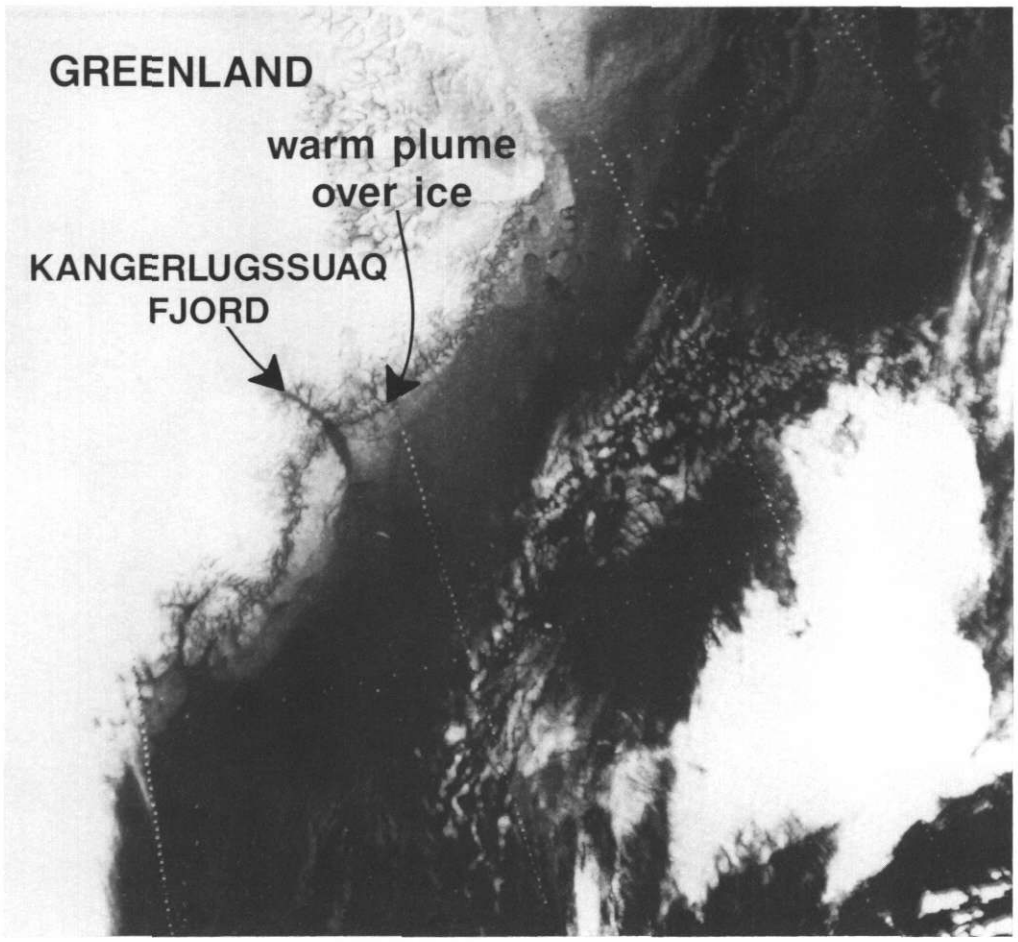

Figure 3-12. The Katabatic Warming of a Greenland Fjord.

out that being a cold source in a cold region it does not fill the upper troposphere with cloud, and the area is, therefore, relatively sunny for the latitude. With an albedo close to 1.0, the land breeze remains a cold source even at midday in June.

The main meteorological feature in the Greenland area is the so-called Icelandic Low. Scorer considered this a misnomer because it is in reality a low formed in the wake of the huge Greenland Icecap, which is a complete obstruction to the advance of air at sea level. For this reason isobars on surface analyses in the Greenland area should not be shown over the plateau. More representative would be the 700-mb height contours.

In any case, as the Icelandic Low moves from Cape Farewell toward Svalbard it causes a prevailing wind from the north or northeast along the lee coast that drives ice as far as Cape Farewell all year round. The Earth's rotation causes the ocean current to turn the cape and occasionally carry ice 50 to 100 n mi (100–200 km) up the west coast. During the fall and winter months many systems moving across the Labrador Sea and the northwest Atlantic split at the southern tip of Greenland, with a portion moving rapidly up the west coast of Greenland and the major portion continuing northeastward. Baffin Bay is often referred to as the graveyard of lows, and many of these lows move through Davis Strait.

