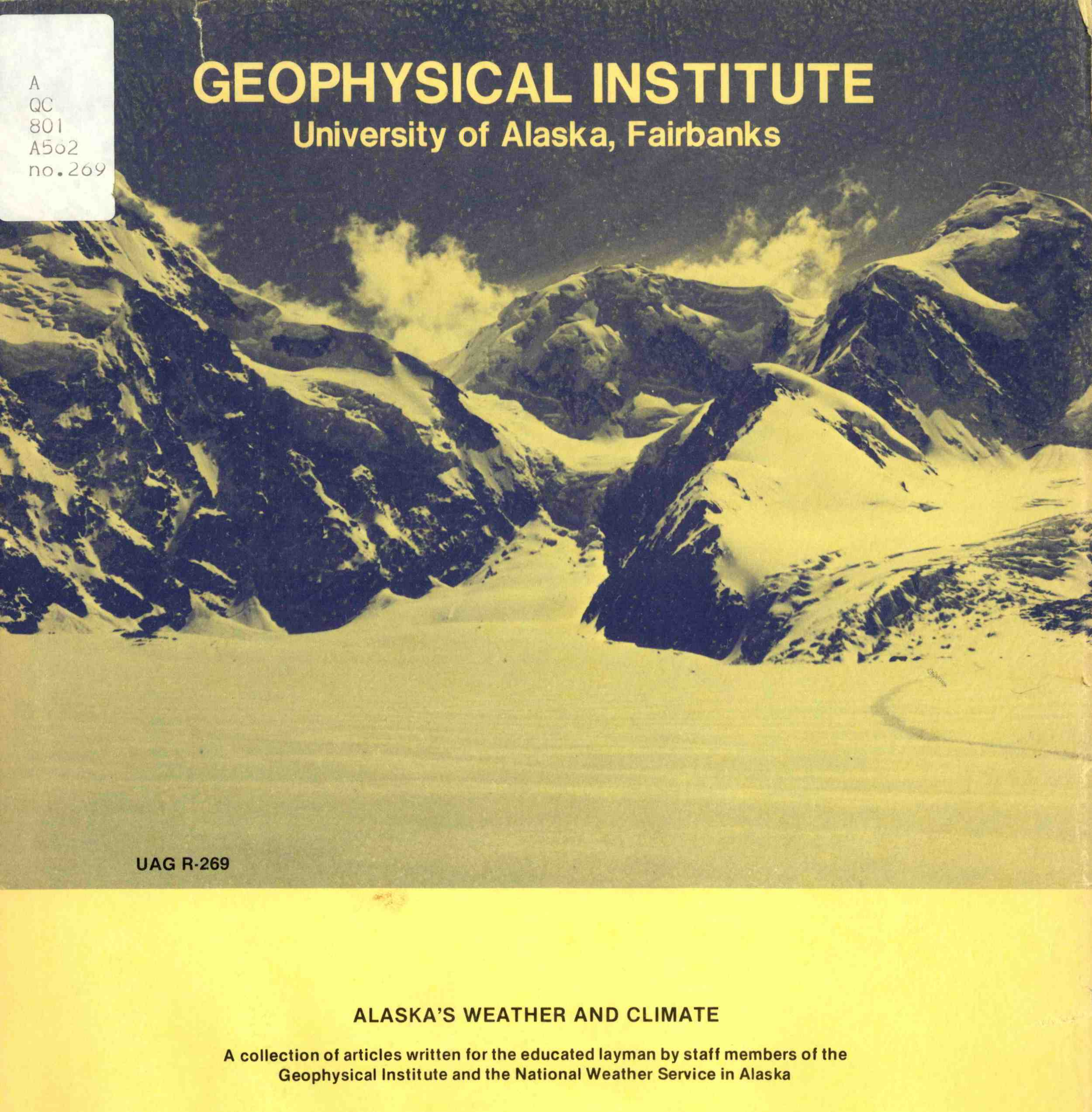
ALASKA'S WEATHER AND CLIMATE

edited by Gunter Weller

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GEOPHYSICAL INSTITUTE

University of Alaska, Fairbanks



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ALASKA'S WEATHER AND CLIMATE

A collection of articles written for the educated layman by staff members of the Geophysical Institute and the National Weather Service in Alaska

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Gunter Weller

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A collection of articles written for the educated layman by staff members of the Geophysical Institute and the National Weather Service in Alaska

Edited by

Gunter Weller

Geophysical Institute
University of Alaska
Fairbanks, Alaska 99701

July 1979

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Gunter Weller

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Preface

Alaska's weather and climate is as fascinating and varied as Alaska's magnificent scenery. In this special report for the educated layman, Alaskan meteorologists, climatologists and geophysicists explain some of the atmospheric phenomena over Alaska and their causes and consequences. Sue Ann Bowling's paper on general weather and climate patterns, which also touches on permafrost, storm surges and floods, sets the stage for the subsequent papers. Richard Augulis' climatology of winter temperature maxima and minima is an example of how Alaskan climatology deviates from the expected norm. Among the climatic controls of Alaska is sea ice, which is extensive in the ocean basins to the west and north of the State and undergoes large annual fluctuations, as described by Bruce Webster. The oceans are also the source of clouds and precipitation which are the subject of Takeshi Ohtake's paper, in which topics ranging from arctic stratus clouds to ice fog and weather modification are discussed. Most of the precipitation ends up as snow on the ground, which lasts up to nine months without melting in some of the snow provinces of Alaska, as explained by Carl Benson. When it melts, this usually occurs almost explosively, and the annual spring breakup is a period of mud, ice jams on the rivers and floods as described in Gerald Nibler's paper. The reasons for snow and ice melt are explained by Gerd Wendler and Frank Eaton in terms of the heat energy balance and its control of microclimates. The energy balance also determines the near-surface mid-winter climate in the interior of Alaska, by creating steep temperature inversions which pose serious pollution threats even in small communities, as pointed out in Carl Benson's and Ken Rizzo's paper. In fact, surprisingly, air pollution is a widespread phenomenon throughout the Arctic Basin, the subject of the next paper by Glenn Shaw. As we move further away from the

near-surface and tropospheric environment and towards the upper atmosphere over Alaska, a number of intriguing and beautiful phenomena occur.

Robert Fischer describes the noctilucent clouds and Syun Akasofu the aurora, that shimmering and pulsating curtain of lights in the Alaskan night sky. Both these phenomena have links or probable links with weather and climate as pointed out in the last two papers.

To help the reader locate the place names mentioned in the various papers, a map precedes the papers. The map is a photograph of part of a 2-meter diameter revolving globe, showing mountains and ocean trenches in three-dimensional relief which is located in the lobby of the Geophysical Institute of the University of Alaska at Fairbanks.

I would like to express my appreciation for the patience shown to the editor by all the authors, and to Dr. Sue Ann Bowling for her help in putting together the material for this report.

Gunter Weller
Editor

Index map of Alaska. The heavy line shows the approximate routing of the Alaska Pipeline. Code numbers represent the following:

Population Centers:

1. Anchorage
2. Aniak
3. Barrow
4. Barter Island
5. Bethel
6. Fairbanks
7. Fort Greeley
8. Propect Creek
9. Valdez
10. Whitehorse, Yukon Terr.
17. Juneau (and Juneau Ice Field)

Mountains:

11. Chugach Mountains
12. Kenai Mountains
13. St. Elias Mountains
14. Talkeetna Mountains
15. Wrangell Mountains
16. Mt. McKinley

Other Land Features:

17. Juneau Ice Field
18. McCall Glacier
19. Worthington Glacier
20. Copper River Basin
21. Glacier Bay National Monument
22. Kenai Peninsula
23. Lake Minchumina

Islands:

24. Kodiak Island
25. Pribilof Islands
26. St. Lawrence Islands

Other Ocean Features:

27. Bering Strait
28. Cook Inlet
29. Kotzebue Sound
30. Morton Sound
31. Prudhoe Bay



Map of Alaska with Place Names

ALASKA'S WEATHER AND CLIMATE

Sue Ann Bowling
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Alaska - The Big Land. The 49th state spans 20° of latitude, from Point Barrow at 71° to Amatignak Island in the Aleutians at 51° N, and 57° of longitude, from Attu Island to Ninemile in the southeast "pan-handle". Slide a map of Alaska over the 48 contiguous states, and you will find that it reaches from the Atlantic to the Pacific, from Canada to Mexico (Figure 1). The altitude range within the state is equally impressive - 6194 meters from sea level to the top of Mount McKinley, compared with 4504 meters from Death Valley to the top of Mt. Whitney in the "lower 48" states. The Alaskan coastline exceeds in length that of all other 49 states combined. Is it any wonder that the range of climatic conditions in Alaska rivals that of the 48 contiguous states as a whole?

Factors Determining the Alaskan Climate

The climate of Alaska is dictated by its latitude, the character of the surrounding land or water, the physical relief of the state, and the interaction of all these with the global atmospheric circulation. Of these factors, latitude is the one in which Alaska differs most from the other states and contributes to much of the variation within the state as well. For example, in the southernmost Aleutians the hours of possible sunlight vary from 8 in winter to 16 in summer. At Anchorage (61°N), the range is from 5 1/2 to 18 1/2 hours. At Fairbanks (65°N), the extremes are 3 3/4 hours and 20 hours. This far north, the midday solar elevation is so low in winter (1°44' on December 21 and less than 3° 30' from December 1 through January 12 - Figure 2) that solar heating is

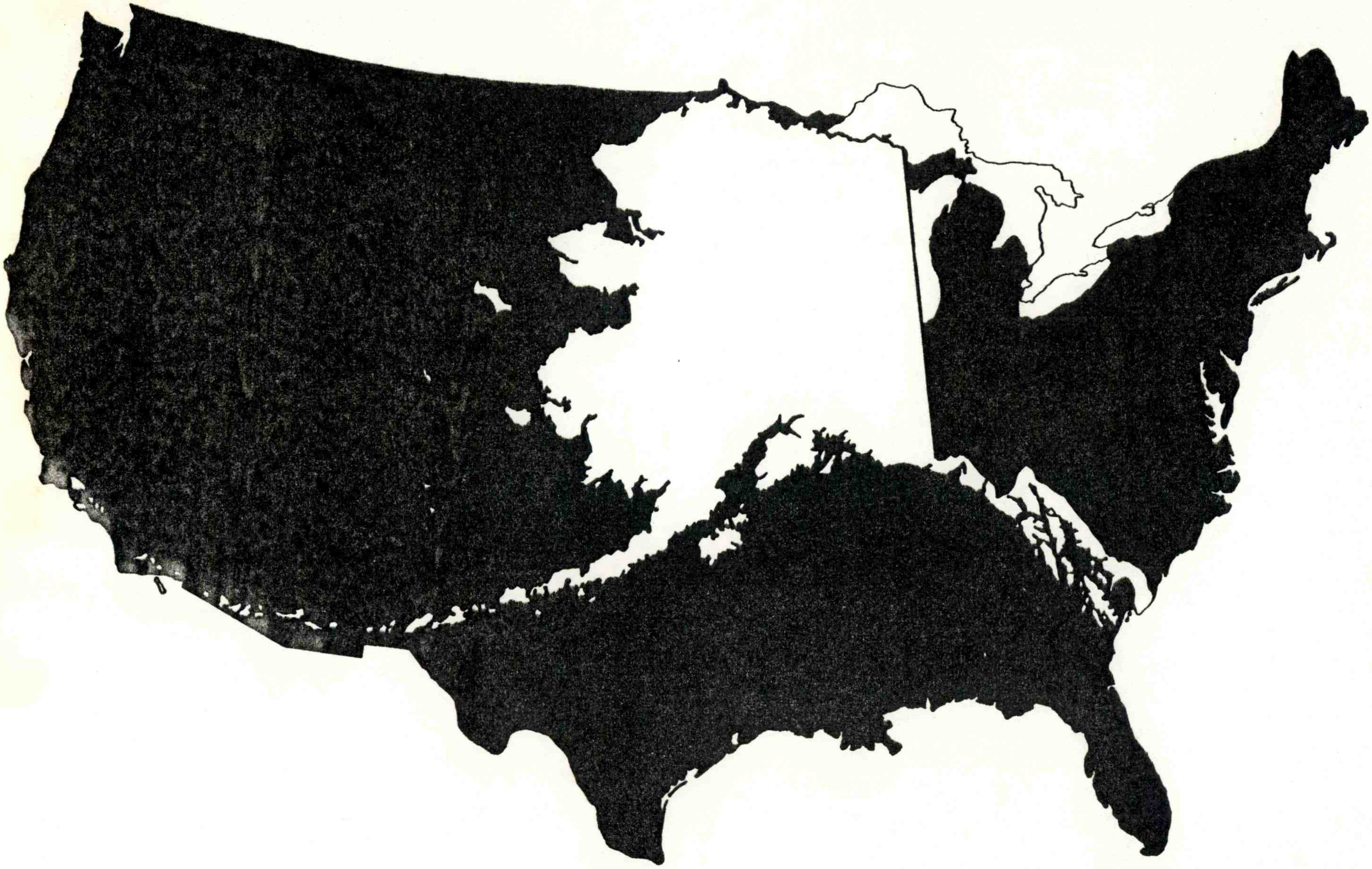


Figure 1: Relative sizes of Alaska and 48 contiguous states.

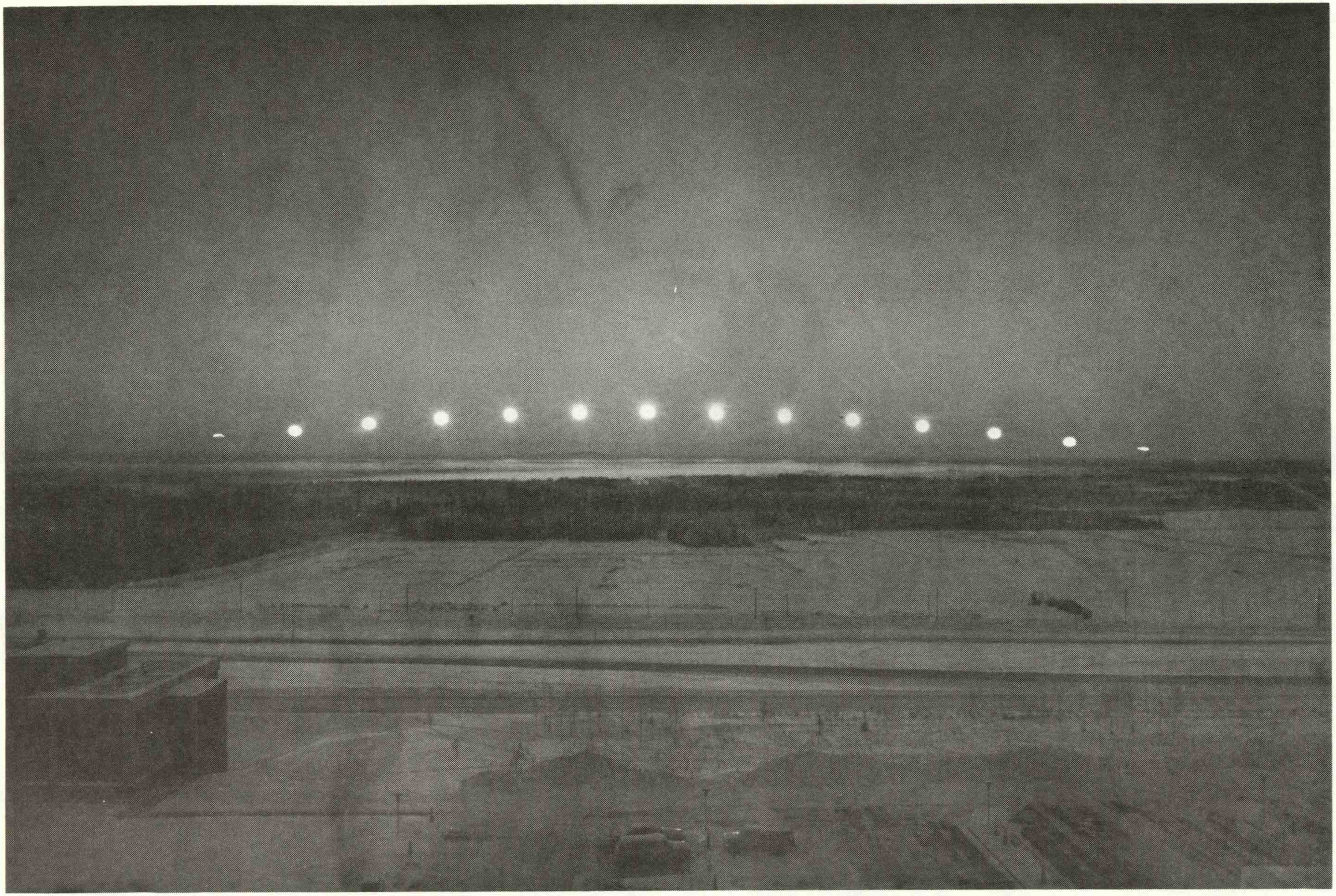


Figure 2: Multiple exposure of the solar path as seen from Fairbanks, 21 December 1972. Photo by Takeshi Ohtake.

negligible - the highest temperature of the day is almost as likely to be at midnight as at noon (see Augulis' article in this report). In summer, on the other hand, there are almost three months when "night" is no more than a brief interval of twilight and it is possible to work outdoors without artificial light at midnight. At Barrow (71° N), there is a two month period in summer when the sun never sets, and a period almost as long when "day" is only a few hours of twilight. Thus in midsummer relatively low solar elevations and long days compensate for each other and the potential insolation does not vary much over Alaska. In midwinter, the total possible solar radiation north of the Alaska Range is essentially zero, so variation of possible insolation with latitude is confined to the southern part of the state. In spring and fall the possible insolation depends strongly on latitude.

The importance of factors other than latitude, however, is brought out by the fact that the southernmost part of the state, the Aleutians, is treeless while the treeline in the Interior reaches north of the Arctic Circle. This is due to the second major influence on climate and weather across the state - the degree of exposure to oceanic air.

The mainland mass of Alaska is a large peninsula of North America. The oceans which form the northern, western and southern boundaries of the state differ markedly in their thermal characteristics and hence in their effects on climate. The Arctic Ocean to the north has year-round sea ice, and while the ice may be driven well offshore in the summer months, the water temperatures are little higher than the temperature of melting ice. In winter, the sea ice is almost an extension of the continent as far as its thermal behavior is concerned. To the west are the Chukchi and Northern Bering Seas, both with seasonal sea ice, and the eastern tip of Siberia. To the Southwest and south are the Southern Bering Sea,

the North Pacific and Gulf of Alaska, all of which are mostly free of sea ice. The major North Pacific warm current, the Kuroshio, does not enter the Bering Sea or penetrate as far north as the Atlantic's Gulf Stream, but it does provide some warm water to the Gulf of Alaska.

A number of mountain chains within the state act as barriers to the free circulation of oceanic air. Along the coast of the Arctic Ocean is the Arctic Coastal Plain, from which rises the Brooks Range-the northernmost extremity of the Rockies. South of the Brooks Range is the Interior - a jumble of unglaciated uplands and broad river valleys. The Alaska Range, dominated by Mt. McKinley, comes next, followed by another complex of lowlands and mountains (glaciated this time) and the Coast Range complex. The overall arrangement of east-west oriented mountain ranges is such that unmodified oceanic air can reach the interior only from the west, while the Arctic slope and the Coast Range are strongly affected by air masses from the Arctic and Pacific Oceans respectively. Major climatic zones within the state are controlled almost entirely by topography.

Climatic Zones

The southernmost barrier, the Coast Range complex, starts in the south-eastern "panhandle" as the Alexander Archipelago. This part of Alaska, which consists entirely of the seaward slope of the Coast Range, has an extreme maritime climate and many glaciers. Mean yearly precipitation near sea level ranges from 1 to 5 1/2 meters of water equivalent, with most stations having over 2 1/2 meters a year. Precipitation is heavy throughout the year but is heaviest of all in autumn. Sea level daytime temperatures are normally above freezing even in January, although nighttime freezes are expected. Summer temperatures,

however, are cool - 10° to 15°C at sea level. Cloud cover is extensive and less than 100 days a year escape overcast skies at most stations.

Farther north, the Coast Range is made up of the glacier clad St. Elias Mountains, the Chugach Mountains, and the Kenai Mountains, with Kodiak Island as its last outpost. This region has more glaciers than any other part of Alaska. It is also highly active tectonically and has produced several major earthquakes, including the 1964 Good Friday earthquake. The seaward slope make up the south coast climatic division, which is slightly cooler and drier than the southeastern division, primarily due to its higher latitude. The landward slopes are considered to be part of the transitional climatic zone.