Another feature often observed in the Greenland area is the marked anticyclonic curvature of lines of cirrus. This cirrus is seen in the air spreading out at a high level from a cyclone over Greenland, which is not accompanied by parallel surface air flow. Cirrus also originates in the high-level wave that exists above the air descending to the sea from the plateau. Because the surface air shows no convergence into Greenland it indicates that snowfall accompanying the cirrus moving inland is slight or nonexistent.

One new feature discovered by satellite imagery is the formation of rows of small, short-lived cyclones. They occur when the cold air behind a front has spread eastwards so that the front is aligned roughly east-west with the cold air on the south side. The warm air, originally on the south side and coming from the southwest, now lies on the north side of the front. In this manner an "inverted front" is regenerated by the easterly motion of the cold air, and new small cyclones come to life on it. This feature occurs in other Arctic locations such as the Bering Sea (Fig. 3-13).

Storm systems are often "steered" around Greenland and only occasionally penetrate over the plateau and then neither far nor for long. While Svalbard, Iceland, and other landforms near Greenland are some of the cloudiest places on earth, Greenland is relatively sunny, and yet it remains a very significant source of cold.

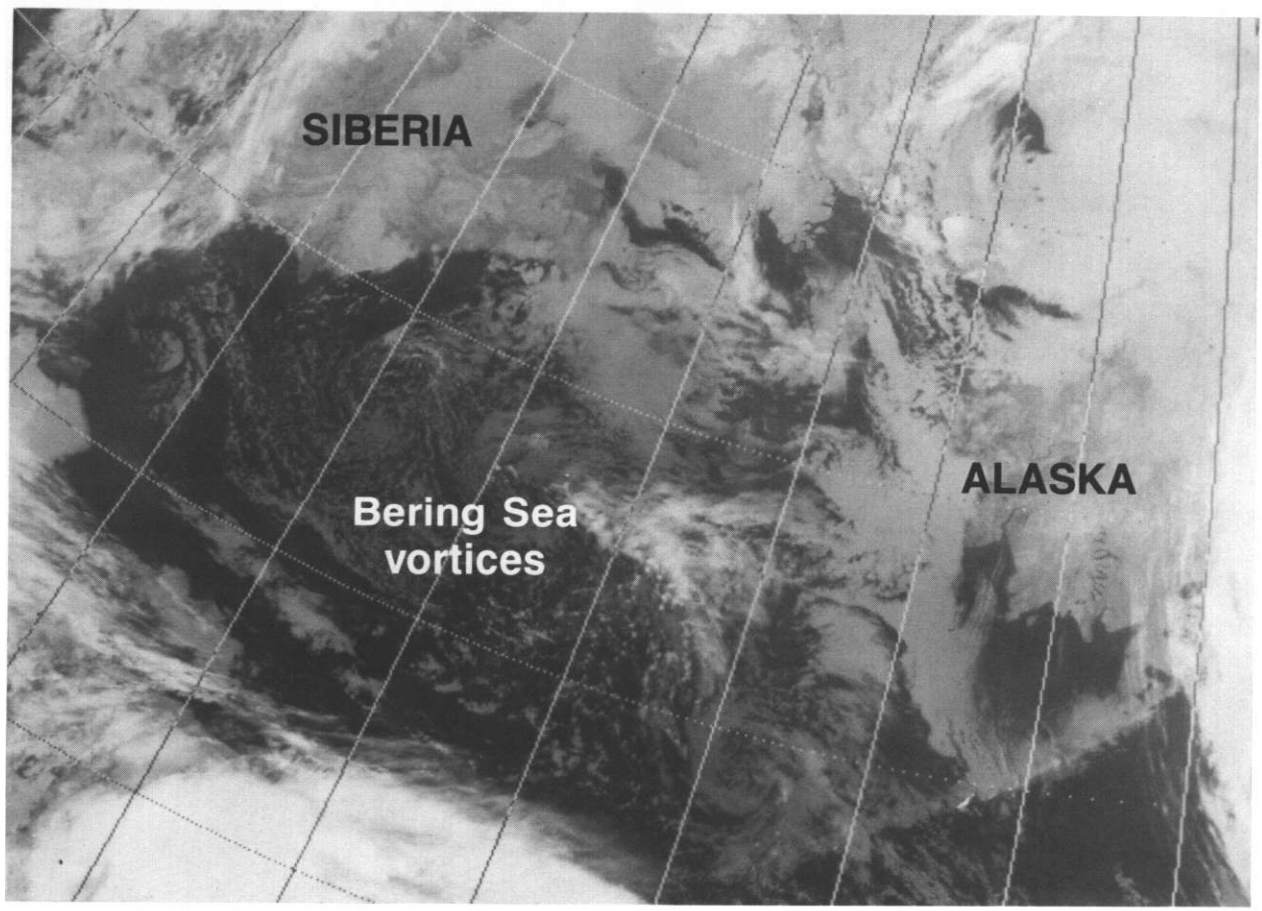

Figure 3-13. Row of Small Lows in the Bering Sea.

During a cruise on an icebreaker in 1979 Wadhams and Squire (1980) observed an ice-water vortex off the Greenland coast in the Fram Strait. The LANDSAT image they acquired appeared on the cover of EOS. Since then, many satellite observations have been made of similar vortices; for example, Figs. 3-14(a) and 3-14(b) are visible and IR images showing ice vortices near the Greenland coast. The eddy-like features are to the west and southwest of Iceland in the Denmark Strait. A comparison of the outline of the ice edges in this region shows that the edges on the IR image do not correspond to the edges in the visible image. In fact, at all points, the cold, bright boundary on the IR image extends seaward (closer to Iceland) than the visible boundary. This observation, which implies that cold, fresh, meltwater exists south of the ice boundary, is apparent in the visible imagery. The boundary of the water is uniquely revealed in the IR data (Fig. 3-14(b).

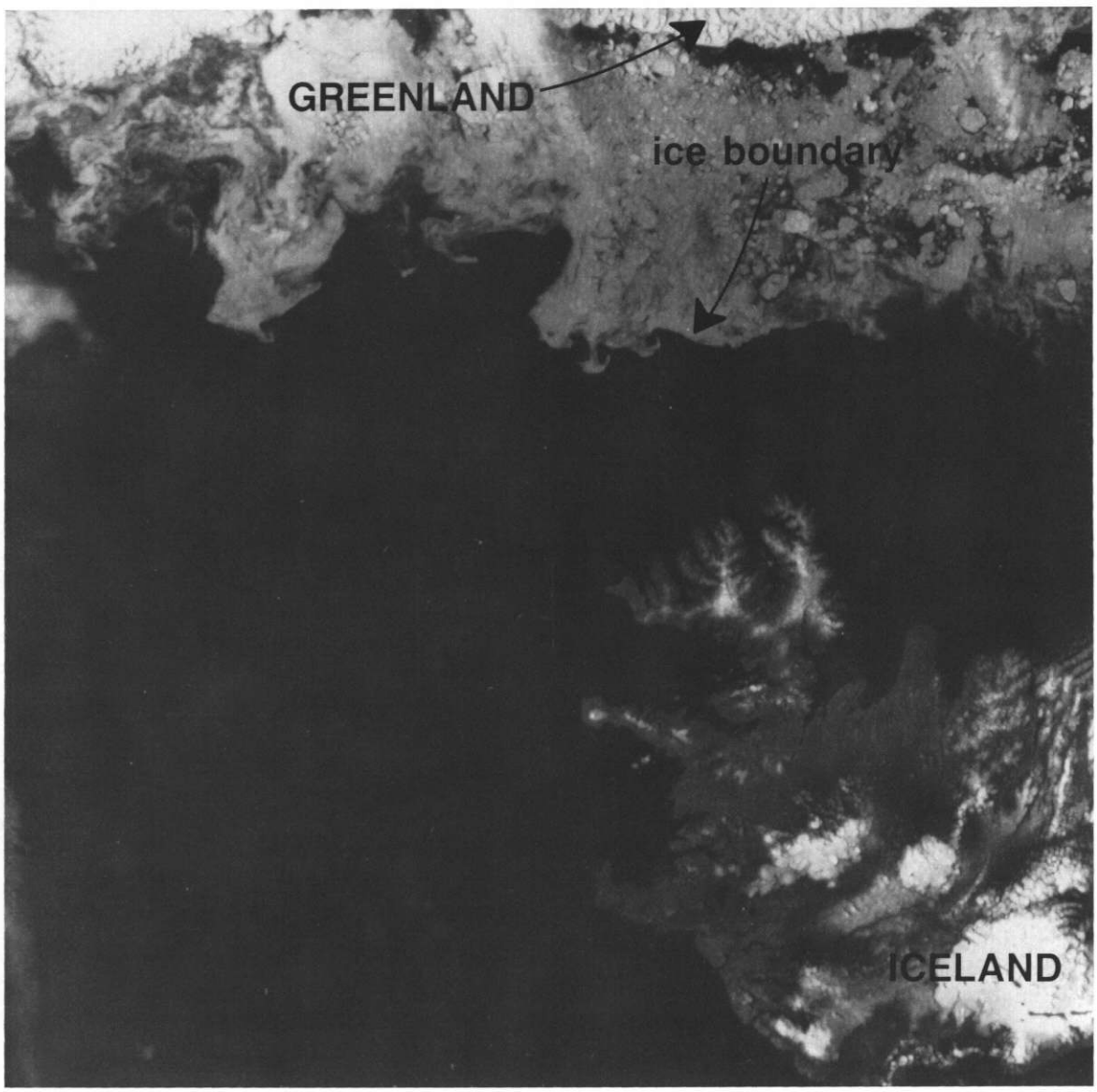

Figure 3-14(a). Ice-Water Vortex in the Denmark Strait (visible).