The major volcanic chain of Alaska, which includes the Aleutian Range, the Alaska Peninsula and the Aleutian Islands, is topographically a continuation of the Alaska Range into a classic island arc. Climatologically, however, this chain corresponds to the Coast Range, rather than the Alaska Range, westward from Kodiak Island. The climate is extremely maritime and very stormy with the highest regional winds in Alaska. Precipitation here is only about half of the south coast-southeastern average, due primarily to lower sea surface temperatures and the far smaller influence of mountains in the Aleutians. Midwinter temperatures hover around the freezing point. Daily minimum temperatures in July are around 7°C throughout this area, while daily maximums range from 17°C along Cook Inlet to 10°C in the southernmost part of the Aleutian Islands. As a result of these low summer temperatures and high winds, virtually the entire area is treeless.

The "lowland" area between the Coast Range and the Alaska Range is interrupted by two mountainous areas. The volcanic, glacier covered, Wrangell Mountains are separated from the Chugach Mountains by the

rather narrow Chitina Valley. The Chitina Valley heads near the Canadian border in the major ice field of the St. Elias Mountains. The Talkeetna Mountains, north of the Kenai Peninsula, are much lower, with glaciation confined to scattered cirque glaciers. Between the Talkeetna and Wrangell Mountains lies the Copper River basin--once an ice-dammed lake, now draining to the Pacific via a narrow canyon through the Chugach Mountains. Although this mountain-walled basin is considered part of the transition zone between maritime and continental influences, its precipitation (37 cm/year) and January temperatures (-20°C) are typical of the interior continental climate. Summers are cooler at night but warmer in the daytime than in the fully maritime zone. The fact that summers are cooler than in all but the northernmost part of the interior zone can probably be attributed to the relatively high elevation of the valley floor. The combination of cool summers and cold winters results in widespread permafrost in the Copper River Basin.

West of the Talkeetnas, the Cook Inlet division of the transitional zone includes the Matanuska and Susitna Valleys and their continuation into the lowlands on both sides of Cook Inlet. This is the most densely settled area of Alaska, and includes the largest city in the state, Anchorage. There is a great deal of variability across this division, but the mean precipitation is around 50 cm. The mean July temperature is around 14°C and January temperatures are around -12°C. Temperatures below -40°C have been recorded in the Susitna Valley, but are not common. Summer temperatures above 30°C are also rare, and most of the precipitation occurs in summer. High winds are not uncommon during the winter months. Permafrost occurs only sporadically.

A different type of transitional zone climate occurs on the Bering Sea coast, in the Bristol Bay and West Central climatic divisions. Here the ocean is partly ice-covered in winter and spring and ice-free but cold in the summer and fall. The result is a maritime climate with coastal conditions only slightly warmer and dryer than the Aleutians in summer, and stormy but cold winter weather. The January mean temperature for the West Central division is around -16°C, and temperatures below -40°C are not uncommon. The low summer temperatures allow trees to grow only in a few favorable locations, and most of the area is underlain by permafrost.

The moderately glacier-covered Alaska Range runs in a southward-concave arc from the north end of the Aleutian Range to the Canadian border, reaching a maximum latitude of nearly 64° N near Fairbanks. The northward-concave Brooks Range stretches across the state at around 68° N. Between these northernmost equivalents of the Sierra Nevada and the Rocky Mountains lies the equivalent of the Basin and Range province of the Southwestern United States - a jumble of broad, flat valleys separated by unglaciated rolling uplands of moderate relief. The valleys, where most of the population is located, have the most continental climate in the state. Yearly precipitation totals range from less than 20 to over 40 cm, with a clear August maximum. Normal January temperatures are around -25°C, while July values are about 15°C. Summer temperatures have exceeded 35°C on occasion and 30°C rather frequently, and warm-season crops such as corn and tomatoes may be grown outdoors where the soil temperatures can be increased sufficiently (Figure 3). Frosts <u>may</u> occur during any month of the year, but are rare from June through mid-August. Thawing periods of more than a few hours duration are practically unknown



Figure 3: Corn growing, interior Alaskan style. Notice the difference between corn grown in unprotected soil (center) and through clear plastic (other rows). The plastic warms the soil by suppressing sensible and latent heat transfer. Interior air temperatures are generally excellent for annual plants if soil temperatures can be kept high enough. Photo by Don Dinkel, Fairbanks Agriciltural Experiment Station.

from November through February, although daytime high temperatures may reach 4°C in any month. Consequently, there is virtually no melting in the snowpack from November until the March-April breakup season (Benson's snow cover article - this report). At Fairbanks a winter without at least a day or two with -40°C temperatures is unusual, and temperatures below -45°C occur in about one winter out of two. -50°C is unusually cold, but not unknown. These low temperatures are invariably due to steep ground inversions (air pollution article by Benson and Rizzo in this report) and cannot be maintained in the presence of any perceptible wind. The lowest temperature ever recorded in Alaska, -63°C, occurred at Prospect Creek in the Interior basin. Permafrost varies from sporadic to continuous within this zone.

The Arctic Drainage division of Alaska includes everything north of the crest of the Brooks Range as well as the west coast north of the Bering Strait. Precipitation is quite light, less than 20 cm a year. Trees are rare and confined to sheltered areas in the Kotzebue Sound drainage, south of the Brooks Range. Mean summer temperatures are about the same as those in the western Aleutians, although stations along the Arctic Coast are colder - Barrow, at the northernmost tip of Alaska, has a mean July temperature of 4°C. Cloud cover is infrequent in spring but almost constant in summer (Ohtake's article in this report). Midwinter mean temperatures are marginally colder than those of the Interior, but the spring warming is far slower and February rathern than January is usually the coldest month. Extreme low temperatures are not as severe as in the Interior, but low temperatures in the Arctic are less dependent on inversions than those in the more sheltered inland valleys, and the combination of low temperatures with high winds is frequent. Permafrost is continuous and extends to considerable depths in the Arctic drainage.

Mountain climates in Alaska have not yet been discussed, and in fact there is very little long-term data available from mountainous areas in Alaska. Mountain areas have generally lower summer temperatures than the adjacent lowlands, and their yearly precipitation is higher - mostly in the form of snow. Wind speeds are also normally higher. Winter temperatures in the mountains may not differ much from lowland temperatures, and under clear-sky conditions all but the highest peaks are quite definitely warmer than the valleys. This anomalous behavior, which is clearly visible on satellite infrared imagery (Figure 4), is due to the prevalence of ground-based inversions and cold air drainage under clear-sky conditions.

Atmospheric Circulation

The seasonal progression of climate in Alaska is closely associated with the characteristics of the mean circulation of the atmosphere. For example, consider the seasonal variation of precipitation. Aside from the Pacific coastal zone, April is the driest month and August is the wettest in every climatic division and at most individual stations. In the Pacific coastal region the driest month is June and the wettest is October. This situation can be partially explained by sea water temperatures and hence atmospheric water contents being minimum in spring and maximum in fall, but 700 mb mean flow and topography suggest an additional factor. During April and May, a strong ridge usually develops over the Yukon and Alaska. Mountain barriers, including those on the Kenai Peninsula, shelter all but the immediate coastal area from the south to north moisture transport. As the season progresses, the onshore flow becomes less intense and the ridge flattens (Figure 5). By August, the flow is southwesterly, almost parallel to the mountain barriers and ideally aligned to bring moisture up the Cook Inlet and



Figure 4: NOAA 4 thermal infrared composite image of Alaska in winter. Note the contrast between cold (light) valleys and warmer (darker) mountains.

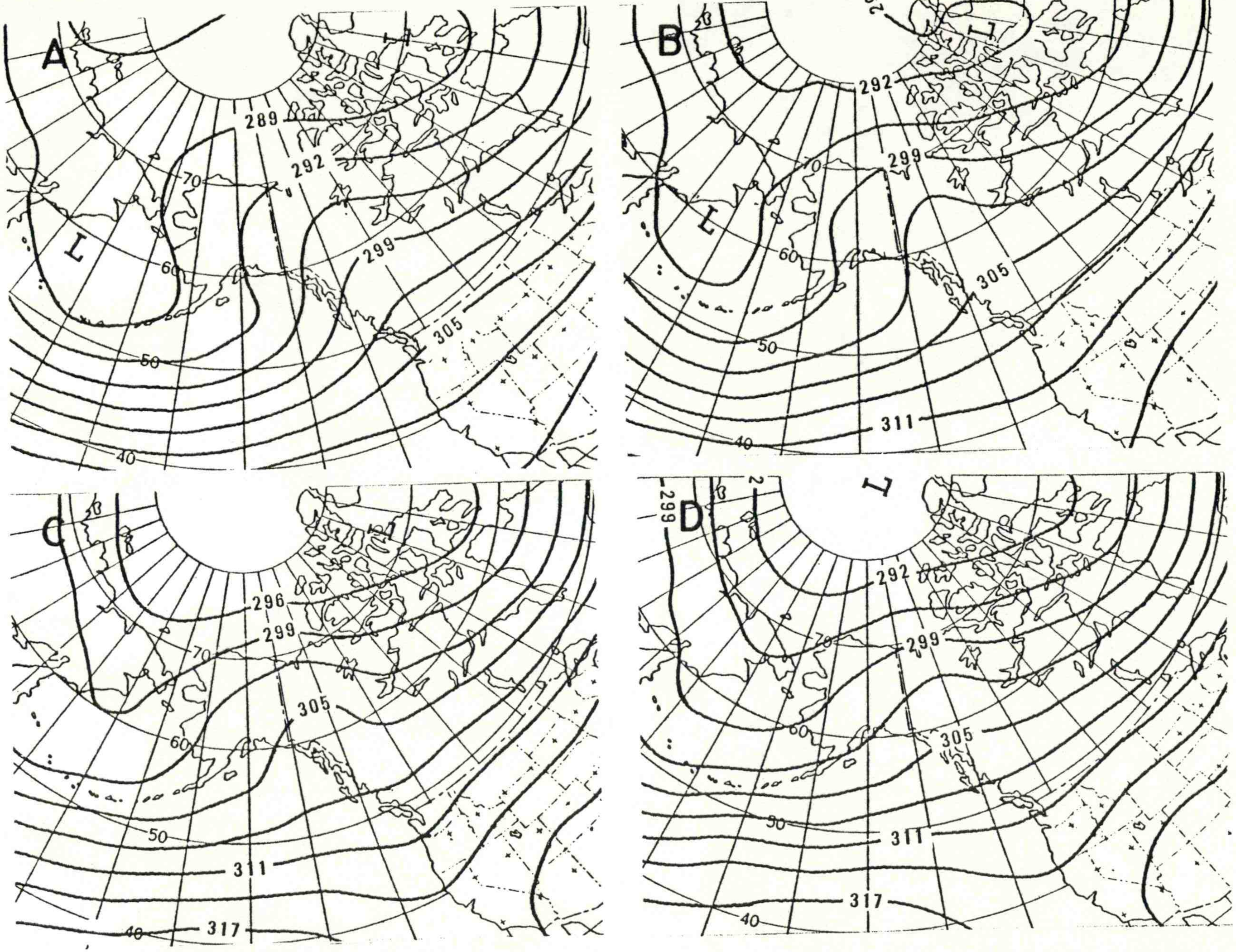


Figure 5: 700 mb normal maps, contours labeled in 10's of meters. A-May; B-June; C-July; D-August. Maps courtesy of NOAA Extended Forecast Division.

around the west ends of the Alaska and Brooks Ranges. The relationship between mountain barriers, upper winds and precipitation is demonstrated by years such as 1977, when a weaker-than-normal ridge in June strengthened, rather than decayed, through August. As a consequence, heavy rains through most of the state in June were replaced by hot, dry conditions in August.

Winter circulation patterns cannot be so neatly tied to an orderly sequence. Certain patterns do persist over periods of weeks, associated with distinct spells of different weather types, but with one possible exception they do not occur in any orderly sequence. Individual winters may be dominated by just one or two patterns, or may have a chaotic mixture of types. Two excellent examples of winters dominated by one or two circulation types were the winters of 1970-71 and 1976-77.

In order to understand the impact of the winter 1970-71 in interior Alaska, and particularly in Fairbanks, it must be kept in mind that, although snow normally covers the ground from October to April, the average snowfall does not exceed 35 cm a month, and the total accumulation is usually around 175 cm. During the last three months of 1970, the 700 mb flow over Alaska strongly resembled the normal August pattern. Precipatation along the Pacific Coast was only about 60% of normal but the Interior lowland snowfall was two to three times higher than the normal. At Fairbanks, a snowfall of 137 cm set a new record for November and was the second highest montly total ever recorded. By late December, almost 3 meters of snow had fallen and remained on the ground. Moose began to have difficulty traveling in the deep snow, and, in many cases, preferred to challenge traffic on plowed roads. Snow-removal crews found it increasingly difficult to keep even the main roads clear.

Fairbanks residents are accustomed to driving on packed snow, which at normal December temperatures of around -20°C to -25°C is not as slippery as one might suppose (salting to remove ice and snow is impractical at temperatures substantially below freezing), but deep snow for weeks on side roads, a night of rain on top of the cold-soaked snow and the moose on the roads were too much even for the local population.

When the August type of 700 mb pattern occurs in the first half of the winter, it tends to evolve into a pattern with northerly flow dominating across Alaska (Bowling, 1977). In 1970-71, this change began shortly after Christmas and was complete by mid-January (Figure 6). Incessant snow was replaced by frigid temperatures and ice fog. The January monthly mean temperature of -35.4°C was the lowest official monthly mean in Fairbanks history. From January 14 to the end of the month the minimum temperature never rose above -40°C and the maximum temperature was -30°C or colder. Five days had minimum temperatures of -50°C or lower, and the temperature was continuously below -40°C for over a week. The worst of the cold weather ended by February, but temperatures continued below normal until May.

Not all winters with stable circulation types are miserable, however. If southerly flow predominates, as in the usual April circulation, generally above-normal temperatures are the rule. However, this type of circulation seems to be stable only if it is part of an amplified west coast ridge-east coast trough over all of North America. The results of such a circulation were superbly demonstrated by the winter of 1976-77, particularly by the month of January (Figure 7). While the Ohio Valley had its coldest month on record, Alaska temperatures were above normal everywhere except in the westernmost Aleutians. Departures ranged from

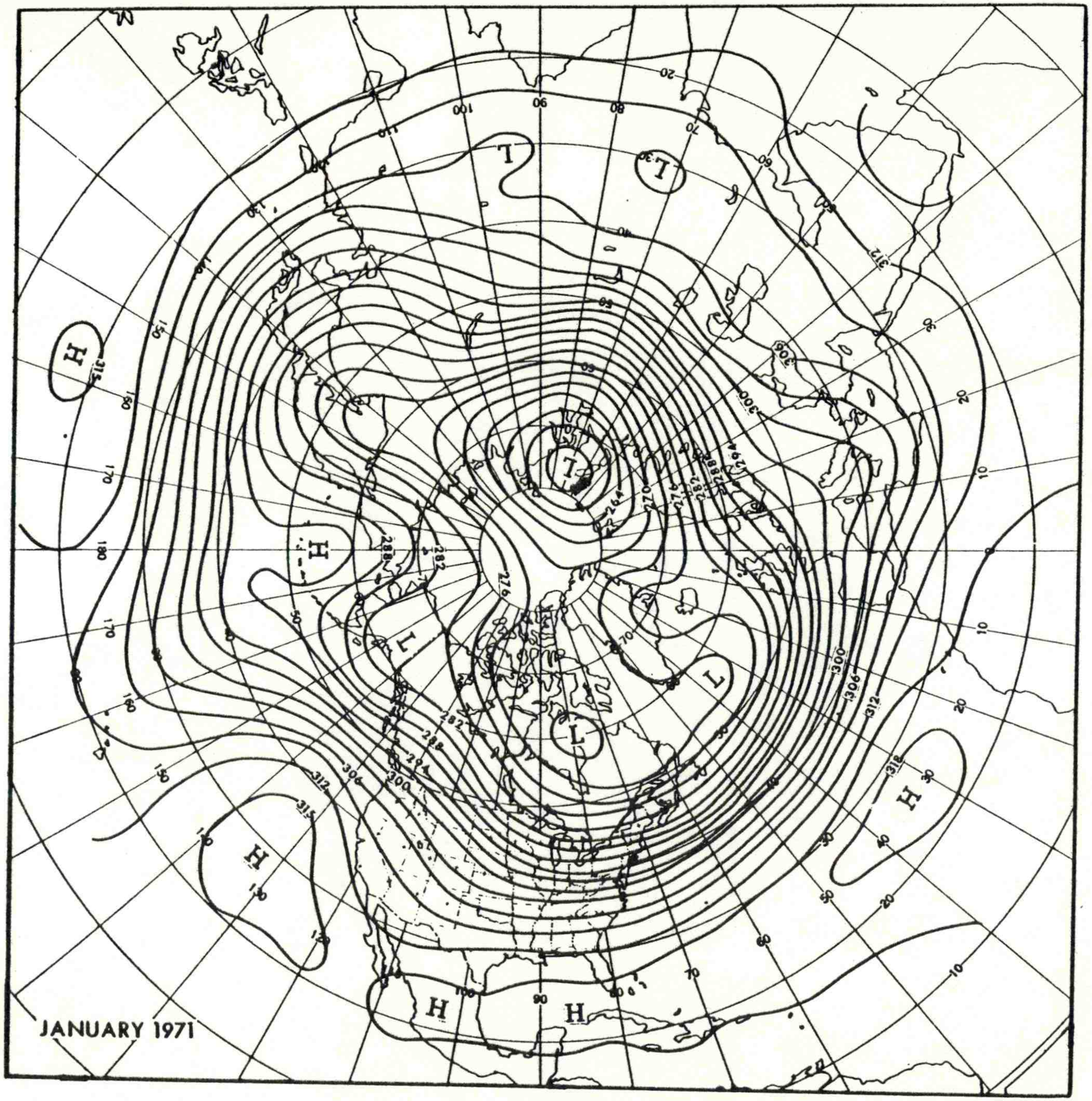


Figure 6: 700 mb mean heights, January 1971. Contours labeled in 10's of meters. Map from Posey, 1971.

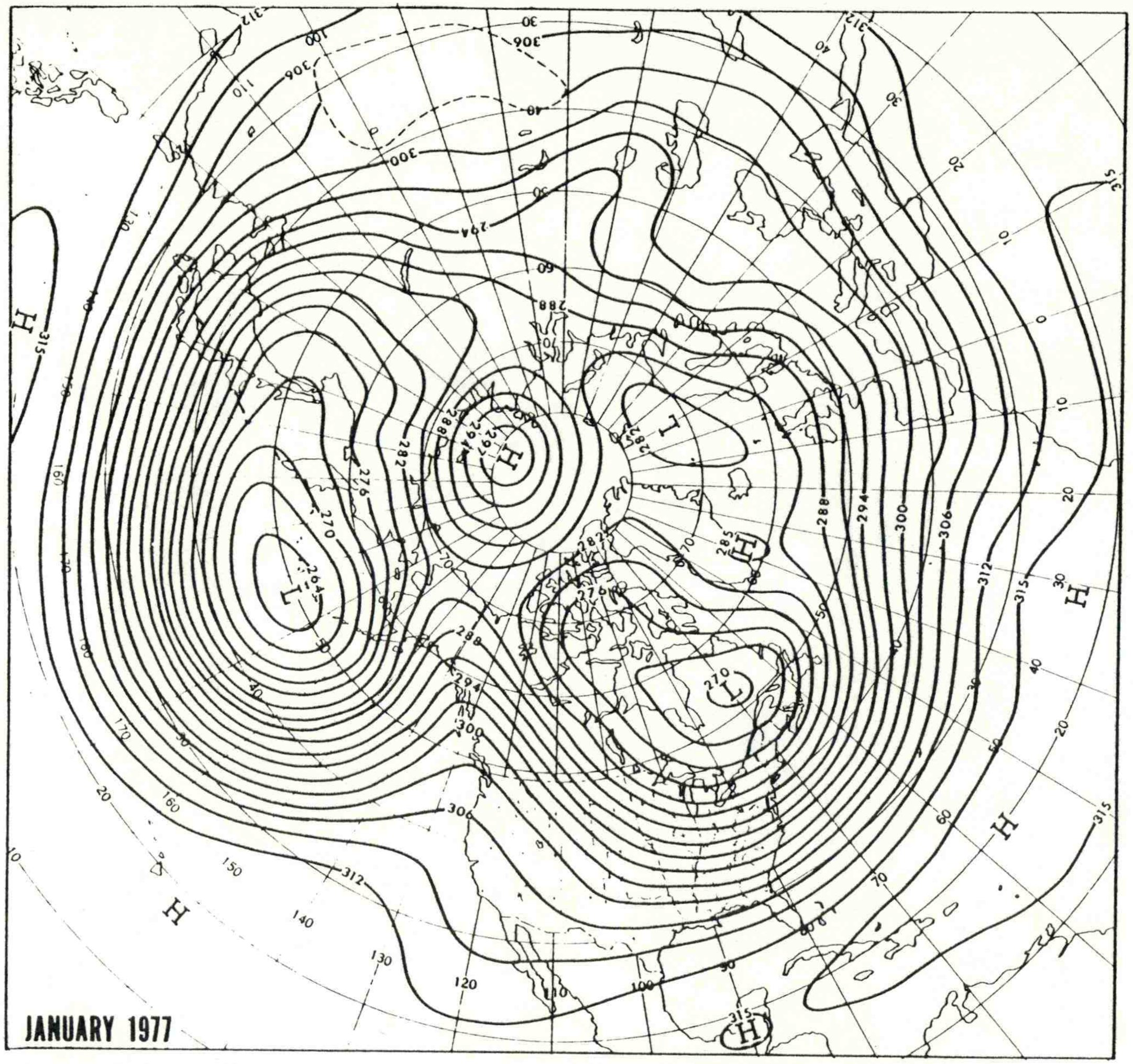


Figure 7: 700 mb mean heights, January 1977. Contours labeled in 10's of meters. Map from Wagner, 1977.