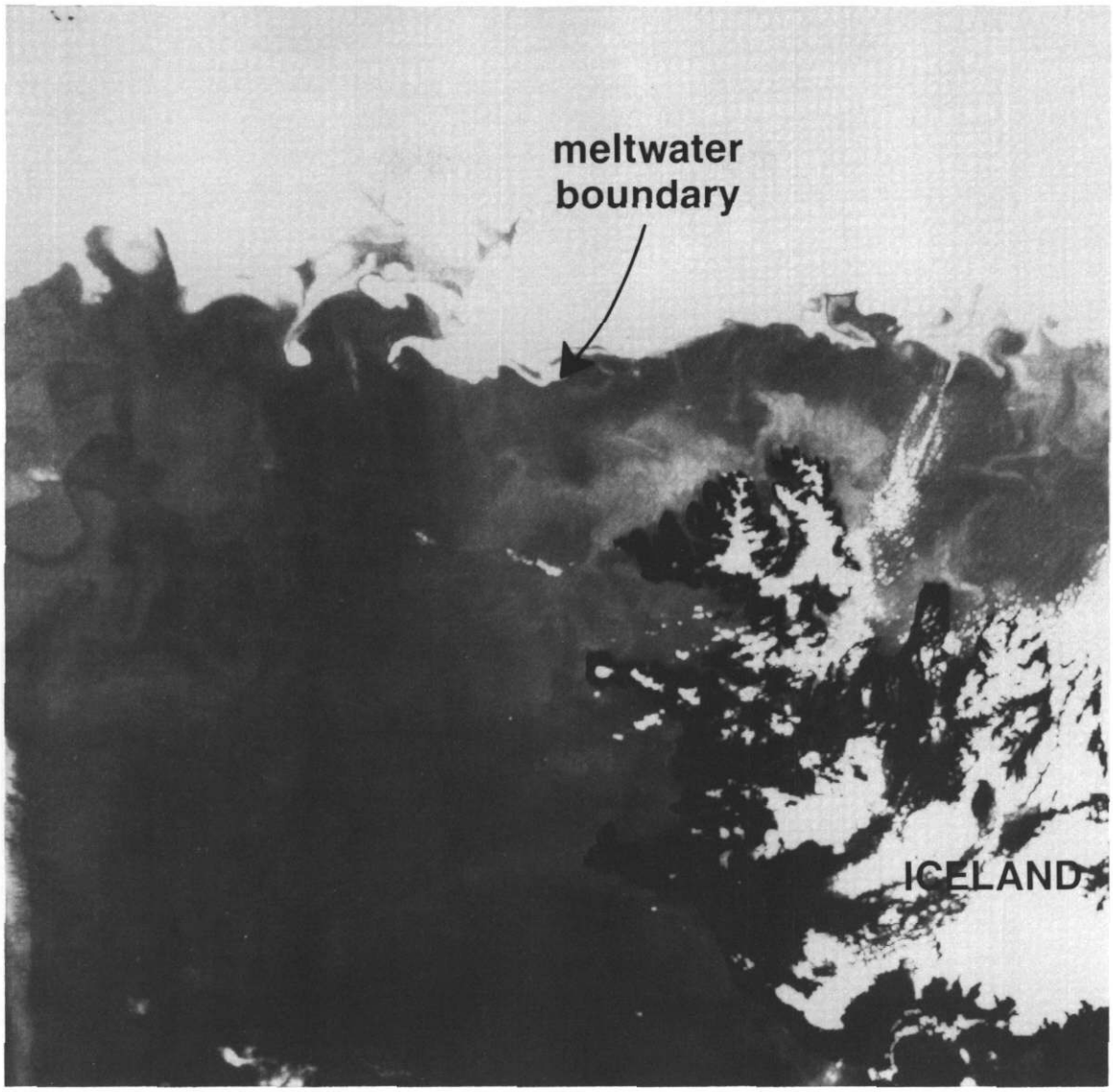

Figure 3-14(b). Ice-Water Vortex in the Denmark Strait (IR).

3.3.4 European U.S.S.R. and Siberia

The weather in this large region (Fig. 3-15) is unified by the movement of cyclonic storms throughout the area, which impart rapid and intense fluctuations in weather that stand in sharp contrast to the more stable conditions to the southeast. In fact, the northern half of the West Siberian Lowland is known to be one of the two areas on Earth (the other being the North American Great Plains) that record the greatest day-to-day variations in weather phenomena, particularly temperature. There are great meridional movements of air masses from Kazakhstan or the southern plain of the European U.S.S.R. As fronts sweep through the area these air masses interchange very rapidly and rival the frequency and magnitude of changes that take place in interior North America, east of the Rocky Mountains.