+4°C in Southeastern Alaska to +10°C and more in the Cook Inlet and Interior regions. With upper air flow perpendicular to the major mountain chains, orographic effects dominated the distribution of precipitation. Although cloudiness was above normal, precipitation was below normal at low-lying inland sites. South coast stations, however, had double and triple their normal precipitation, and seasonal snowfall at mountain snow-gaging stations throughout the state was high. At Fairbanks, 1976-77 was the first winter on record during which temperatures never dropped to -35°C; in fact the lowest temperature observed through the end of February was -31°C. In March the persistent longwave pattern broke down and Alaska received its first seasonable weather of the winter, but the minimum temperature observed at Fairbanks was still only -33°C.

Extreme Weather Events

Even a superficially normal circulation pattern can produce a decidedly abnormal weather occurrence - such as the Fairbanks flood of August 1967. Within Alaska, the overall direction of flow was normal, but there was a good deal more confluence and faster flow than usual (Streten, 1967; Figure 8). The fast westerly flow was actually established in July, and the July precipitation at Fairbanks and throughout the Tanana drainage was well above normal. In fact, flooding of the Tanana River, sufficient to disrupt normal rail service from Anchorage to Fairbanks, occurred at Nenana July 25-29, and August started with rivers high and the ground saturated throughout the Tanana drainage. Although southerly flow brought some clearing of skies during the first few days of the month, resumption of westerly flow resulted in over 140mm of additional precipitation - almost 2 1/2 times the August normal - at Fairbanks in the 7 days from August 7 through

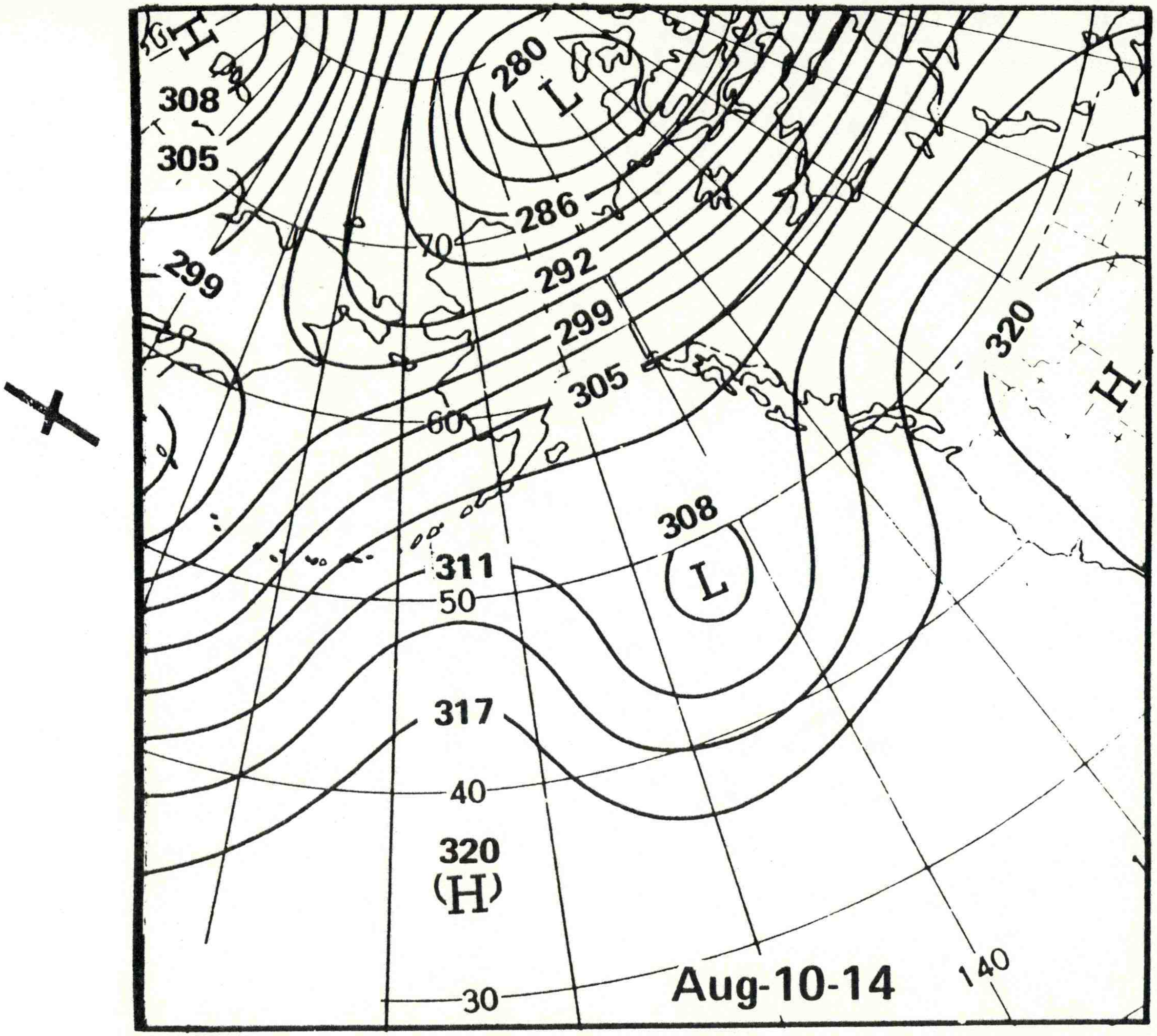


Figure 8: 700 mb mean heights, 10-14 August 1967. Contours labeled in 10's of meters. Map from Posey, 1967.

August 13. By the afternoon of the 11th, a decaying tropical circulation in the North Pacific was feeding moisture into this confluent flow and the precipitation on the 12th - over 85mm - set a 24 hour record that still stands. Orographic effects probably produced even greater amounts upstream in the basin of the Chena River, the tributary to the Tanana which runs through Fairbanks. However, neither rainfall nor river level data were available at that time from any point upstream from Fairbanks. The Chena passed flood stage at Fairbanks about midday on the 13th. By the 14th, both the Richardson Highway and the Alaska Railroad - the only surface routes from Fairbanks to Anchorage or Canada - had been cut by rising waters in the Tanana River. The Chena had meanwhile risen enough to force evacuation of a few families in low-lying areas, and a predicted crest of around a meter above flood stage that night led to sandbagging operations around the hospital. Water was well above the predicted crest and still rising on the 15th, however, and the crest - which occurred that afternoon - was more than 2 meters above flood stage. Virtually everthing inside the city limits of Fairbanks, as well as a substantial area around the city, was flooded (Figure 9). Luckily, the runways at the airport stayed above water, and the airport, the high school and the University of Alaska (built on a hill) were all packed with evacuees. The river dropped back to 1 meter above flood stage on the 19th and below flood stage on the 21st. Total damage from this flood was estimated at 200 million dollars. In view of the rapid and unanticipated early-morning rise on the 15th, the confirmed death toll of 5 was surprisingly low. (As an example of how little warning was available during the final stages of the rise in water levels, the author, then living near the edge of the flood plain and working at the University, felt extremely foolish about loading both



Figure 9: "Traffic jam" near the University of Alaska close to noon on August 15. The interaction shown had been dry an hour and a half before the photo was taken and was impassable to vehicles an hour and half later.

cats, their food and a sleeping bag into the car before driving to work over dry streets the morning of the 15th - but somewhat less foolish after four nights of sleeping on the office floor until the water level dropped enough so that it was possible to wade home). River-level and rain gages have since been installed upstream on the Chena with data being telemetered directly or via satellite into the Fairbanks Weather Service office.

Satellites have been a boon to Alaskan forecasters in more ways than serving telecommunications. Alaskan weather tends to come from the North Pacific (with one permanant station - weathership Papa - between Alaska and Hawaii), from Siberia, or from the Arctic Ocean (with no permanent weather observing stations). Today, satellite images in both visible and thermal infrared wavelengths allow storms to be monitored year-round in all three areas. In October 1973, however, thermal images were not available, and TIROS visual images were generally good only to about 65°N, even had the Arctic Ocean not been sinking into polar night at that time. Consequently, the record storm which came out of the Arctic Ocean to strike Barrow on October 3 was almost completely unanticipated.

Normally, sea ice tends to damp storm surges and high waves. But in this case, the pack ice boundary was an unusually distant 350 km or more north of Barrow. With winds of 25m sec^{-1} gusting to 35m sec^{-1}, the observed 3.7 m storm surge and wave heights of 3 m were devastating (Figure 10). Barrow Spit was breached in numerous places, the fresh-water supply for the Arctic Research Laboratory was inundated with seawater and an estimated 20 years worth of normal erosional processes were accomplished in less than a day. Some 3 million dollars worth of

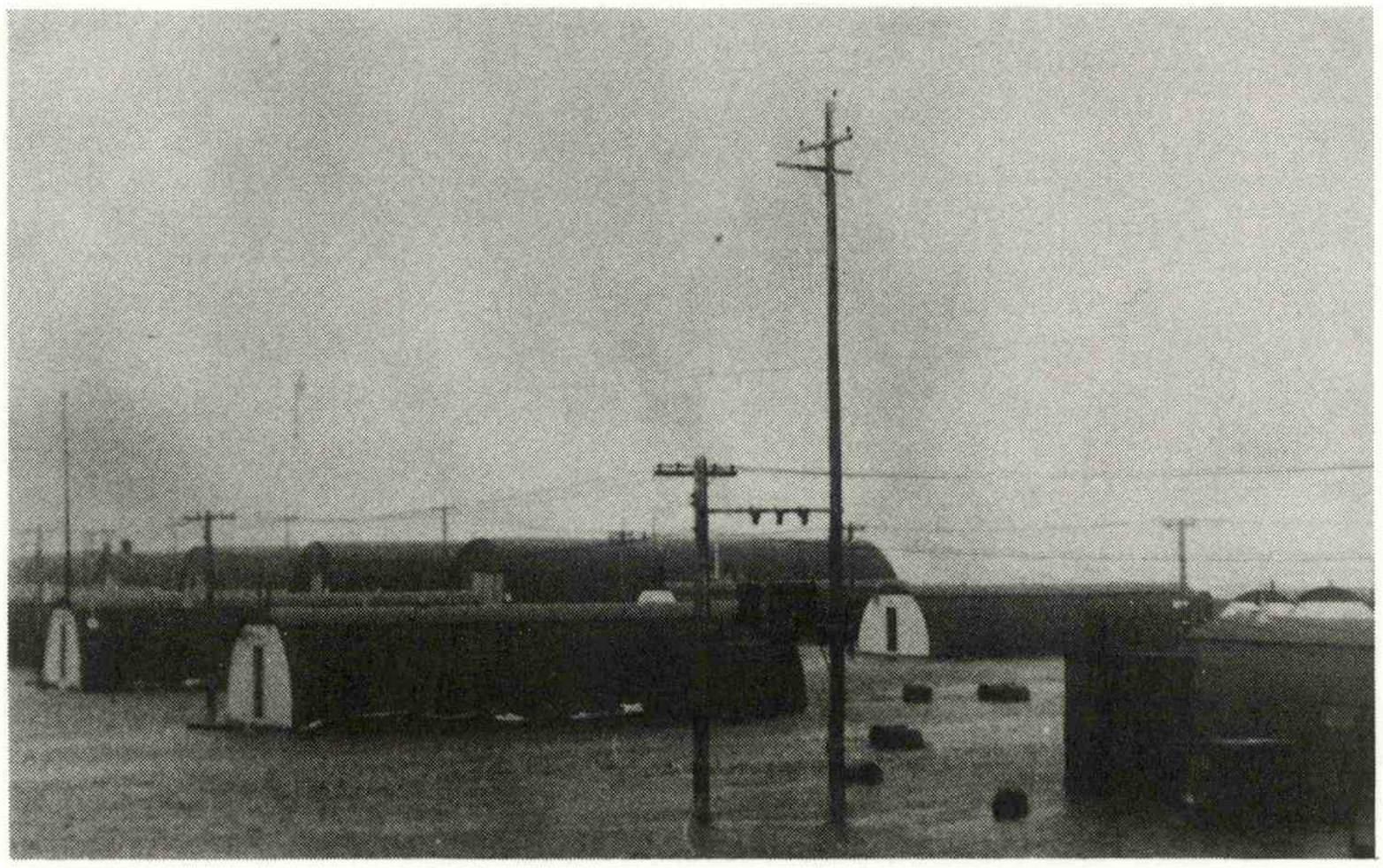


Figure 10: Storm surge at Arctic Research Laboratory, Barrow, October 3, 1963. This photo was taken by Jerry Brown several hours before the peak of the flooding. Photo courtesty of Naval Arctic Research Laboratory.

damage to government installations and $250,000 worth to Barrow village (on higher ground) were sustained. Sea ice is usually considered to be a negative factor in the Arctic environment, but in this case its absence was responsible for much of the damage (Hume and Schalk, 1967).

Conclusion

The Alaskan atmosphere, from sea level to auroral heights, includes far more of interest than can possibly be covered in a single article, or even in a report of this nature. This article and those following attempt to cover some of the highlights only.

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HOURLY CLIMATOLOGY OF DAILY HIGH & LOW TEMPERATURE OCCURRENCE

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Observations in middle and low latitudes show that, most of the time, daily maximum temperatures occur during the mid to late afternoon period, and daily minimum temperatures around the time of sunrise. This is because the maximum and minimum temperatures occur when the net energy exchange at the earth's surface becomes zero - as solar heating drops below the radiative and turbulent energy losses in the afternoon, and rises above radiative losses shortly after dawn. However, it is not always so in the polar or high latitude locations, especially not during the winter months. During mid-winter, at the latitude of Fairbanks, the sun just barely climbs above the horizon. There the winter sun reaches only about 2 degrees above the horizon at noon on December 21st, the winter solstice, and no more than 7 degrees above the horizon by the end of January. Under these circumstances, there is very little heat in the sun's rays and the temperature of the air is not greatly influenced by it. Consequently the times of daily high and low temperatures can be quite variable.

Although the sun does have some effect on the times of occurrence of the day's maximum and minimum temperatures, Figure 1 shows that a different pattern of timing is the dominant one.

The daily maximum temperature is most likely to occur near the end or beginning of the calendar day. There is a secondary preferred time around 2pm, that shows some effect of the sun. This behavior simply means that temperatures frequently fall or rise steadily all day long. This is caused by relatively slow changes of cloud cover and other weather characteristics that affect temperature over Fairbanks.

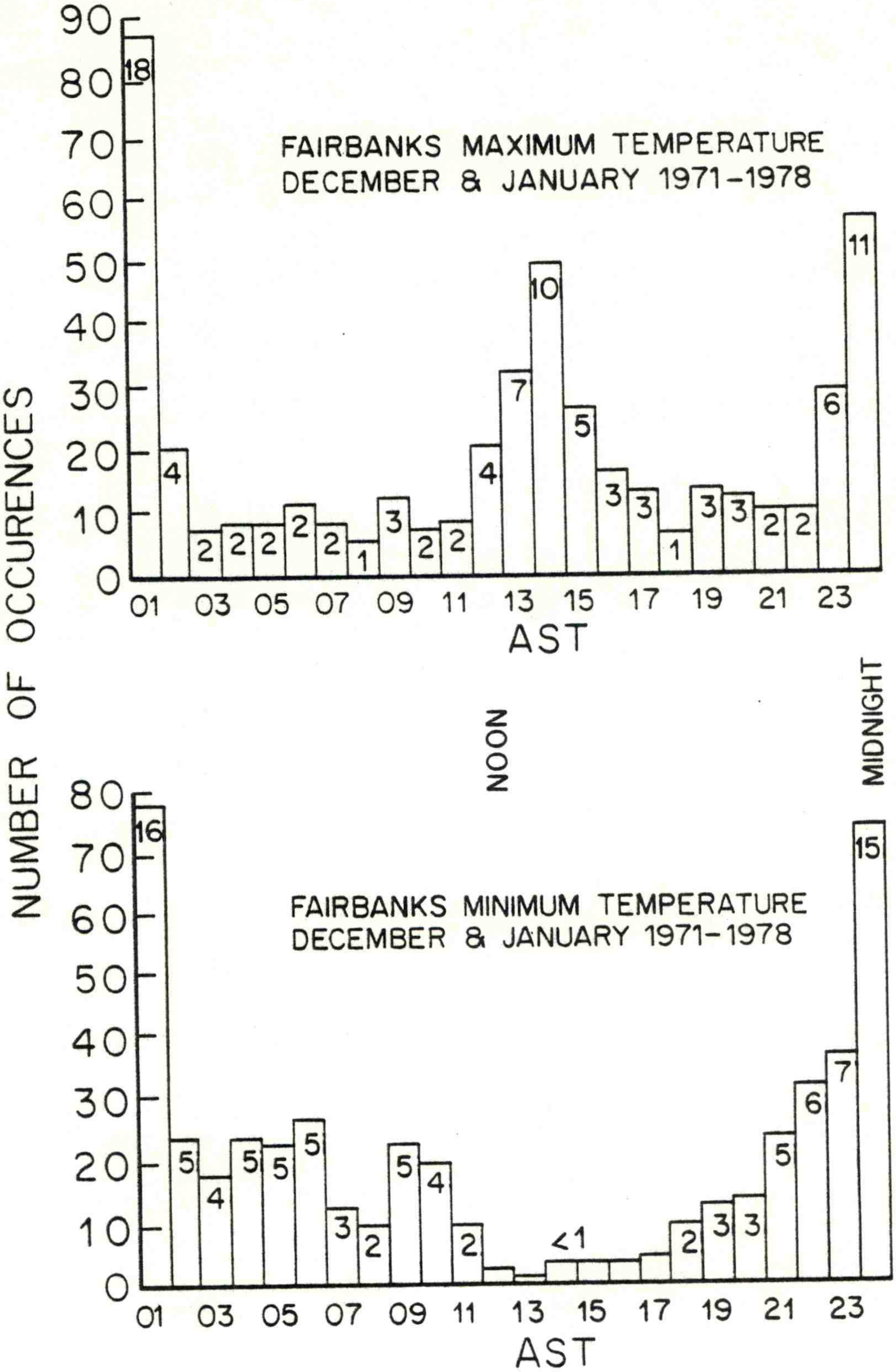


Figure 1 Number of times that the daily maximum (top) and minimum (bottom) temperatures during December and January at Fairbanks, Alaska, were reached in each hour of the day. Based on data for 1971-1978.

The daily minimum temperature shows the same tendency to occur near the end or beginning of the calendar day. The reason for this is the same as in the case of the maximum temperature. Sunrise occurs between 8:30am to 10am during the winter months in Fairbanks. It can be seen that the daily low temperature does not tend to occur most frequently around those times, as it does further south.

We see that Alaskans cannot always expect a forecast to say "High this afternoon 10 below and low tonight 40 below", at least not during the winter months. Instead, they might hear "Temperatures rising to 10 below early this morning, then falling to 40 below by midnight".

ALASKA'S SEA ICE

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Viewing Arctic pack ice evoked this poignant description from one of the early Arctic explorers: "It was like being in some gigantic sculptor's yard; stacked as far as the eye could see were glistening marble blocks cut in a million fantastic shapes." Indeed, Arctic sea ice is far from being a white and featureless plain; from the time of its formation it is moved about, broken up and deformed by the action of winds, currents and tides, and the prediction of its motion is a difficult undertaking.

Sea ice forms every fall and winter in the vast seas surrounding Alaska's western and northern sides; the Bering, Chukchi and Beaufort Seas, respectively, and in Cook Inlet on the south central coast (Figure 1). The arrival of the pack ice and the animals it carries is an annual event which the Eskimos have historically depended on for survival during the long and severe winters. The success of the harvest is closely related to the coastal ice conditions from where most of the harvest is conducted. For instance, persistent offshore winds can keep the drifting pack ice away from shore, a condition more likely to occur on Alaska's west coast where the pack ice has room to move than on the Arctic coast where it is more compacted. Anomalous pack ice movements are thought to have been responsible for the decimation of St. Lawrence Island's Eskimo population during the 1878-79 winter. Southerly winds supposedly kept the pack ice and its resident walrus herds from reaching the island during the critical harvest season causing the starvation-deaths of 1,500 people. The island's population has never recovered from this.

The influence of Alaska's sea ice extends indirectly to all of its population alike, in shaping the climate and affecting local weather conditions

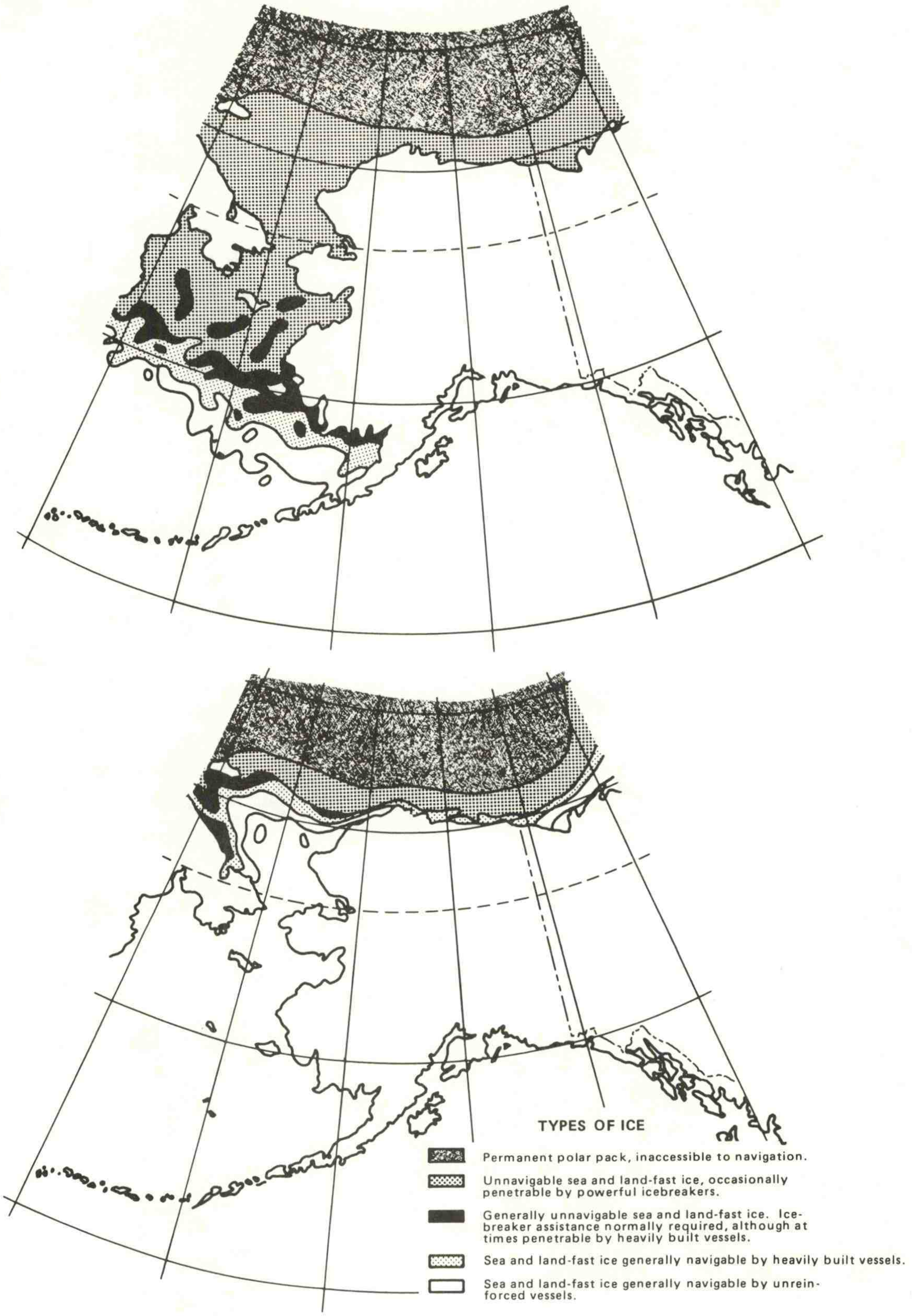


Figure 1 Average sea ice conditions in the Bering, Chukchi and Beaufort Seas in April (top) and August (bottom). (From the U.S. Navy Ice Atlas of the Northern Hemisphere.)

while directly determining the health and well-being of Inupiat subsistence hunters, the prosperity of fishermen and the success of oil companies drilling for offshore oil and gas.

Extent and Characteristics of Alaska's Sea Ice

Each of the seas surrounding Alaska has unique sea ice characteristics. On the southern shores of Alaska, sea ice forms only in Cook Inlet which is surrounded by cold land masses during the winter. With increasing latitude, sea ice thickness and aereal extent increases; typical thicknesses are: half a meter or less in Cook Inlet, 30 to 100 cm in the Bering Sea, 100 to 120 cm in the Chukchi Sea, 120 to 180 cm in the southern Beaufort Sea (annual or first-year ice), and over three meters in the Arctic Ocean (multi-year ice). However, imposed on this basic pattern, ice conditions are affected by sea floor and coastline topography, tides, ocean currents, storms and the presence of islands; all serving to deform the ice into a wide range of different ice thicknesses and concentrations ranging from open water polynyas (irregular but large openings in the ice cover) to ice pressure ridges tens of meters in depth.

In Cook Inlet, large tidal fluctuations freeze and pile up the ice in layers on the mudflats and beaches, creating thick and hard chunks much thicker than the ice formed *in situ* - some pieces have been reported to be as thick as three to four meters. These chunks are pushed into the inlet by strong winds, menacing ship traffic and, in the words of one observer, "literally jar the teeth" of the people who work on the oil platforms in Upper Cook Inlet (Schulz, 1977). Usually the ice stays in Cook Inlet, however strong northwesterly winds can sweep the ice out of the inlet and down the Shelikof Strait between Kodiak Island and the Alaskan Peninsula.

The ice cover over the Bering Sea is influenced by winds generated by storms moving northeastward from the Pacific Ocean, and by warm northward-setting ocean

currents (Dehn, 1971). These influences, along with the diverging Siberian and Alaskan coastlines, keep the ice cover along a wide ice front loose and relatively thin. In fact, the ice conditions over the entire southern portion of the Bering Sea are characterized by a mixture of ice cakes, floes and detached ice fields which usually stay north of the Pribilof Islands but can be pushed southward by strong northerly winds following the passage of storms. The ice cover thickens and compacts farther northward - away from the ameliorating influences of winds and currents. St. Lawrence Island obstructs ice movement southward through the Bering Straits, and results in a heavily deformed and thick ice field extending from the north side of the island to the Bering Straits. During the course of the winter, shorefast ice, so-called because it is attached to the coast, develops 10 to 15 km seaward, being more extensive in Norton Sound and on the northern side of the islands. During the summer, the Bering Sea is ice-free.

Chukchi Sea ice is quite different in character from Bering Sea ice. Where the ice cover in the Bering Sea is virtually unrestrained, that of the Chukchi Sea is contained by the converging Siberian and Alaskan coastlines and the Bering Straits. Moreover, the annually expanding polar ice pack presses the Chukchi Sea ice toward the narrowing coastlines, causing the ice to buckle and become heavily deformed. Shorefast ice forms in Kotzebue Sound and in a narrow band along the coast. The drifting pack ice grinds against the edge of the shorefast ice, piling the ice haphazardly into walls or ridges the heights of which are usually two to three meters, but are sometimes much higher; a nine meter high pressure ridge grounded in 20 meters of water was observed in the Bering Straits by Coast Guard personnel in 1970 (Dehn, 1971).

The Beaufort Sea, unlike the other seas around Alaska, is never completely ice-free, due to the proximity of the multi-year polar pack ice. Polar ice is

pushed toward Alaska by a clockwise-circulating current, called the Beaufort Gyre, lying between Alaska, the Canadian Archipelago and the North Pole (Sater et al., 1971). In fall and winter (beginning in the latter half of September), ice forms along the Beaufort Sea coast and spreads northward while the polar ice pack expands southward. Initially, polar ice mixes with the outer areas of the loose and thin shore ice (the shore ice tending to impede the southward progression of the polar ice floes). However, the shorefast ice stabilizes with increasing thickness and the moving pack ice shears against its seaward edge, deforming the ice into ridges (Figure 2). The ridges ground in shallow water, effectively anchoring the coastal ice sheet (Barnes et al., 1978) (Figure 3). The ridging is most intensive off the eastern Arctic coast where the ice is shoved shoreward by the Beaufort Gyre, and decreases toward the west as the drift becomes more parallel to the coast (Barnes et al., 1978).

Sea Ice Hazards

Although the seasonal advance of the pack ice may be a propitious occurrence for Alaska's coastal inhabitants, it is not as esteemed by shippers or fishermen. Marine navigation along Alaska's western and northern coasts can occur only during the summer (or melt) seasons which grow progressively shorter from south to north. Vessels can navigate freely in the Bering Sea by July 1, over most of the Chukchi Sea by late August, and across the nearshore waters of the southern Beaufort Sea during a six-to-eight-week period in August and September. Cook Inlet ice, because of its thinness, can be negotiated by shippers during the ice season (November to March) but not without occasional difficulty.

The Bering Sea and the northern Gulf of Alaska are noted for their rich fishing grounds. During unusually cold winters, Bering Sea pack ice encroaches on the fertile crab grounds in the vicinity of the Pribilof Islands and south

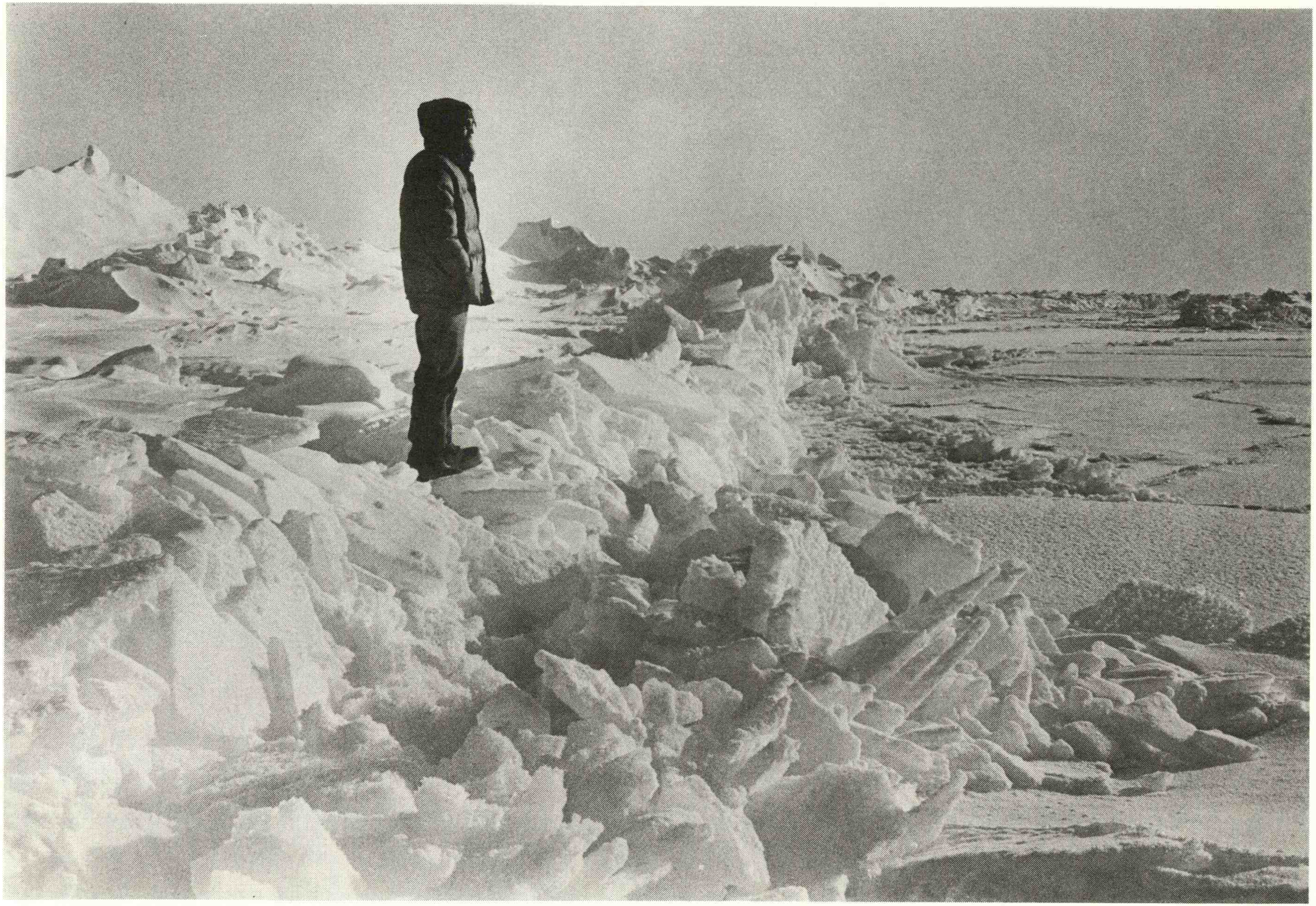


Figure 2 Typical ice pressure ridge formation in the nearshore areas of the Beaufort Sea.

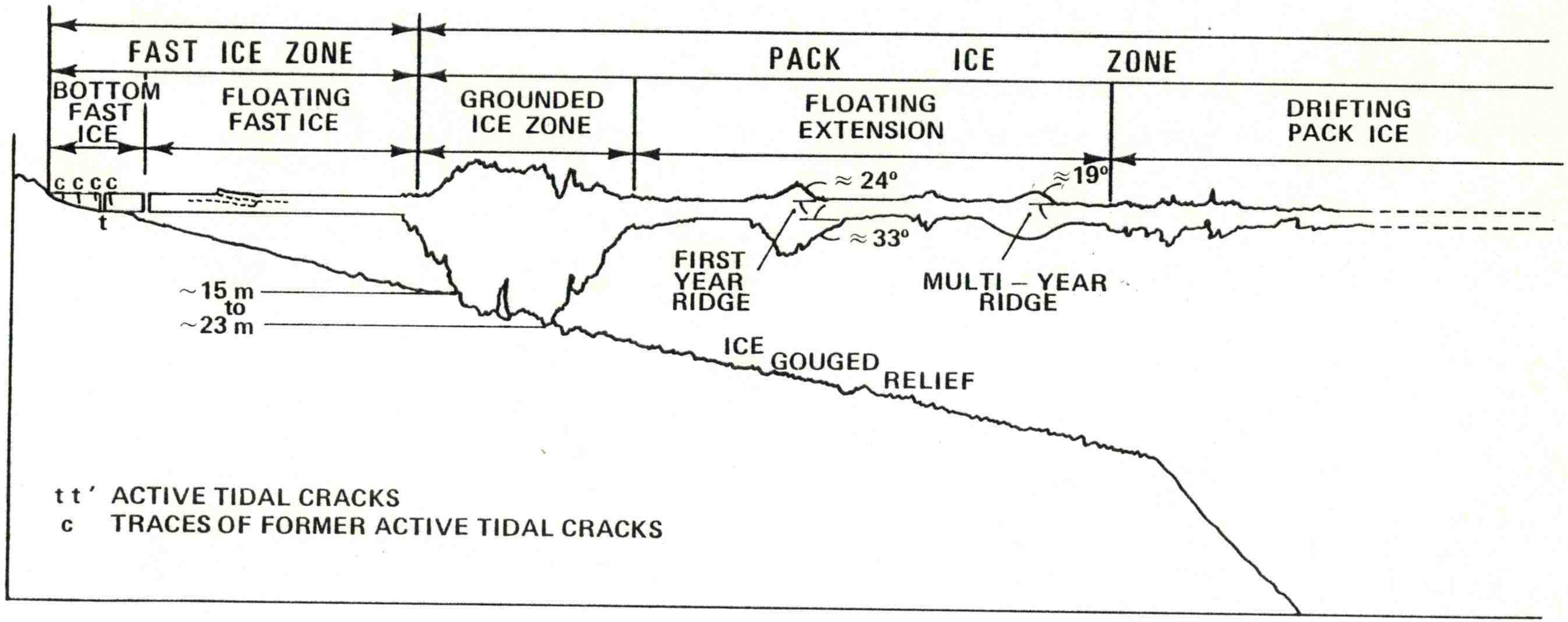


Figure 3 Late winter ice zonation of the Alaska Beaufort Sea Coast (modified from Kovacs, unpublished).

of Bristol Bay, destroying valuable fishing equipment (approximately 70,000 crab pots are spread over these areas valued at $3.5 million). Severe winters also expand Cook Inlet ice farther southward than usual and, with the onset of strong northerly winds, pieces of ice can drift into the Shelikof Straits where crab fisheries are located. Furthermore, the Bering Sea fisheries are expanding to include bottomfish (pollock and sablefish, for example) and herring (Pennington, 1978). Sea ice poses an increasing problem to commercial fisheries.

The excellent prospects that Alaska's continental shelves harbor new sources of petroleum add a new dimension to the sea ice problem. Most of the areas under consideration for oil exploration are, for at least part of the year, covered by a veneer of sea ice. Sea ice complicates the technology of oil drilling in marine areas since the oil platforms must be designed to withstand the forces of moving pack ice while remaining stable (Coon et al., 1977). The presence of sea ice also complicates the manner in which petroleum products can be shipped to market. Supertankers with ice-breaking capabilities are a possible solution (Sater et al., 1971). Furthermore, the technology that works in one area may not suffice in another because of dissimilar sea ice conditions.

Ice Surveillance and Forecasting

The deleterious effects of sea ice on Alaskan marine operations emphasizes the need for ice reconnaissance and forecasting. Such a program was initially formulated by the U.S. Navy in the mid-50's to benefit ships ferrying cargo to the Arctic for the construction of defense installations (Dehn, 1974). The successful yearly resupply of these sites (except in 1975, when persistent northwesterly winds kept the pack ice close to Alaska's north coast) attests to the Navy's expertise in this area. At the present, both government agencies and private consulting firms are engaged in ice prediction. Their primary tool for analyzing sea ice conditions on a year-round basis is the weather satellite.

Satellite data is corroborated with ship and shore ice reports and, during the critical resupply of Alaska's North Slope, with aerial reconnaissance ice information. Armed with the knowledge of existing ice conditions, the ice forecasters undertake the job of predicting ice movement, changes in concentration, development of navigable leads (and their closing), etc., utilizing empirical rules and formulas and computer-generated forecasts. Planning shipping operations is facilitated by the promulgation of seasonal outlooks which make general predictions on when the shipping corridor along the Beaufort Sea coast is expected to open, and on the types of ice that might be encountered (polar ice, floebergs, etc.).

As nature would have it, clouds and fog can prevail across the Arctic coast during the shipping season for several days, and over the southern Bering Sea in winter for weeks, precluding sea ice observation from both satellite and aircraft. Critical decisions regarding tug and barge movement or the evacuation of fishing equipment are hampered. The technology exists to overcome this problem, but unfortunately it is not yet in widespread use. Both satellite and aircraft can be equipped with radar systems which have the ability to penetrate clouds, fog and all but the heaviest precipitation, and detect sea ice, day or night. Aircraft mounted with such systems (Side Looking Airborne Radar) are particularly useful in monitoring ice conditions in support of shipping since the ice information (radar signals) can be transmitted simultaneously to the ships and to ground stations and can thus be used without delay (Schertler et al., 1977). Presently, the only operating satellite capable of "seeing" ice through clouds is the Nimbus V weather satellite, but it detects only heavy ice concentrations (50 percent or more) and is, therefore, of only limited use in detailed ice analysis.

Forecasting the behavior of sea ice is also subject to limitations because of a lack of comprehensive knowledge of the environment in which sea ice forms and moves about. For instance, it is not unusual for sea ice to form over broad expanses of the ocean in short periods of time. How does one forecast such an occurrence? It is a matter of intensive scientific inquiry: several cases have to be studied to understand the process, environmental parameters (air temperature, water temperature, wind speed, etc.) have to be known to establish the initial conditions, and, most importantly, changes in those parameters have to be predicted. At present, not enough environmental data can be routinely collected to do a good job at ice forecasting. Since the wind forces the ice to move, accurate weather analyses are essential to forecasting weather changes, hence, ice conditions. Except for a few Soviet drifting ice stations and U.S. and Canadian ice buoys, there are no routine reports of wind, pressure and temperature from the Arctic Ocean; however, it is hoped that sooner or later this will be rectified.

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CLOUDS AND PRECIPITATION IN ALASKA

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Alaska with its vast geographic extent covers many climatic regions. Cloud and precipitation features, accordingly, vary from zone to zone. Because of Alaska's high latitude, clouds and precipitation throughout have some singular aspects, as described below. While the interior of Alaska has strongly contrasting summer and winter climates (i.e. temperature range, cloud form and mode of precipitation), the contrast in the other zones is less severe due to the oceans, which act as a heat reservoir.