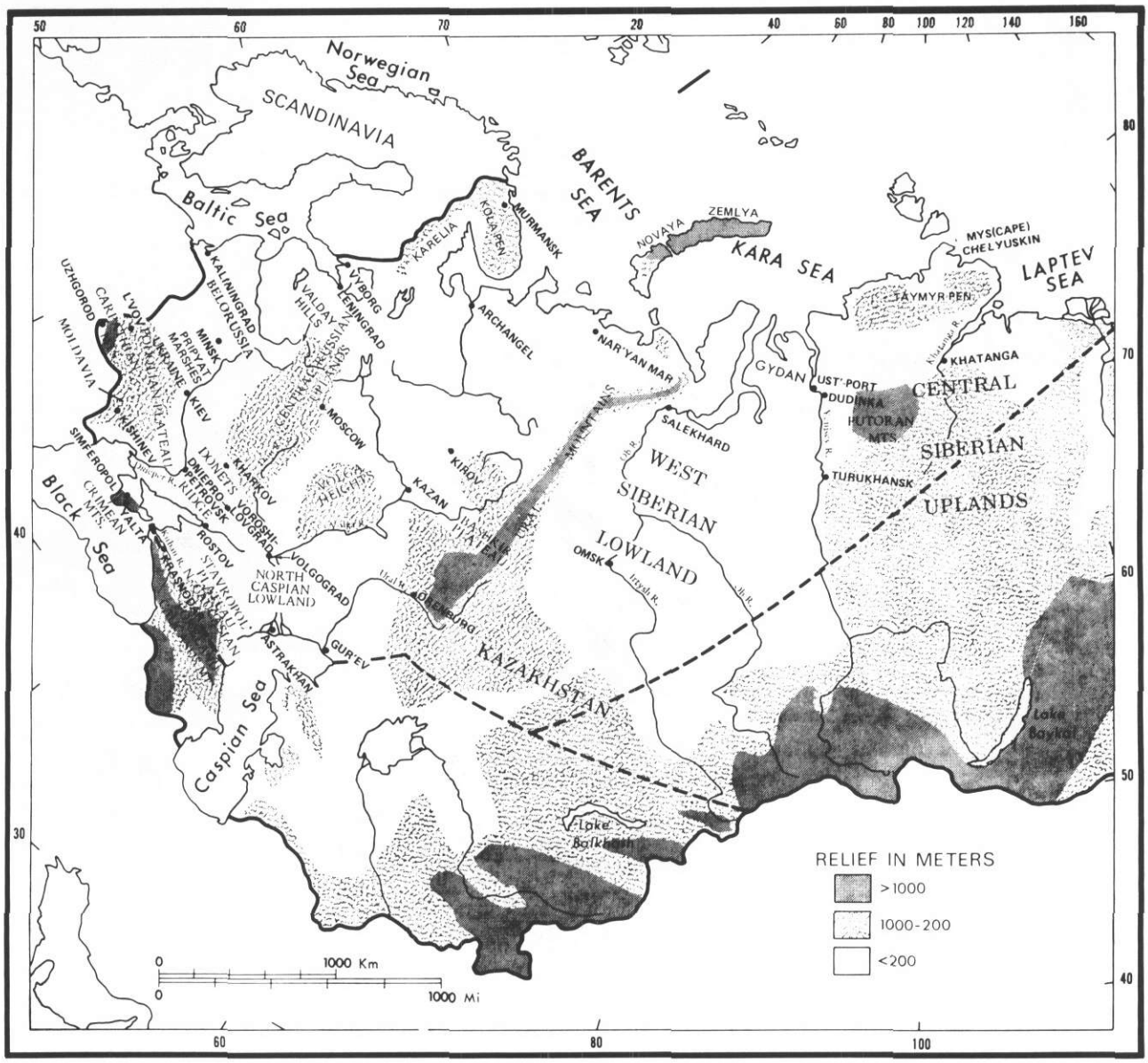

Figure 3-15. European U.S.S.R. and Siberia (Lydolph, 1977).

The entire Arctic coast experiences a monsoonal reversal of winds by season because of the great temperature contrasts between the land and sea. During midsummer, under constant sunlight, surface temperatures in the open water along the coast are held consistently near 32 °F (0 °C) by melting ice floes. The land, with higher thermal conductivity than the water, warms more strongly. Winds blow inland from the sea with average speeds during summer of 12 to 15 kt (6–7 m/s), which is significantly greater than even in the steppe regions to the south. These winds carry the cool humid influence of the sea onto the land and produce much cloudiness and fog. North of the Arctic Circle the region receives about 240 hours of sunshine (total hours the Sun is up minus cloudy hours). As cloudiness decreases somewhat inland, the hours of sunshine increase to about 300 in the southern

part of the tundra, which is the greatest number of sunshine hours during the summer of any part of western Siberia. Fog occurs about 80 days per year along the coast, and more than half of it occurs during the 3 summer months.

Added to the severity of the winter is the dreariness of the skies, which are cloud covered much of the time. During January much of the tundra zone averages at least eight-tenths cloud cover day in and day out. In the estuaries of the Ob and Yenisey more than 210 days per year experience overcast skies.