Cloud climatology

The commonest summer clouds in the Arctic are extensive covers of stratus and stratocumulus (Figure 1). Their formation mechanisms need further study, but the clouds are probably formed by the cooling and condensation of water vapor emitted from the open water and wet pack ice of the Arctic Ocean in the presence of steep temperature lapse conditions over the melting ice. The turbulence in the clouds is rather minor, but the clouds are frequently supercooled and pose an icing hazard to aircraft. The cloud base frequently lowers to the ground, hampering airport activity. The clouds cover not only the arctic coastal zone but the entire Arctic Ocean and occur about 74% of the time from May to August. Due to this prolonged extensive cloud cover, solar energy to warm the ground is substantially reduced and consequently the surface temperature remains low. The arctic summer climate is thus heavily affected by the stratus clouds. In the winter, cloud climatology in the Arctic is similar to that of the Interior.

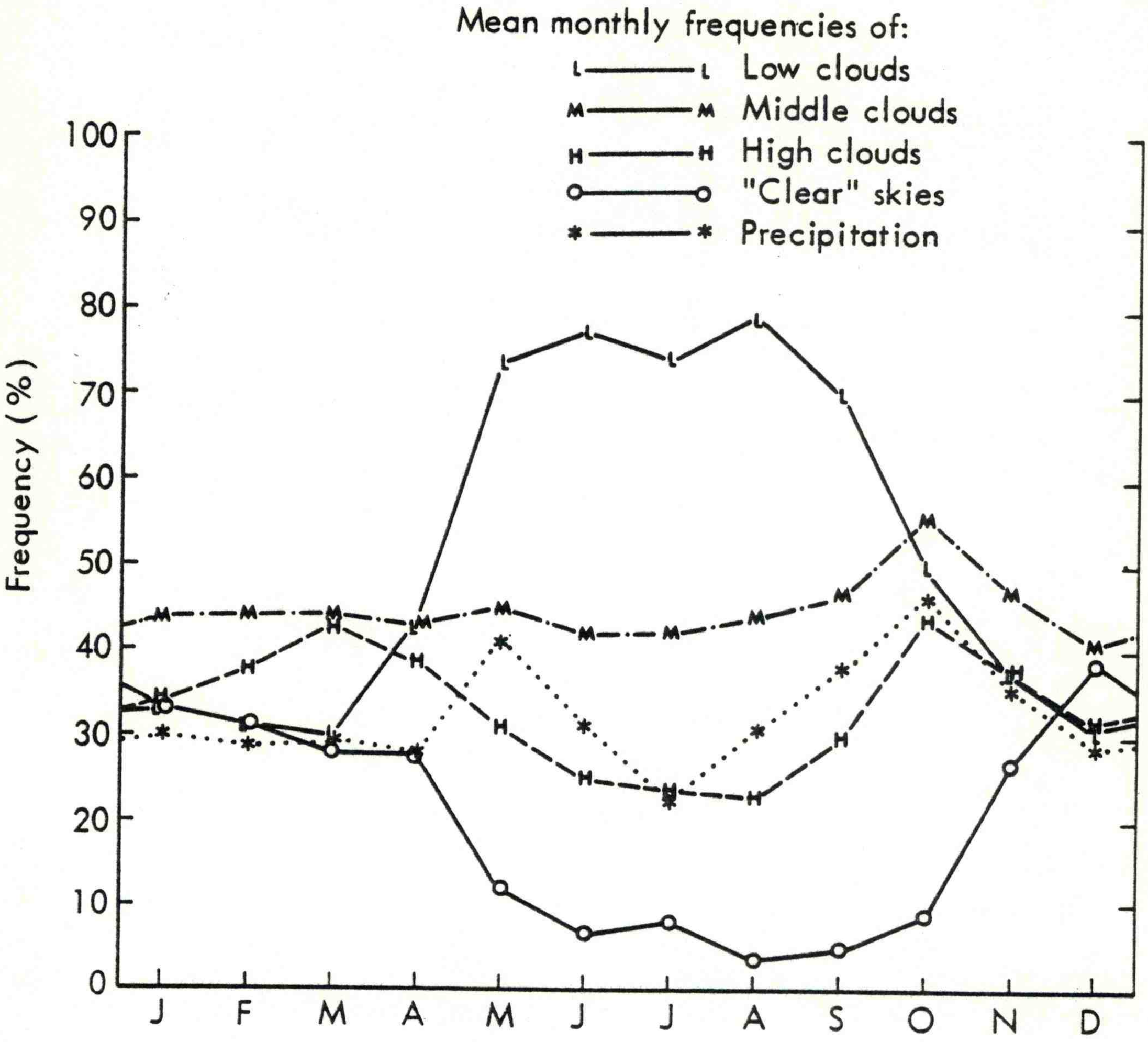


Figure 1: Frequency of cloud conditions over the Arctic Ocean. L-L indicates stratus clouds. (After Huschke, 1969).

Summer clouds in the interior region are similar to those in continental regions elsewhere. Both frontal cloud systems and cumuliform clouds are common. Frontal systems are most often of the occluded type. Cumulus clouds are associated with relatively gentle convection, and rarely reach the heights observed in the more southern states. Since the convection leading to cumuliform clouds is initiated by solar heating, these clouds normally disappear from early evening to morning.

Thunderstorms and lightning are mainly observed in the interior of Alaska during the summer season. Since convective activity in cumuliform clouds is moderate, as previously described, electrical activity is not violent. The initiation of convective activity is definitely related to the solar heating of the ground. The development of cumulus clouds into thunderstorms depends on the availability of moisture at levels higher than 1500 m. This moisture is transported by advection via large scale weather systems. Although lightning and thunder are rather weak, most forest fires in Alaska are caused by cloud-ground lightning strokes. Many of these fires in tundra burn continuously even under the snow cover throughout winter. Satellite photographs in both visible and infrared wave bands have been used for identification of thunderstorm and forest fire locations. Cloud top temperatures lower than -28°C derived from the satellite imagery show that the clouds have grown to great vertical heights; their locations coincide well with ground observations of thunderstorms (Biswas and Jayaweera, 1976).

Thunderstorms in the arctic zone and over the Arctic Ocean are not impossible. Although some were reported as early as 1580, they seem to occur only once in a few years. One of the most remarkable thunderstorms was observed at the AIDJEX camp (76° 25'N 147° 7'W) on July 7, 1975.

Several occurrences of lightning, thunder and rain gushes (total 0.65 mm precipitation) were observed at surface temperatures between +0.2° - +0.9°C (Hanson, 1979).

As the nights become longer, and nocturnal radiative cooling becomes more important, unfrozen steam fog can often be seen near swamps and lakes as moist air cools radiatively.

Winter clouds in the interior are predominantly stratiform, and include most cloud types typical of occluded fronts. Classical warm and cold fronts are rarely observed. Ice crystal cirrus clouds may at times be observed at much lower elevations than in mid-latitudes. Probably the most striking "winter cloud" is ice fog, a type of air pollution with water vapor as the pollutant. At sufficiently low temperatures water added by human activity condenses and is quickly frozen into tiny ice crysatls of irregular shape. Typical ice fog crystals and their size distribution are shown in Figures 2-4, respectively. As the temperature goes down, more and smaller ice crystals are produced, particularly spherical crystals at temperatures below -37°C. Below this temperature ice crystals tend to be formed without nuclei (Ohtake, 1970). Major water vapor sources for these crystals are the combustion of fuels in heating and power plants, evaporation from cooling ponds or open water from power plants and from automobile exhaust (Benson 1970). Ice fog is much like the condensation trails made by aircraft flying at high altitudes, and in Fairbanks it may last for a few days to a few weeks. During days with ice fog in December and January the air temperature does not change much even in the daytime, because of the very low solar angles in mid winter (see Augulis' article in this report). Clear skies occur about 1/3 of the time in winter.

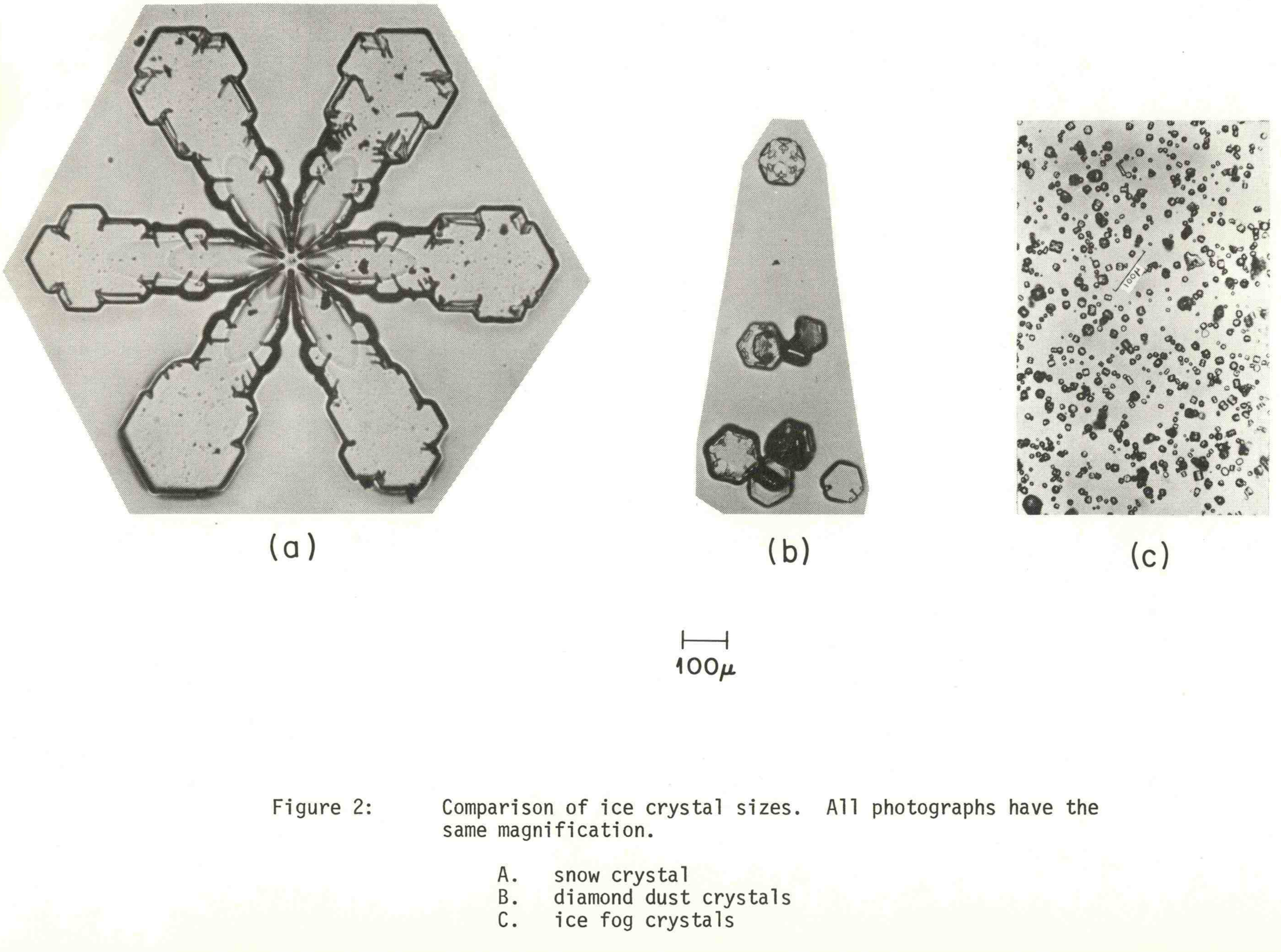


Figure 2: Comparison of ice crystal sizes. All photographs have the same magnification.

A. snow crystal
B. diamond dust crystals
C. ice fog crystals

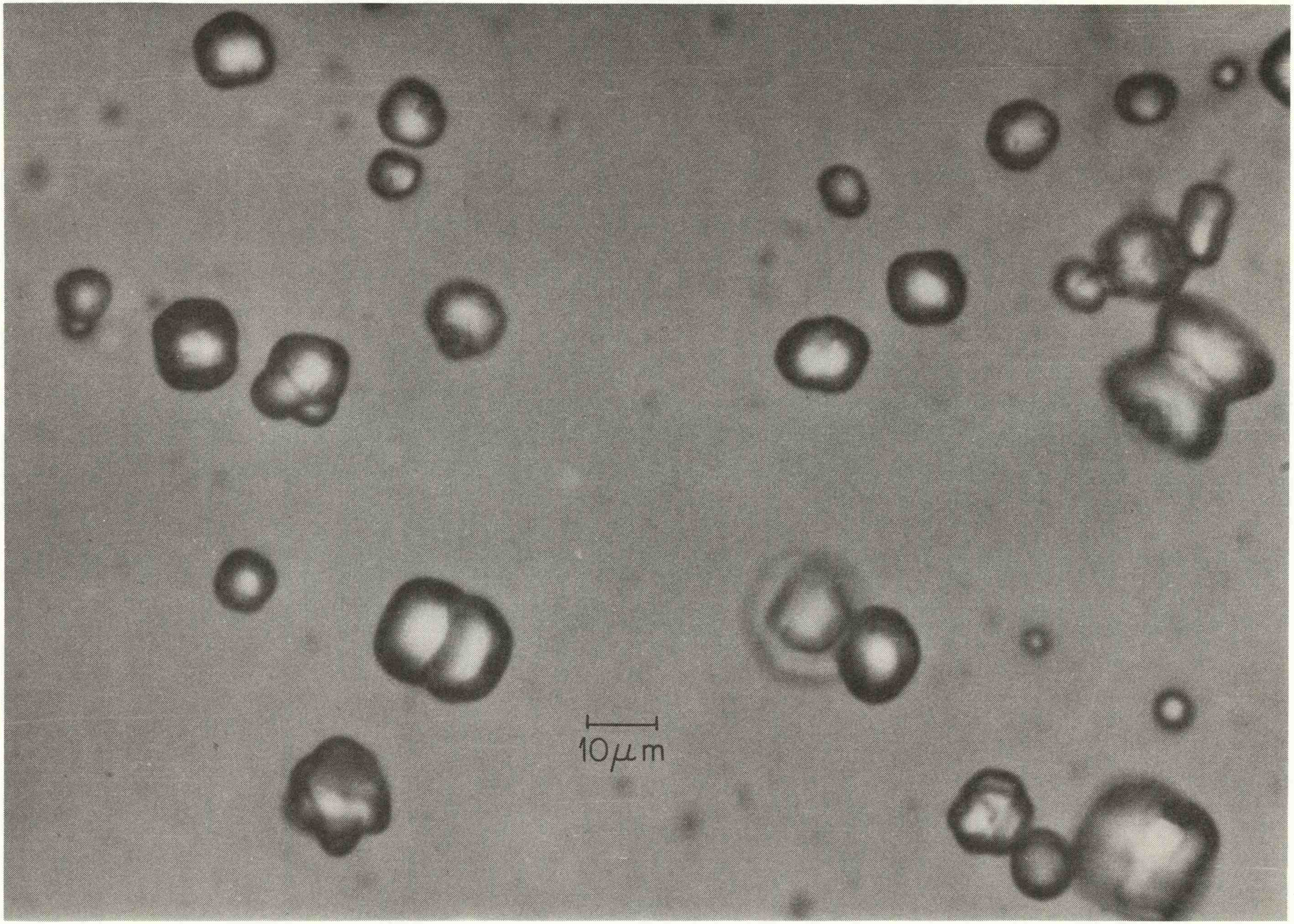


Figure 3: Typical ice fog crystals at temperatures below -37°C.

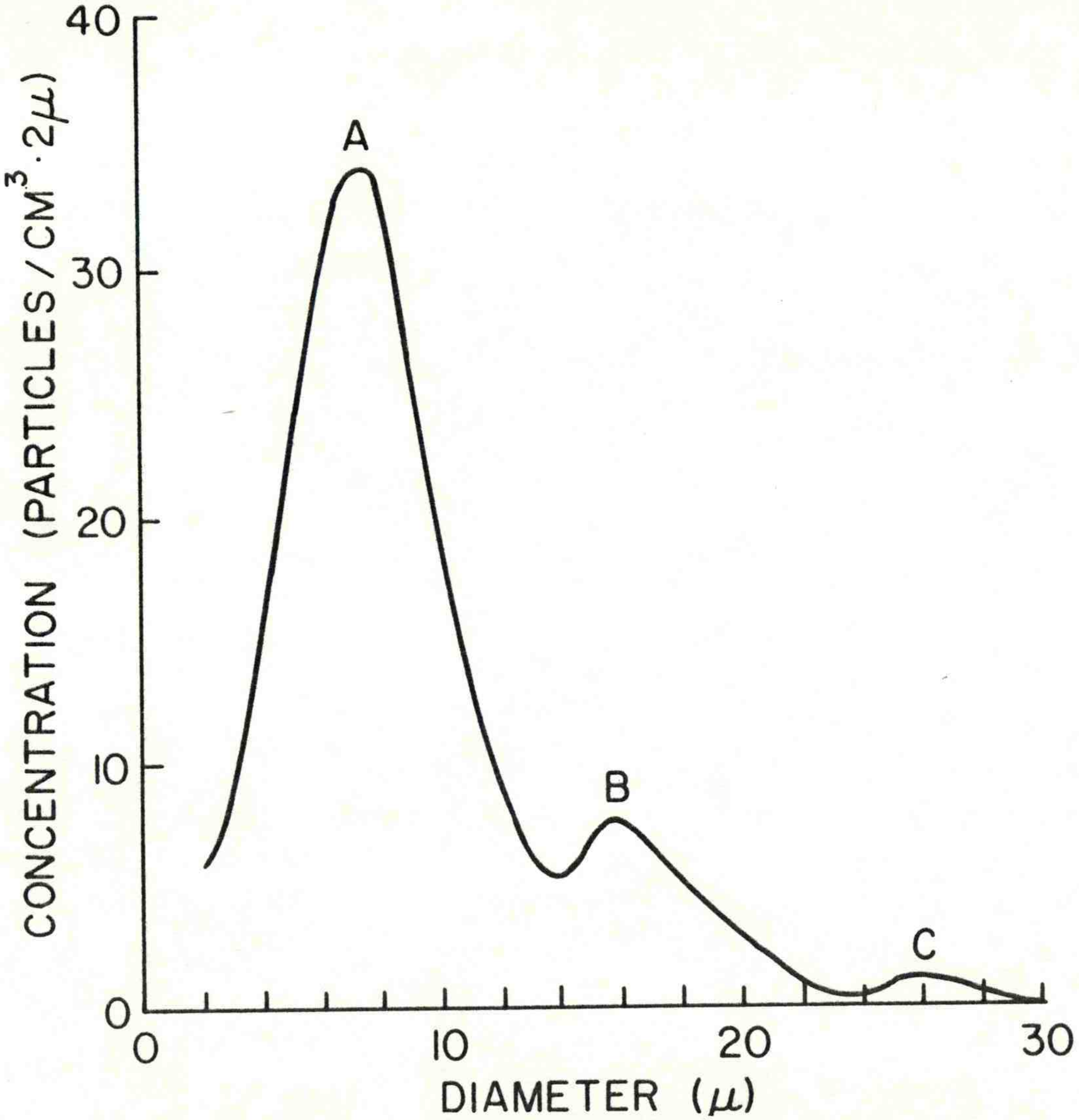


Figure 4: Typical size distribution of ice-fog crystals in downtown Fairbanks, at -36.5°C. Total water content 0.23 mg l^{-1}, solid water content 0.12 mg l^{-1}, visibility 180 m, particle concentration 153 p cm^{-3}, at 1240 AST on 8 December 1968. Peaks A, B and C represent crystals from car exhausts, open water and heating plants (commercial and residential) respectively. Heights of the peaks are different from place to place owing to different moisture and temperature conditions.

Air traffic within Alaska is predominately at fairly low altitudes, but contrails are common and at times spectacular along the polar route to Europe (Figure 5). The frequency of spreading contrails attest to the frequent presence of air supersaturated with respect to ice.

The maritime zone, which includes southeastern Alaska, the south central coast and the Aleutian Islands, has warm foggy winters and cool wet summers, and its cloud conditions are rather similar to those of the Pacific Northwest. In general, most clouds over Alaska are associated with cyclones or low pressure zones, as confirmed by satellite photographs.

Precipitation Climatology

Most of the fall, winter and spring precipitation in the Arctic is observed as snow, but a small amount is in the form of ice crystals, often called "diamond dust", observed under cloudless skies. The moisture to form the ice crystals originates from the open waters of the Arctic pack ice (as do the arctic stratus clouds) and from man-made sources such as heating plants and vehicle exhaust. The water droplets from those sources freeze into ice crystals which grow by vapor diffusion, and the ice crystals precipitate from a clear sky after drifting a long distance (Ohtake et al., 1978).

Another interesting phenomenon of precipitation in Alaska is warm drizzle. This occurs mostly in maritime tropical regions. However, in spite of the cold climate, warm rain is observed throughout Alaska, including the Arctic, and especially in the maritime zone (Ohtake, 1963). Even in the interior supercooled warm drizzle, with associated icing, can be observed once in a while.

"Warm rain" is rain which has not passed through the frozen state during the formation of precipitation particles from cloud droplets.



Figure 5: Expanding aircraft condensation trail over McCall Glacier, Brooks Range. This location is under the transpolar flight route.