Precipitation is frequent but light. Some precipitation falls during 160 to 170 days of the year, and the annual total amounts to about 10 in (25 cm) along the coast and 14 in (35 cm) further inland. The precipitation maximum usually comes in late summer when frequent instability showers occur in the cool Arctic air that is being warmed as it moves southward over the land surface. Few thunderstorms occur.

Although the Ural Mountains have not been accorded the status of a separate region, they obviously cause the climate to change locally with elevation and induce enough perturbations in the west-east air flow across the Eurasian plain to produce significant differences in weather on both sides of the range. Although they are relatively low and broken, in many places they rise to 4,900 to 6,550 ft (1,500–2,000 m); such elevations have very significant effects on radiation, temperature, precipitation, and atmospheric circulation.

During the winter the topography of the Urals is significant in stagnating cold air, so that great temperature differences occur between enclosed lowland basins and upper slopes. Strong temperature inversions are the rule. In summer, usually a rather steep lapse rate develops along the mountain slopes, averaging approximately 4.4 °F/1,000ft (8 °C/km). One significant difference between the Urals and the surrounding plains is the amount of cloudiness. At the observation station of Taganay in the south-central Urals at an elevation of 3,616 ft (1,102 m), overcast skies occur more than 60 percent of the time every month of the year. During the average year fog is recorded for 230 days at Taganay. This high fog frequency must relate to stratus clouds that have been formed by orographic uplift and are lying on the ground at this elevation. Snow storms occur an average 97 days per year at Taganay, and the maximum depth of snow cover averages about 35 in (90 cm).

Apparently, a rain shadow effect exists on the eastern side of the Urals; these precipitation amounts may be due more to favorite routes of cyclonic storms than to the mountains themselves. Precipitation on the western side of the mountains is often about 12 in (30 cm) per year less than on the windward eastern side. Eastern Siberia (Fig. 3-16) is characterized by the highest degree of continentality on Earth. The Asiatic High brings consistently cold, clear, calm weather to the region. The area is predominantly mountainous, with the most rugged topography along the southeast. The nature of the topography has great influence on all local climatic elements.

3.4 Storm Tracks and Cyclogenetic Regions

Section 3.1.2 described the nature of the Arctic circumpolar vortex. Owing to the strong southerly and northerly components of the main westerly current around this vortex (caused by superimposition of the long wave pattern onto the polar vortex), cold, dry Arctic air is carried far to the south over central and eastern North America and eastern Asia. Mean-

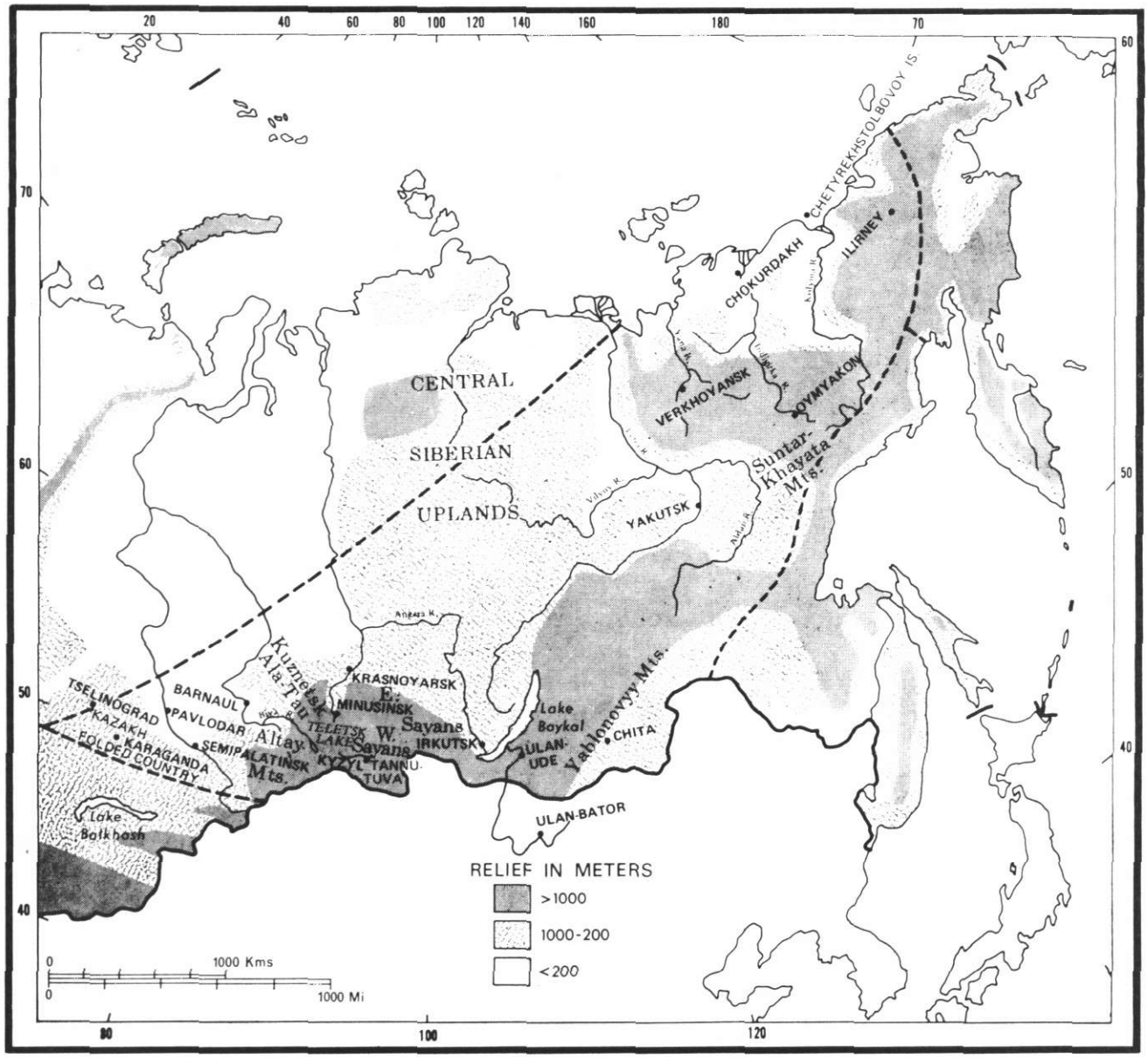

Figure 3-16. Eastern Siberia (Lydolph, 1977).