It thus differs from raindrops originating from melted snowflakes or any other frozen particles. Precipitation can be identified as definitely warm if the temperature between cloud top height and the ground is everywhere above freezing, and as probably warm if no part of the cloud has temperatures below -5°C. Supercooled warm drizzle can occur but can only unequivocally be identified if liquid precipitation is observed on the ground while no portion of the atomsphere below the cloud tops is above freezing - i.e., precipitation particles have had no opportunity to melt and therefore liquid particles must have been liquid all along.

In the interior of Alaska diamond dust is also observed at temperatures lower than about -13°C when the atmospheric water content is above ice saturation. The ice crystals composing the diamond dust result from the freezing and subsequent vapor-phase growth of water droplets which in many cases originate as condensed water vapor from power plants (Ohtake and Jayaweera, 1972). Diamond dust occasionally displays beautiful halos and sun pillars (Figure 6). These optical phenomena are caused by refraction and reflection of sunlight or street lights by plate-type ice crystals. Ice fog, on the other hand, does not display such halos or pillars because its crystals have near-spherical shapes. Both diamond dust and ice fog usually occur when winds are weak and variable.

In Alaska, especially in the interior and the arctic zone, snow is often observed to fall from rather thin stratiform clouds with very minor convective activity. The growth rate of the snow flakes seems to be higher than ever reported elsewhere but this is a problem that needs further attention.

The precipitation climatology in the maritime zone is similar to that of the Pacific Northwest.



Figure 6: Sun pillar caused by diamond dust.

Weather Modification in Alaska

Due to the different nature of Alaskan weather conditions, weather modification interests in Alaska have also been somewhat different from those in other areas. These interests include thunderstorm suppression, production of artificial clouds, cold fog seeding and dissipation of arctic stratus and ice fog.

A) Thunderstorm suppression

Cloud-to-ground lightning strokes are causing most of the forest wildfires in Alaska. In the late sixties and early seventies the Bureau of Land Management conducted summer cloud seeding programs to prevent and control forest fires by reducing the number and intensity of lightning strokes. Three aircraft dispensed silver iodide into the tops of potential thunderstorm clouds. A subjective evaluation of the success of seeding seemed to indicate that the majority of the seeded clouds glaciated well. Glaciation reflects the conversion of supercooled water droplets to ice particles, thereby reducing the electrical charging and separation mechanism, and thus lightning (Franks, 1975).

B) Artificial cloud making to prevent radiative cooling

Low surface temperatures on the ground normally result from radiative cooling under clear skies in the interior of Alaska. At times, these clear skies occur even though the air is below freezing temperature and supersaturated with respect to ice. Under these circumstances, an artificial ice cloud can be formed by introducing pellets of dry ice into the air. Such clouds could be made to reduce cooling at the ground, since the clouds radiate heat back to the ground until a radiative equilibrium between ground and air is established. Artificial creation

of clouds was successfully achieved in Fairbanks and in Barrow (Jayaweera and Ohtake, 1972).

C) Cold fog dispersal

Cold fog is composed of suspended liquid water droplets at temperatures below 0°C near the ground. Airport activities in the maritime zone of Alaska are occasionally hampered by the cold fog, since it restricts visibility. An experiment in cold fog dissipation at Elmendorf Air Force Base, near Anchorage, was initiated in the 1965-66 winter season. Although the early experiments utilized dry ice seeding from tethered free balloons or from an aircraft, liquid propane released at ground level was later found to be effective in clearing cold fogs and this technique has since become operational (Wise, 1973). Some supercooled fog droplets are converted to ice crystals upon sudden cooling caused by the expansion of the propane during the spraying process. As these ice crystals grow by consumption of water vapor from the remaining water droplets and fall out to the ground, visibility is improved. The amount of propane used in seeding is small enough so that there is no danger of fire even downwind of the propane release points.

D) Artificial dissipation of stratus clouds over the arctic zone

During the summer months, supercooled stratus clouds cover the arctic coastal area and the Arctic Ocean. The stratus clouds considerably reduce the amount of sky and solar radiation received at the ground. These clouds also interfere with many human activities. The most important of these is air transportation, on which almost every other activity of man in these regions depends and which provides a lifeline to some villages. Removing or dissipating such clouds by seeding would allow solar energy to reach the ground or sea ice, thus raising the temperature

and clearing additional fog or clouds. Seeding experiments to achieve this started in the fall of 1978. A small aircraft was used to disperse liquid carbon dioxide, which produced a high concentration of ice crystals within supercooled stratus clouds in the same manner as does liquid propane seeding. These experiments were successful in at least temporarily dissipating the clouds.

E) Ice fog control

Because ice fog is thermodynamically stable, conventional (and unconventional) modification techniques have not succeeded in reducing the ice fog problem. Some success has been attained in reducing the fog density by controlling water sources--specifically, by reducing evaporation from cooling ponds (McFadden, 1978). Much remains to be done in this field, however.

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ALASKAN SNOW

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Alaska contains the most magnificent array of mountain glaciers outside of the polar regions. Some of Alaska's glaciers are huge, the Bering Glacier covers an area of 5800 km^2 by itself; that's 700 km^2 more than the total area of Delaware. The total area covered by glacier ice in Alaska is 73,000 km^2; ten states have sizes smaller than that. Although Alaska is justly famous for its glaciers they cover less than 5% of its area of 1.5 million km^2, all of which is covered by seasonal snow for part of the year. This article briefly summarizes some highlights of the seasonal snow cover of Alaska which varies so greatly in quantity, length of stay and physical properties in different parts of the State.

Snow forms a thin veneer over most of Alaska's surface area of 1.5 million km^2 for one-half to three-fourths of the year. Alaska has three different types of snow cover corresponding to its three climatic regions. The arctic polar climate is separated by the Brooks Range from an extreme continental climate, which yields, by way of a transitional climate zone south of the Alaska Range, to a maritime climate. The snow varies from a dry, hard, wind-packed and shallow "tundra snow" on the Arctic Slope to the low-density, dry, largely recrystallized "taiga snow" of the interior, and finally to the thick and often wet "maritime snow" of the southern coastal regions. The terms "Tundra Snow" and Taiga Snow" were introduced by Pruitt and have proved to be useful, for completeness the term "Maritime Snow" is added.

Alaskan snow differs from the hydrologically important mountainous snow of the western United States in that its temperatures are lower, steeper temperature gradients occur in it, and there is less of it per unit area; however, it lasts longer and enters more directly into human activity as snow itself rather than serving primarily as a cold storage water reservoir. The physical properties of this snow and the physical processes which occur in, above, and below it are important and fascinating. All forms of life in Alaska must adapt to the presence of snow and the complex heat and mass transfer processes that occur within it. Some animal species, such as lemmings, spend most of their lives beneath the snow.

The Tundra Snow of the Arctic Slope

Aside from perennial snow in mountainous areas and on glaciers, the snow cover in Alaska lasts longest on the north slope of the Brooks Range. For three quarters of each year, the entire Arctic Slope, from the foothills across the tundra to the Arctic Ocean, is covered with dry, wind-packed snow. This tundra snow forms a wind-swept rough surface which strongly resembles the year-round appearance of the Greenland and Antarctic ice sheets, or the winter snow surface of the adjacent Arctic Ocean. It commonly has wind-sculptured, hard packed snow features which are referred to by their Russian name, "sastrugi" (Figure 1).

The similarity between tundra snow and snow on the polar ice sheets does not stop at the surface. Indeed, the structure of the tundra snowpack resembles the top annual stratigraphic unit of the Greenland or Antarctic Ice Sheets; it consists of a hard, high-density, wind-packed layer, overlying a coarse, low-density, depth hoar layer consisting of big loosely-bonded crystals (Figure 2). Although there is considerable

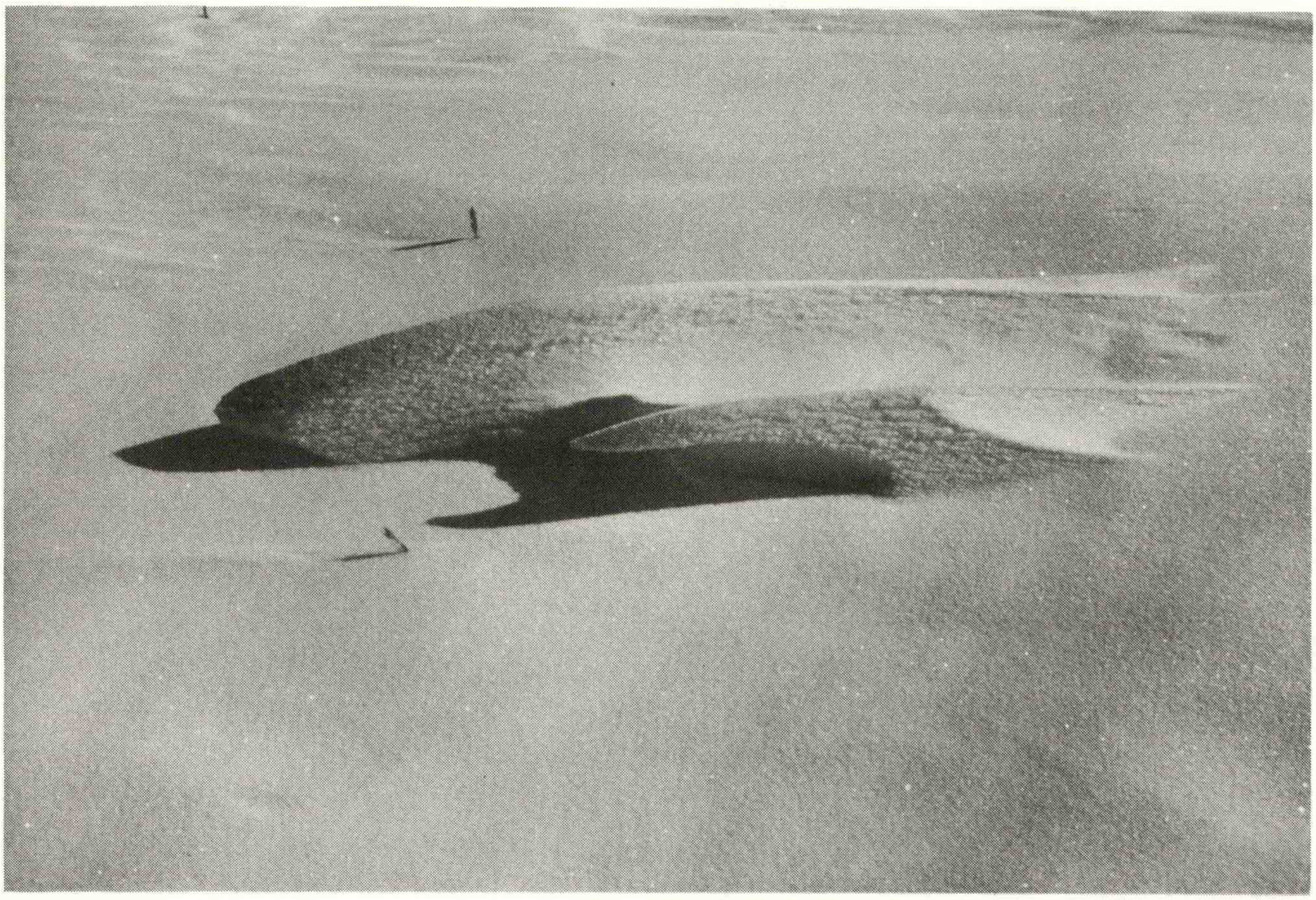


Figure 1. Sastrugi are sharp, irregular ridges formed on the snow surface by wind erosion and deposition. The ridges are parallel to the direction of prevailing wind which was from the northeast (left to right) in this photo. The feature shown here is about 1.5 m long, and bits of vegetation are seen protruding through the snow surface.

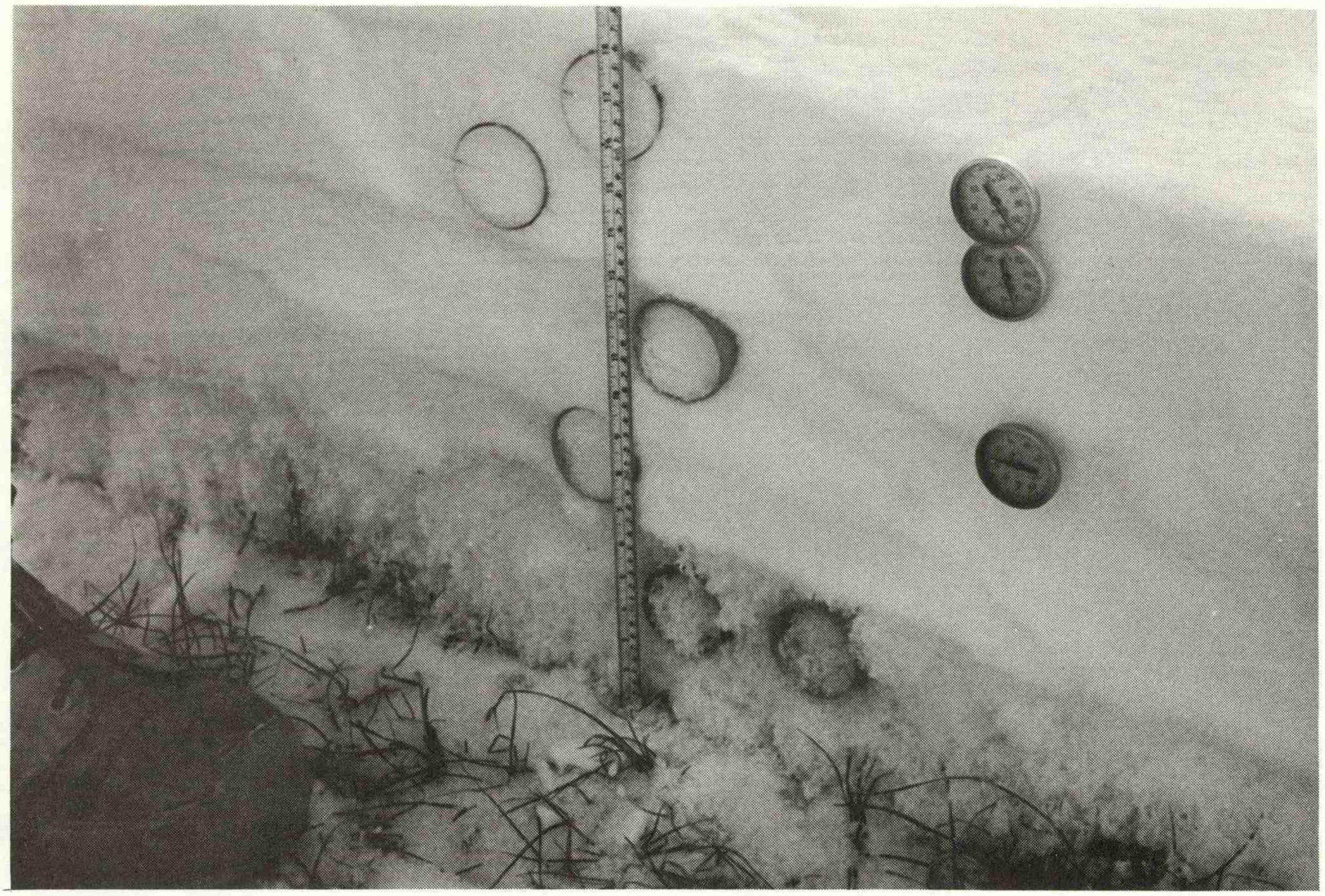


Figure 2. A typical cross section of a 47 cm thick tundra snow pack with sample tubes and thermometers inserted. The two bottom tubes are in the basal depth-hoar layer (average density = 0.19 g cm^{-3}); the middle tubes are in fine to medium grained snow, (average density = 0.27 g cm^{-3}) and the top two tubes are in a hard winter-packed layer (average density = .36 g cm^{-3}).

variability in the stratigraphy of this snow, one can generally describe it by referring to only four major varieties of snow. In approximate order from top to bottom in the snowpack these are:

Snow Type	Range of Grain Size (mm)	Range of Density ($g\ cm^{-3}$)*
1. Fresh new snow, variable crystal forms	0.5 to 1.0 sometimes < 0.5	0.15 to 0.20
2. Wind slab, hard, fine-grained	0.5 to 1.0	.35 to .45
3. Medium grained snow	1 to 2	.20 to .35
4. Depth hoar, coarse loosely-bonded crystals	5 to 10	0.15 to 0.30

*The density ranges are only approximate. However, they indicate the differences one may expect between the various layers.

The total quantity of snow on the Arctic Slope is about 10 cm of water equivalent per year but is difficult to measure accurately because of the way it is blown about by the wind. In general, precipitation gages err on the side of underestimating the amount of precipitation. In order to improve precipitation measurements, experiments with wind shielding of the gages have been carried out for many years and the best and most elaborate system appears to be the Wyoming Blow-Fence developed by Rechard and Larson (1971). Use of this shielding increases the catch to about 100%. Several Wyoming gages were modified for arctic tundra and installed on the Arctic Slope (Figure 3). Comparisons showed that the Wyoming gages catch 3 to 5 times more precipitation on the Arctic Slope than do the unshielded Weather Bureau gages.

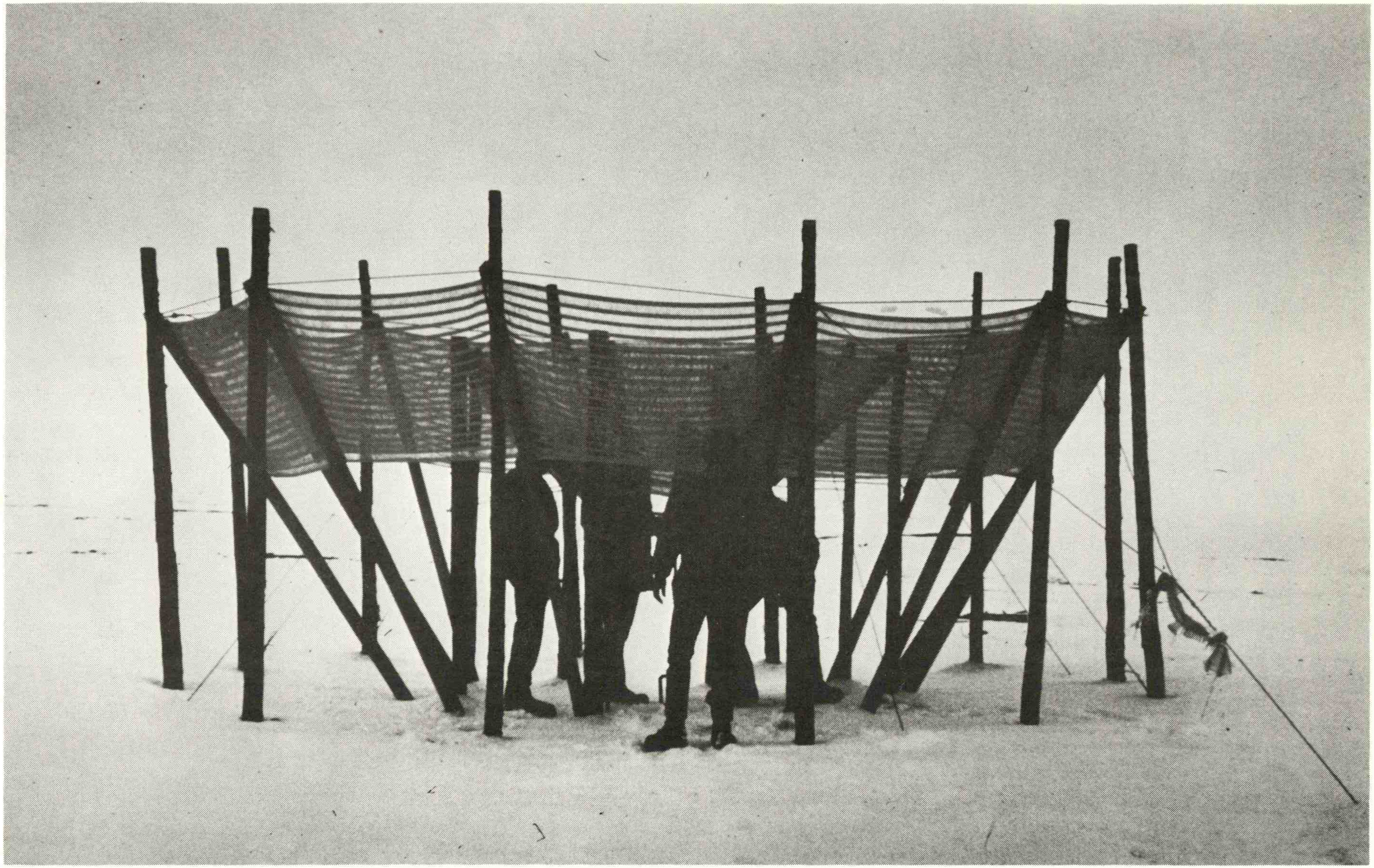


Figure 3. A precipitation gage surrounded by a Wyoming blow-fence. This design on wind shield, developed and tested in Wyoming, has proven itself on the Arctic Slope. It consists of two snow fences concentric about the storage gage and catched 3 to 5 times more snow than an unshielded gage.

The fact that annual precipitation has been underestimated at Barrow and Barter Island, by about 35%, has been known since Black compared the records with snow on the tundra in the early 1950's (Black, 1954). Since 1961, the U.S. Soil Conservation Service has been conducting an ambitious snow survey program in Alaska. Compared with similar research in the "lower 48" states, it is still in its infancy but growing fast, from ten snow courses in 1961 to 124 snow courses now. The sites on the Arctic Slope consist of eight Wyoming snow gages which were installed during the past three years.

The virtually unbroken tundra snow surface has a high reflectivity, or albedo, of about 80% for incoming solar radiation. When the snow disappears, in late May or early June, it goes rapidly and the albedo drops to about 15%. This five-fold decrease in albedo occurs at the same time that the amount of incident radiation is rapidly increasing, and results in a dramatic change in the amount of absorbed radiation at the surface (Weller, et al., 1972; Benson, et al., 1975). Similar observations were made on the Canadian tundra by McFadden and Ragotski (1967) who estimated a six-fold increase in absorbed radiation at the surface when the snow melted away. With the disappearance of the snow cover, air temperatures can therefore increase, initialing the beginning of summer.

The Taiga Snow of Interior Alaska

The U.S. Weather Bureau has maintained an office in Fairbanks since 1929. It is one of the few sources of long-term climatological data in interior Alaska and shows an annual average precipitation of about 30 cm water equivalent, of which 1/3 is in the form of snow which covers the ground for

more than half of each year. Departures from this average during some years are significant, however. For example, the snow pack in April 1971 was the largest on record but during April of 1970 it was 5 times less and nearly at an all-time minimum. The history of a typical taiga snow pack is shown in Figure 4.

The land south of the Brooks Range is forested and the forest cover prevents taiga snow from exercising the same dominant control on the albedo as does tundra snow. The boreal forest has about half the albedo of clear snow-covered areas within it, such as lakes and bogs (McFadden and Ragotzkie, 1967). In general, South of the Brooks Range the albedo is lower and more variable than it is on the North Slope. Nevertheless, spring break-up is abrupt, as can be seen in Figure 4, and can cause serious flooding, as discussed in the next article.

The winter climate of the interior lowlands of Alaska is typified by cold air and calm winds which strongly influence the physical characteristics of the snow pack. Despite very low temperatures at the upper surface of the snow the ground beneath it does not experience low temperatures, and strong temperature gradients therefore prevail in the snow. A comparison between the top 50 cm of snow on the Greenland ice sheet and that of the Fairbanks area will illustrate this. Assume that both have surface temperatures of -45°C. If we go 50 cm below the snow surface in the Greenland case, we find a temperature of about -40°C, while at the same depth in the Fairbanks case we would encounter the soil-snow interface at temperatures not lower than -5 to -10°C. The much higher temperatures at the bottom of the snow pack in interior Alaska are due to the effect of latent heat released in the soil by soil water as it freezes combined with the insulating value of the overlying snow. The large temperature

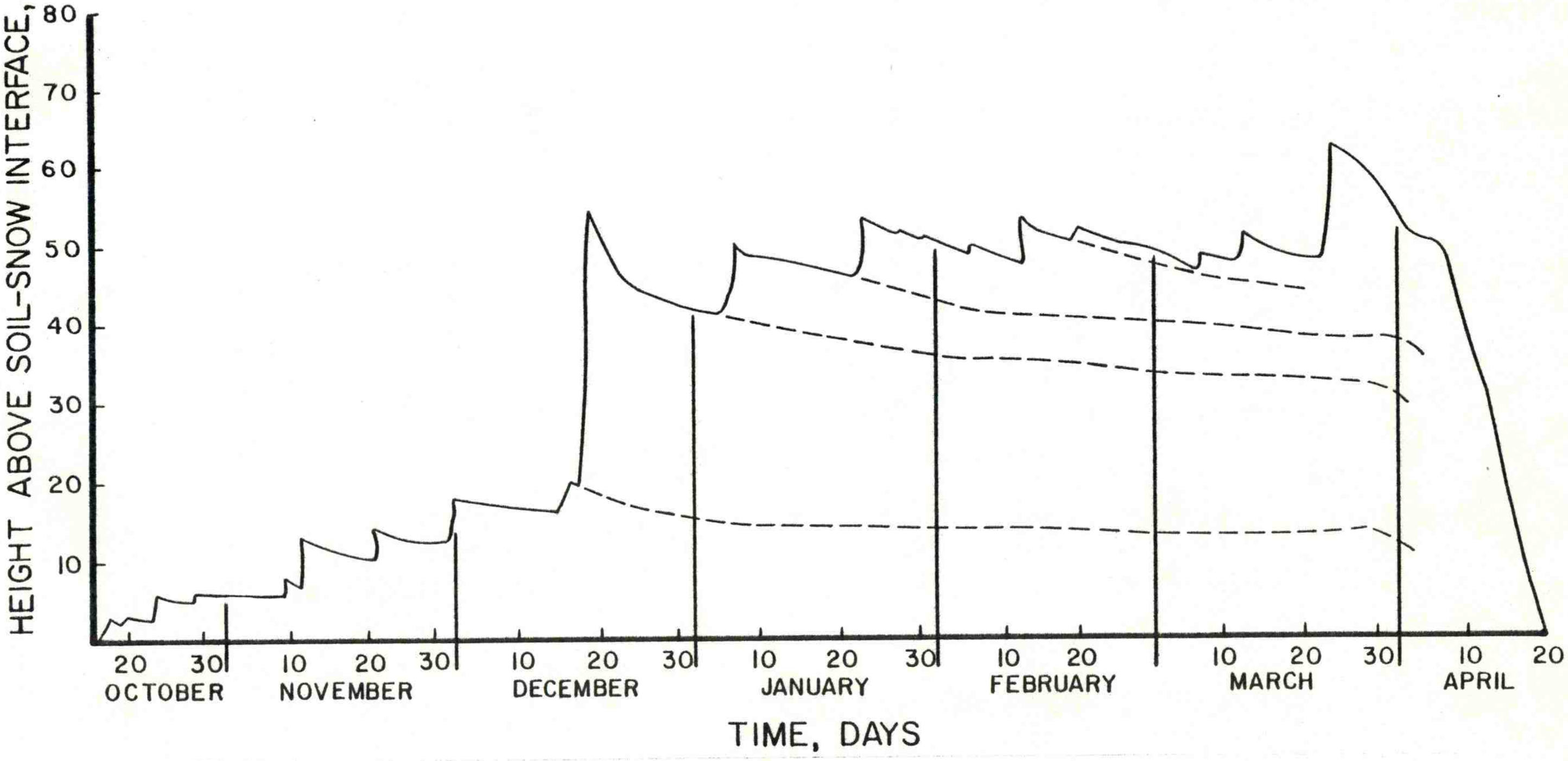


Figure 4. The 1968-69 snow pack near Fairbanks, Alaska. No melting occurred in this snow from October until nearly the end of March. The settling of the snow surface, especially immediately after a fresh snowfall, is due to compaction of the snow. Once melting began, on about 8 April the snow pack disappeared in less than two weeks.

gradients in the snow lead to extensive depth hoar development (Figure 5), and in extreme cases the snow pack becomes predominantly depth hoar. This is the dominant feature of Taiga snow (Trabant and Benson, 1972).

Depth hoar is the coarsest grained snow structure that can form in the absence of liquid water, it consists of large well-developed skeletal crystals which are weakly bonded together, and it forms in the snow as a result of upward vapor transport along the vapor pressure gradients which accompany temperature gradients in the snow. Depth hoar in interior Alaska has the lowest density (less than or equal to 0.20 g cm^{-3}) of depth hoar formed anywhere.

The detailed physical mechanisms involved in the development of depth hoar are the subjects of continuing research. Diffusion of water vapor along the vapor pressure gradients in the snow is certainly involved and convective overturn of the air within the snow appears to occur in extreme cases. The strength of the temperature gradient, and the maximum temperature in the system are the important variables. Where these reach extreme values, as in interior Alaska, the development of depth hoar is also extreme.

The upward flux of water vapor in the snow also exerts a significant drying action on the top 5 cm of soil and vegetation below the snow, which may be as important to winter hardiness of plant species as the low temperature.

One of the most important aspects of the well-developed depth hoar layer is its biological role. Small animals, such as mice, lemmings shrews, etc., can easily tunnel along the snow-soil interface through this low density, coarse, weak snow (Figures 2 and 5). This enables



Figure 5. A thin section of shallow taiga snow, exposed by carefully digging so that the sun could shine through it. In this case almost the entire snow pack was transformed into depth hoar. Scale is revealed by the meter stick lying on top of the thin section.

them to feed and to nest beneath the snow all winter. They count on its insulation to reduce loss of body heat, and on its cover as protection from predators. Large animals such as moose and caribou have developed techniques for travelling and feeding in snow covered areas. These matters have been treated extensively and in a highly readable, informative way by Pruitt (1960).

Not all of the snow pack of the interior of Alaska is as described above, however. Hard-packed wind slabs develop on the hilltops above the Tanana Valley, where winds can be strong, in contrast to the valleys. About 300 m above the valley floor wind action on the snow becomes noticeable, and significant wind packing occurs every winter at altitudes in excess of 400 m above the valley floors. Above the altitude where wind acts on the snow, the taiga snow becomes tundra snow, with wind-packed layers overlying depth hoar layers.

The Maritime Snow of Southern Alaska

The coastal mountains and lowlands of southeastern and southcentral Alaska receive heavy precipitation from Pacific cyclonic disturbances which move through the Gulf of Alaska. The maritime snow pack is thick, L. Mayo and D. Trabant of the U.S. Geological Survey have measured snow depths up to 15 m on coastal glaciers. These snow packs often contain more than 5 m of water equivalent. This maritime snow is common along the Pacific rim northward from Japan and California. Its temperature is significantly higher than that of tundra or taiga snow and it is often wet, especially at low altitudes.

In the transitional climatic zone, conditions alternate between continental and maritime. In the interior this transitional zone becomes apparent as one progresses westward toward the Bering Sea, especially west

of Koyukuk (about 158°W) on the Yukon-Kuskokwim delta. Here, the temperatures and winds are higher than farther east and the climate becomes more maritime; many of the snowstorms are mixed with rain, and the snow cover is characterized by significant amounts of icing with depth hoar at the bottom.

The mountainous maritime snow is responsible for most of the avalanche hazards in Alaska. It also includes all of the largest glaciers in North America. The huge glaciers of the Chugach, St. Elias and Wrangell Ranges which cover 50,000 km^2 owe their existence to the enormous snowfall from the North Pacific.

Acknowledgements

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Weather and Spring Breakup

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The melting of the winter snowpack and the disintegration of the ice cover on Alaskan rivers and lakes are two of the most conspicuous processes that form the annual spring breakup. The annual destruction of the only bridge crossing the Chena River near Fairbanks during the spring breakups in the earlier part of this century is a dramatic reminder of the severity of these events (Fig. 1). These breakups continue to pose problems even now. The character of each breakup is determined by the snow and ice covers developed during the winter, and the weather conditions occurring during the months of April and May. This article will illustrate the effects of weather on producing two different types of spring breakup.

On the large river basins in interior Alaska, such as the Yukon and Kuskokwim Rivers, there is a regular sequence of events that determines the breakup process. It begins in April, when air temperatures rise above the freezing point and the low elevation snowpack starts to melt. The melt rate is fairly slow at first, due to cool temperatures, and most of the melt water is absorbed by the soil. By late April the air is warmer and the melt rate more rapid, causing some of the melt water to run off into streams and rivers. This runoff raises the level (stage) of the rivers, lifting the ice. The drag on the bottom of the ice from the increased stream flow fractures the ice into large floes which may in turn break into smaller chunks. If a large floe is resistant to



Fig 1 Spring breakup of the Chena River at Fairbanks earlier this century, when the wooden bridge over the river was destroyed annually by the ice. Photo made available from the Charles Bunnell Collection in the archives of the University of Alaska in Fairbanks.

further fracturing due to the strength of the ice, it can become lodged in a sharp bend or narrow reach of the river channel, forming an "ice jam." As more ice drifts into the "jammed" floe a tight wedge of broken ice pieces develops in the channel. This wedge can build downward toward the river bed and effectively reduce the flow of the river. Flooding upstream of the jam point will result if the flood plain is low and flat. Fig. 2 shows an ice jam that lasted for 5 days in May of 1972, flooding the village of Aniak, Alaska.

The severity of flooding from ice jams is directly related to the strength of the ice when it is initially lifted and broken. A snow cover over ice will shield it from solar radiation which is the most effective agent in the ice weakening process. Since the snow cover on both the ice and land surfaces melts simultaneously, the snow melt rate becomes a critical factor in defining the character of spring breakup.

Historically the most severe breakups have been associated with lower than normal spring temperatures and, conversely, the easiest breakups occur with above normal spring temperatures.

During a warm spring the low elevation snow cover usually melts and runs off before the mountain snows begin to contribute meltwater. River stages rise slowly allowing time for the ice weakening processes to set in. When the mountain snowpack begins to contribute runoff later in the season and river stages rise high enough to break the ice, the ice is weak and easily crumbles into small pieces that are flushed down the river channel with little or no ice jam formation.

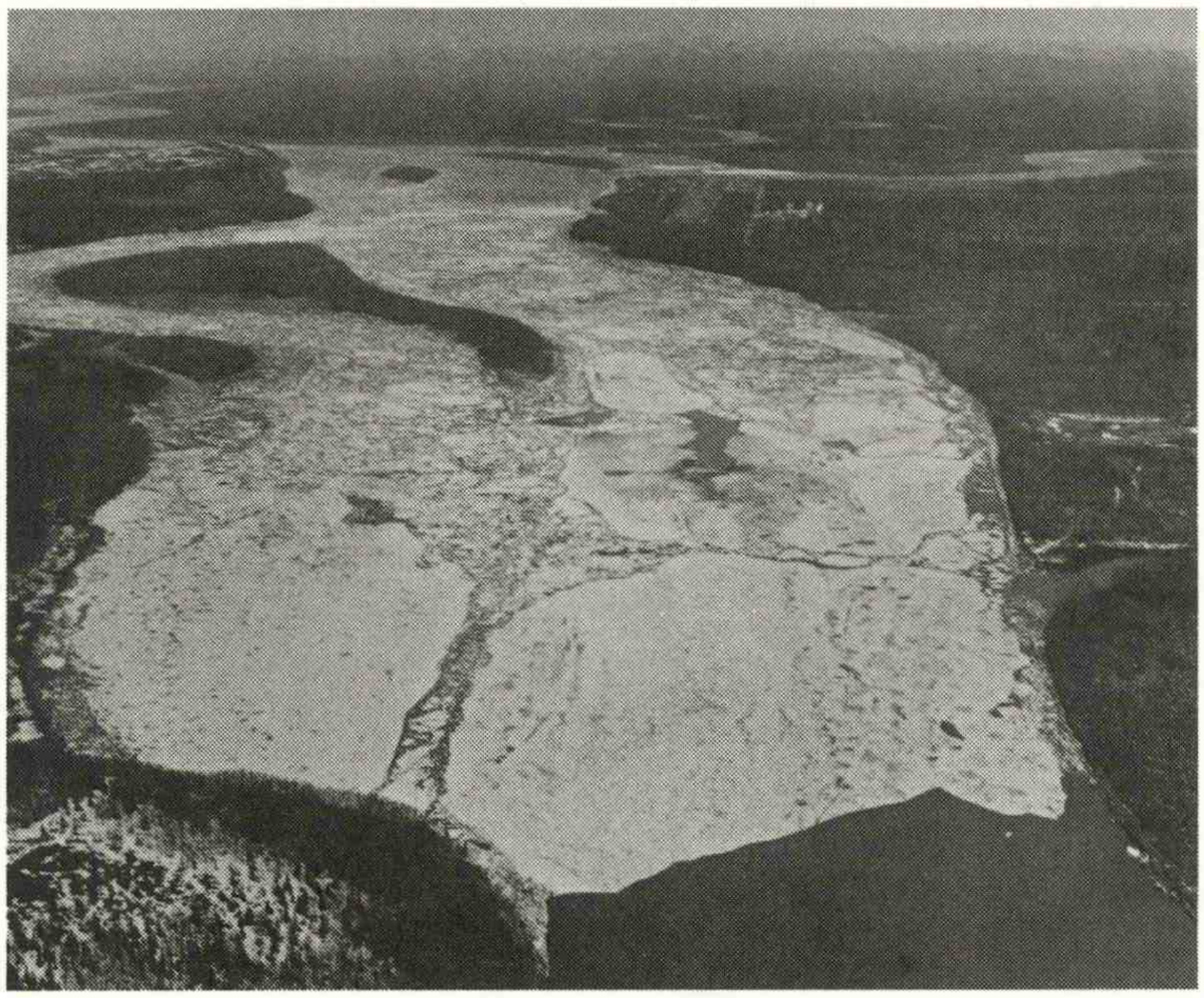


Figure 2. An ice jam formed at a constriction in the channel of the Kuskokwim River about 5 Km downstream of Aniak, Alaska. The airport runway at Aniak is visible on the right river bank in the upper center of the picture.

A long spell of cold, snowy weather during the spring sometimes ends with an abrupt warming to temperatures well above normal. This situation permits both the high and low elevation snowpack to melt and run off concurrently, causing the rivers to rise to high stages in a relatively short time. The ice is then subjected to forces that can break it into the large strong floes which can initiate ice jams.

The net energy available to drive the breakup process is dependent upon solar radiation, air temperature, humidity, cloud cover, wind speed, and the temperature of the water. However for illustrative purposes the daily maximum air temperature can serve as a useful index of the daily net energy.

A classic example of a severe breakup was the spring of 1964. Numerous ice jam floods occurred at many locations on the Yukon and Kuskokwim Rivers. (Anchorage Daily Times, 1964)

Snowpack conditions were near normal over most of Alaska at the beginning of April (EDS, 1964). Occasional periods of snow during April and the first three weeks of May maintained a high albedo on the snow cover which helped to reduce the amount of solar radiation available for weakening the river ice and melting the snow.

In contrast, the spring of 1978 was a very easy breakup. No ice jam flooding was reported on either the Yukon or Kuskokwim Rivers. The ice jams which did form did not have the strength to impound much water.

During April and May of 1964 a record cold spring preserved the strength of the ice until the sudden warming in late May from +2°C to +15°C in 6 days. At that time the snow began to melt very fast from the combined influence of warm temperatures and relatively high solar radiation. River stages rose rapidly on the Kuskokwim River and on the headwaters of the Yukon River. Some of the worst ice jam and snow melt flooding on record in Alaska followed during the first week in June.

The 1964 and 1978 spring breakup are good examples of rather extreme conditions. Most seasons lie somewhere in between. During the average breakup, ice jam problems are the rule rather than the exception. But the number of ice jams and the duration of an individual jam will be dependent upon the day to day weather conditions during breakup.

Snow cover was about normal in the Yukon Basin and slightly below normal in the Kuskokwim drainage at the beginning of April 1978 (EDS, 1978). A light snow fall occurred in mid April which was the last snowstorm of the season.

Since there was not a significant difference in the total winter snow accumulation between the 1964 and 1978 seasons, and rainfall was not a factor, the difference between the two breakups may be explained by the distribution of temperature in April and May.

A comparison of the maximum temperature traces for Bethel, Alaska, during the two seasons are shown in Fig. 3. Note that the climatological mean maximum temperature crosses the freezing point in mid April. In 1964 the temperature steadily diverged from the normal for the 6 weeks from mid April until late May, with the maximum temperature averaging close to the freezing point during the period. In contrast the 1978 maximum temperatures maintained a steady climbing trend during April and May averaging about 5°C above normal.

It is also important to note that the climatological average clear sky solar radiation for central Alaska increases from about 400 ly/day on April 1 to 760 ly/day on June 1.

The 1978 breakup was a case of a sustained slow melt rate beginning in early April when solar radiation was moderate and temperatures were low. By the time the temperatures began to exceed +10°C in early May the ice had turned to slush on the Yukon and Kuskokwim Rivers and ice jams were no longer a threat.