while, warm, moist air over the oceans of lower latitudes is drawn northward over the western Atlantic and Pacific Oceans toward the Arctic. In the troughs over eastern North America and Asia temperature gradients between continental and oceanic air are strongest, and here are located the major cyclogenetic regions of the hemisphere. The surface cyclonic disturbances, steered in the direction of the upper air flow, move northeast toward polar latitudes and frequently spiral in toward the upper vortex centers. As the disturbances approach high latitudes, they generally lose much of their temperature contrast, and the warm air is undercut and lifted off the ground by the colder air (i.e., the cyclone becomes occluded).

Over the north Pacific Ocean (Bering Sea area) and the Atlantic Ocean (Davis Strait area) many storms stagnate and fill. Occasionally cyclones are regenerated at the ice-water

margin, and storms continue toward the central Arctic Basin. Anticyclogenesis occurs just east of the major ridges, with the Yukon showing a particularly high frequency. Occasionally systems move out to the southeast, but they can remain stationary for many days at a time.

3.4.1 Alberta Low

Cyclogenesis occurs over Alberta where systems from the Pacific produce closed surface lows on the eastern (lee) slope of the Rockies. These cyclones generally move east-southeast to a center of maximum frequency over the Great Lakes region where tracks converge from the southwest and south. Most of these disturbances are then steered northeast to Newfoundland. Here they are joined by cyclones from the east coast of the United States, an area that favors cyclogenesis because of thermal contrasts between the cold continental and warmer maritime air masses. From Newfoundland storms generally migrate either toward Iceland (their high frequency resulting in the semi-permanent Icelandic Low) or along the west or east coasts of Greenland. A few disturbances also enter Davis Strait by way of Hudson Bay or drift erratically from western Canada below the cold upper vortex center, and an area of high cyclonic frequency is located west of Greenland.

In summer the most marked change over North America is the northern displacement of the track of the Alberta Low. The Canadian prairie provinces have one of the highest frequencies of cyclonic activity in the Northern Hemisphere. Most of these storms cross Hudson Bay to the Davis Strait-Baffin Bay area where they stagnate. Over the Atlantic, to the east of Iceland, the primary track is more zonal in July with a drop in the frequency of storms entering the Arctic Basin, and a well-defined storm path extends from the eastern Atlantic into Siberia at about 60 degrees.

3.4.2 Icelandic Lows

Storms breaking away from Iceland usually move northeast to the Norwegian Sea-Barents Sea area, and a primary track continues along latitude 75 °N. These frequent storms play an important role in the climate of the northern Europe and Siberia coastlands. This primary track is joined by tracks from the Baltic, Black, and Caspian Seas. Some storms stagnate near Novaya Zemlya, others regenerate and curve toward the pole, where cyclones outnumber anticyclones by a factor of 2 to 1.

3.4.3 Greenland Lows

Another area of high cyclonic activity appears over Davis Strait and Baffin Bay. In cases where deep lows lie south of Greenland-Iceland, the easterly flow over the Greenland ice sheet will trigger rapid intensification to the west over Baffin Bay. In other cases, surface lows will often regenerate on the east side of Greenland after air passes over the huge ice sheet.

3.4.4 East Asian Lows

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

When a strong blocking ridge over the Urals weakens, surface centers move to the southeast causing outbreaks of intensely cold air over China and Mongolia. These outbreaks are associated with strong cyclogenesis off the coast of east Asia. The majority of these east coast storms are steered northeastward toward the semipermanent Aleutian Low to then stagnate or curve north. Occasionally, storms enter the Arctic Ocean over northeastern Siberia or the Bering Strait. From time to time new storms develop on the north side of these mountains after primary storms have stagnated to the south. In general, storms over the Beaufort Sea appear to be largely of the cold low type that is characterized by colder air near its center rather than around its periphery.