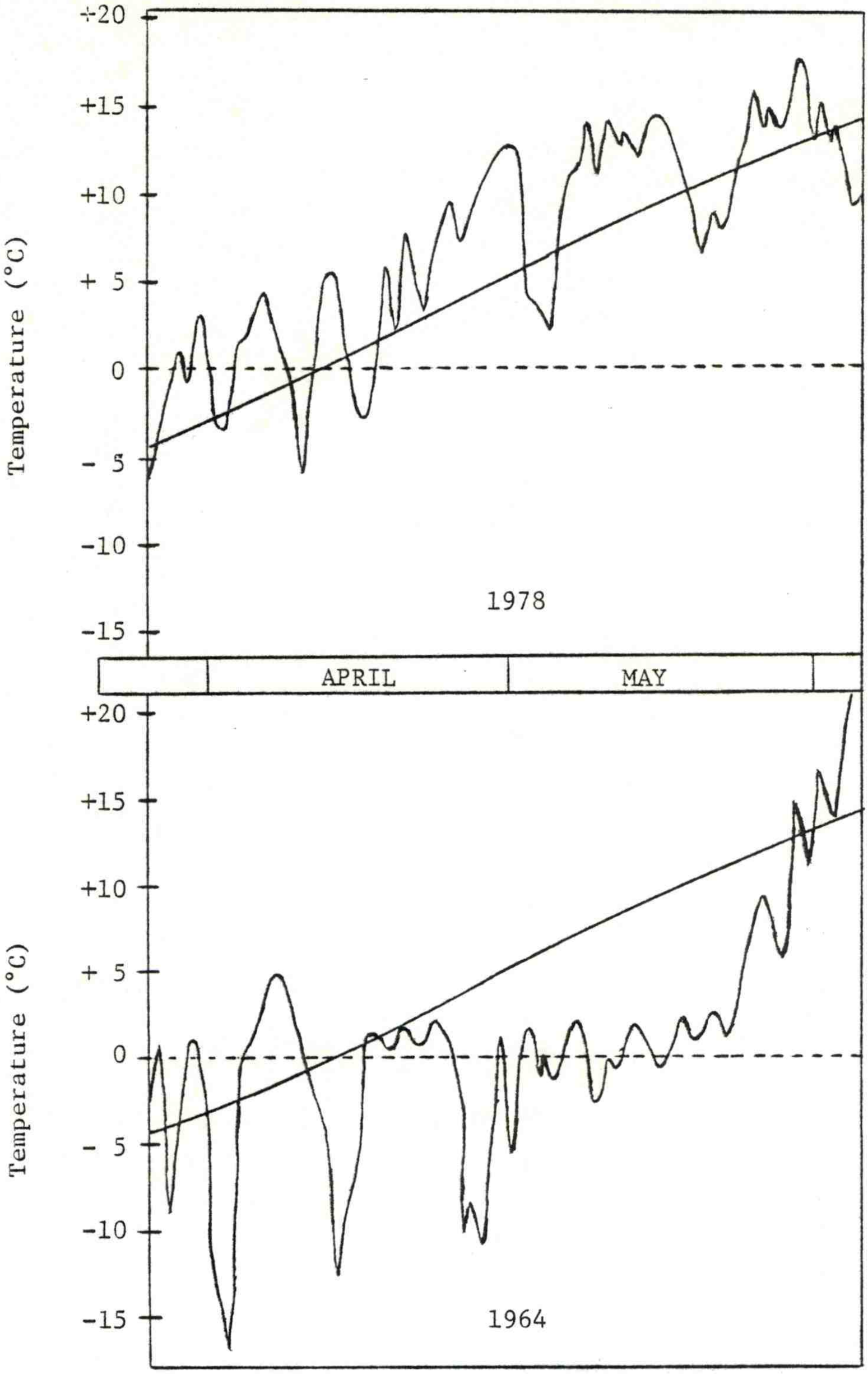


Figure 3. Daily maximum temperature trace for Bethel, Alaska, during the months of April and May. The smooth sloping line is the climatological normal maximum temperature.

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MICROCLIMATES

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The term microclimate refers to the climate of a relatively small area, between several meters to several kilometers in horizontal dimensions and only a few meters in depth from the ground up. Micrometeorology and Microclimatology are the studies of the meteorology and climatology on these scales, respectively. Meteorological conditions near the ground can vary widely over small distances. When approaching the surface, wind speed is reduced by friction from the ground, with the effect of reducing the amount of mixing of the air, which otherwise would have reduced temperature differences over small areas. Terrestrial surfaces also absorb and reflect solar radiation, and lose energy by radiating in the infrared as a function of their temperatures. Natural surfaces found in Alaska vary widely in reflectivity, e.g. values above 80% are found for new snow, ice typically has a value around 30%, vegetation between 10 and 20%, while a lake has a reflectivity below 10%. The physical characteristics of any surface determine how heat energy that reaches the surface by radiation, conduction from the ground and convection in the air is distributed to heat the ground, evaporate water or melt snow. These characteristics thus determine the climate of the layer near the ground in any locality. Since all vegetation and most human and animal life exists in this layer, knowledge of it and processes within it are of great practical importance.

To demonstrate the large variability of conditions near the ground, Wolfe, et al. (1949) compared the microclimate of an area of 0.6 km^2 in Ohio, where they established 109 micrometeorological stations, with the macroclimate of the whole State of Ohio, (133,000 km^2) as reported by 88 official weather stations. The results, as seen in Table 1, show the much greater range of variability observed by the micrometeorological stations, despite the fact that they were located in a much smaller area.

Table 1
Comparison of macroclimate and microclimate

Meteorological variable	At the 88 meteorological stations in Ohio	At the 109 microclimate stations in the Neotoma Valley in Ohio
Highest annual temperature (°C)	33-39	22-45
Time of occurrence of the above	17-19 July	25 Apr.-19 Sept.
Lowest January temperature (°C)	-21 to -29	-10 to -32
Latest spring frost	11 Apr.-11 May	9 Mar.-24 May
Earliest autumn frost	25 Sept.-28 Oct.	25 Sept.-29 Nov.
Length of frost-free period (days)	138-197	124-276

Similar differences can be expected to exist in Alaska. The meteorological measurements that are necessary to determine the microclimate basically quantify how much heat is available to heat the air or ground, evaporate water, melt snow, etc., by establishing an energy balance at the surface in which the algebraic sum of all energy sources and sinks at the surface is equal to zero (i.e. energy is being conserved).

$$N + Q + E + S + M = 0$$

Where N = Net flux of radiation or radiation balance (solar and infrared)

Q = Turbulent heat flux (sensible heat) in the atmosphere

E = Latent heat released by condensation or heat used by evaporation

S = Conductive heat flux in the soil

M = Heat available to melt snow.

The first term can be directly measured with a radiometer, the next two terms can be determined by measuring gradients of temperature, wind and humidity in the air and calculating the heat fluxes by using turbulent exchange formulas, and the last term can be found by measuring temperature gradients in the soil and calculating the heat conduction as the product of the temperature gradient and the thermal conductivity of the soil. Since the wind speed close to the ground increases roughly exponentially with height, a logarithmic spacing of sensors at 0.5, 1, 2, 4, etc., meters is usually chosen, as seen in Fig. 1. A few typical examples will be given to illustrate some of the many different microclimatological regimes in Alaska.

Glacier microclimates

Alaska has a great number of glaciers in its three major mountain ranges. Glacier changes are, of course, good indicators of climatic change and therefore heat balance studies of glacier surfaces have been carried out for a long time. The heat balance studies show whether a glacier is growing or melting and at what rates. A typical example of the heat balance of a high latitude glacier, in this case the McCall Glacier in the Brooks Range, studied by Wendler and Ishikawa (1973) is shown for a limited period in Fig. 2. The figure shows the substantial

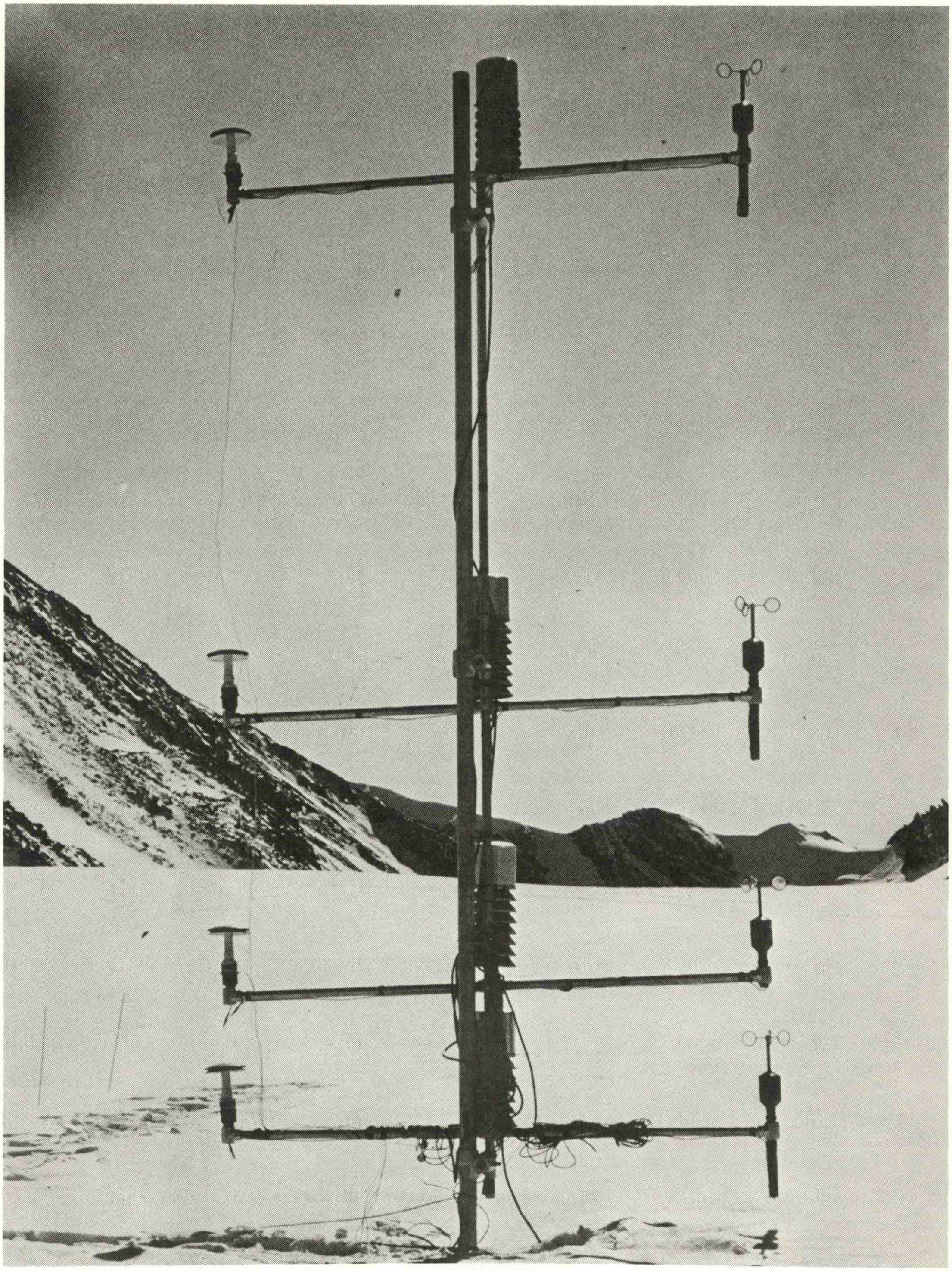


Figure 1. A typical station for determining heat balance components on a glacier. On the left are shielded and ventilated temperature sensors at 0.5, 1, 2 and 4 meters height, in the middle are humidity sensors and on the right anemometers. The whole array can be raised or lowered after snowfalls or melting to keep a constant height above the surface.

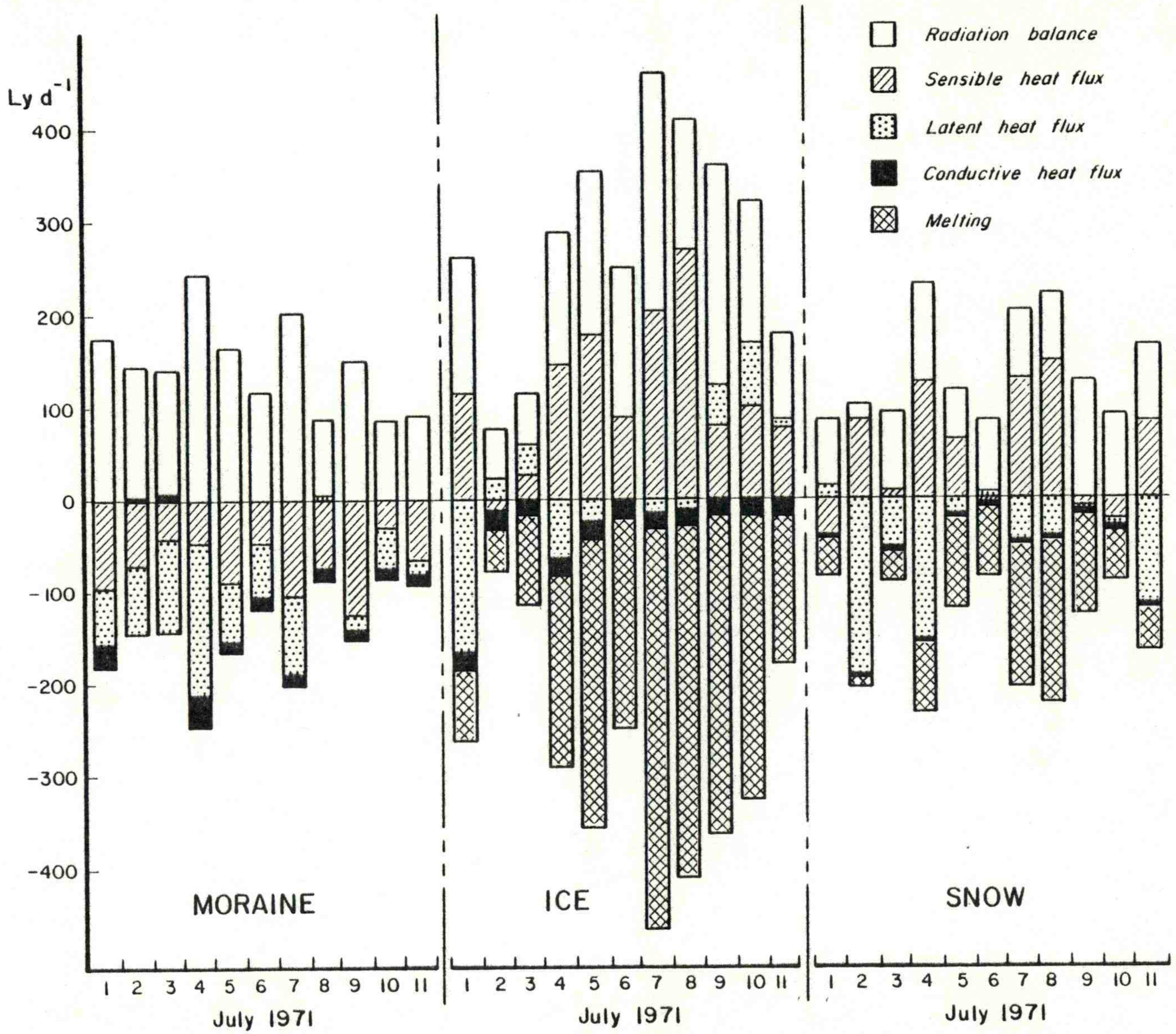


Figure 2. Daily values of the components of the heat balance for three different surfaces in close proximity, on McCall Glacier, summer 1971.

differences in the heat balance components of three surfaces in close proximity to each other. Similar studies have been carried out by other glaciologists and meteorologists on Alaskan glaciers. The kind of conclusions that can be drawn from such investigations are illustrated by the following examples.

Hubley's (1957) studies on the Lemon Creek Glacier on the Juneau Ice Field showed how the different atmospheric and surface conditions can influence the partitioning of the available heat energy. Among his findings are the following interesting details:

1. There was frequently more energy available for melting snow with overcast skies than under clear conditions. This often happens at relative low solar elevation angles when the long wave (infrared) radiation from the clouds increases more than the short wave (solar) radiation reduction by the clouds. High surface reflectivity (albedo) aids this phenomenon. A high albedo can be caused by multiple reflection between the snow surface and the underside of the clouds. Holmgren (1971) found the same results in Devon Island, and later on Ambach (1974) verfied it for Greenland.
2. A term that was not added to the energy balance equation above is the energy imported to the surface by falling rain. Hubley showed that the energy transfer to snow by rain falling on an isothermal melting snow pack is an insignificant part of the total energy transfer. When rain falls on snow at temperatures below 0°C, some latent heat will be transferred from the rain to the snow however.

3. Hubley also found that the snow's reflectivity varied strongly with solar elevation angle. This can be explained in terms of increased forward scattering of incoming radiation, but instrument errors have to be carefully evaluated, since solar radiometers normally change their sensitivity with solar elevation angles less than $\sim 15°$.

Streten and Wendler (1968) and Wendler and Streten (1969) carried out similar short term heat balance studies a decade later over ice and snow in the same location. It can be clearly seen by comparing column 1 with column 2 (snow) of Table 2 (representing the ice surface) of another glacier in the coastal range that the net radiation is substantially lower over snow than over ice. This is, of course, caused by the higher albedo of the snow as compared with ice. In both cases the latent heat flux was positive, which means that condensation is higher than evaporation. This is rather typical for coastal Alaskan areas, where wet, cloudy or foggy weather is frequent. The net radiation in both cases is about half of the energy needed for melting while the other half is provided by the sensible and latent heat fluxes from the atmosphere.

In the already mentioned McCall Glacier studies in the Brooks Range in northern Alaska, the greatest differences from the more southerly glacier studies were found in the latent heat flux (Table 2). While in southern Alaska condensation is higher than evaporation, in the drier climate of northern Alaska - the evaporation is on the average higher than the condensation. This reduces the amount of energy available for melting, and the melting rates found for both snow and ice are about half of those observed for southern Alaska as shown in Table 2.

One of the other unexpected findings on McCall Glacier was the fact

Table 2

Heat budget measurements in ly day^{-1}

	1	2	3	4
Net radiation*	262	84	149	88
Sensible heat flux	141	74	116	52
Latent heat flux	97	16	-10	-55
Heat flux of snow or ice	0	0	-19	-1
Snow melt	-508	-181	-236	-83

*All fluxes towards the surface are considered positive, away from the surface negative

1. Worthington Glacier (lat 61° 10'N long 145° 45'W)
   Ice, 16 July - 1 August 1967 (Streten and Wendler, 1968)

2. Lemon Creek Glacier (lat 58° 23'N, long 134° 21'W)
   Snow, 5 - 20 August 1968 (Wendler and Streten, 1969)

3. McCall Glacier (lat 69° 18'N, long 143° 41'W)
   Ice, 1 - 11 July 1971 (Wendler and Ishikawa, 1973)

4. McCall Glacier (lat 69° 18'N, long 143° 41'W)
   Snow, 1 - 11 July 1971 (Wendler and Ishikawa, 1973)

that the temperature at 10 m depth was higher at high altitudes than at low altitudes. This finding is significant. Normally, temperatures decrease with height (in the atmosphere at 1°C/100 m) and the 10-meter depth snow temperature is often taken as the approximate annual mean air temperature of that location. The cause for the anomalous altitude dependence on McCall Glacier is that melt water penetrates the snow cover, releasing latent heat by refreezing, and warming lower depths to a much larger extent than would be possible by conduction alone. Hence, the temperatures found within the glacier at this altitude are far above the annual mean air temperature. At lower altitudes, water cannot penetrate the solid ice and a temperature much nearer to the annual mean is observed. Normally slope, exposure and mountain screening of the sun have a strong influence on the microclimate of a specific setting. This is seen in the Brooks Range, where most of the remaining glaciers (the Brooks Range was heavily glacier-covered during the last ice age) are protected from the sun by a northerly exposure and located in narrow valleys.

The microclimates over snow-covered terrain

Snow covers most areas of Alaska for long periods each year (see Benson's article elsewhere in this report). Its presence modifies the climate on a large scale as well as on the microscale. Steep temperature gradients usually exist within the snowpack which provides protected and relatively warm shelters for a variety of animals at the snow-ground interface. Some animals, for example the lemmings, live most of their lives under the snowpack. The first snowfall and the melting of the last snow mark the endpoints of a number of distinctly different annual

climatic intervals. Weller and Holmgren (1974) have used heat balance considerations to classify the microclimates of the arctic tundra into six distinct periods, all having quite different heat balances. These six periods are: freeze-up, mid-winter, pre-melting, snow melting, post-melting and mid-summer. It is interesting to note that the snow-free mid-summer period north of the Brooks Range is only about eight weeks long. The most explosive change from one period to another occurs when the snow disappears, generally within a week or so, drastically reducing the surface reflectivity and increasing the amount of solar radiation absorbed by the ground. Wendler (1967) has carried out similar measurements on the snowpack of interior Alaska, near Fairbanks, with similar results.

Typical temperature gradients in the air, snow and soil are shown in Fig. 3, measured at different times during the year at an arctic tundra site near Barrow. The modifying influence of both the snow cover and the low tundra vegetation on the soil surface temperature can be clearly seen. Also, typical winter inversions, as shown by the temperature data of 10 February indicate air temperature differences between the snow surface and 16 m height of over 10°C. The great variability of temperatures within a layer only a few meters thick is quite striking.

Studies of other micro-meteorological and climatological phenomena in Alaska

With the advent of the discovery of large oil reserves in northern Alaska, interest in new micrometeorological studies was immediately generated. Blowing snow on the North Slope was soon found to hinder oil exploration and drilling operations. Wendler (1978) examined relation-

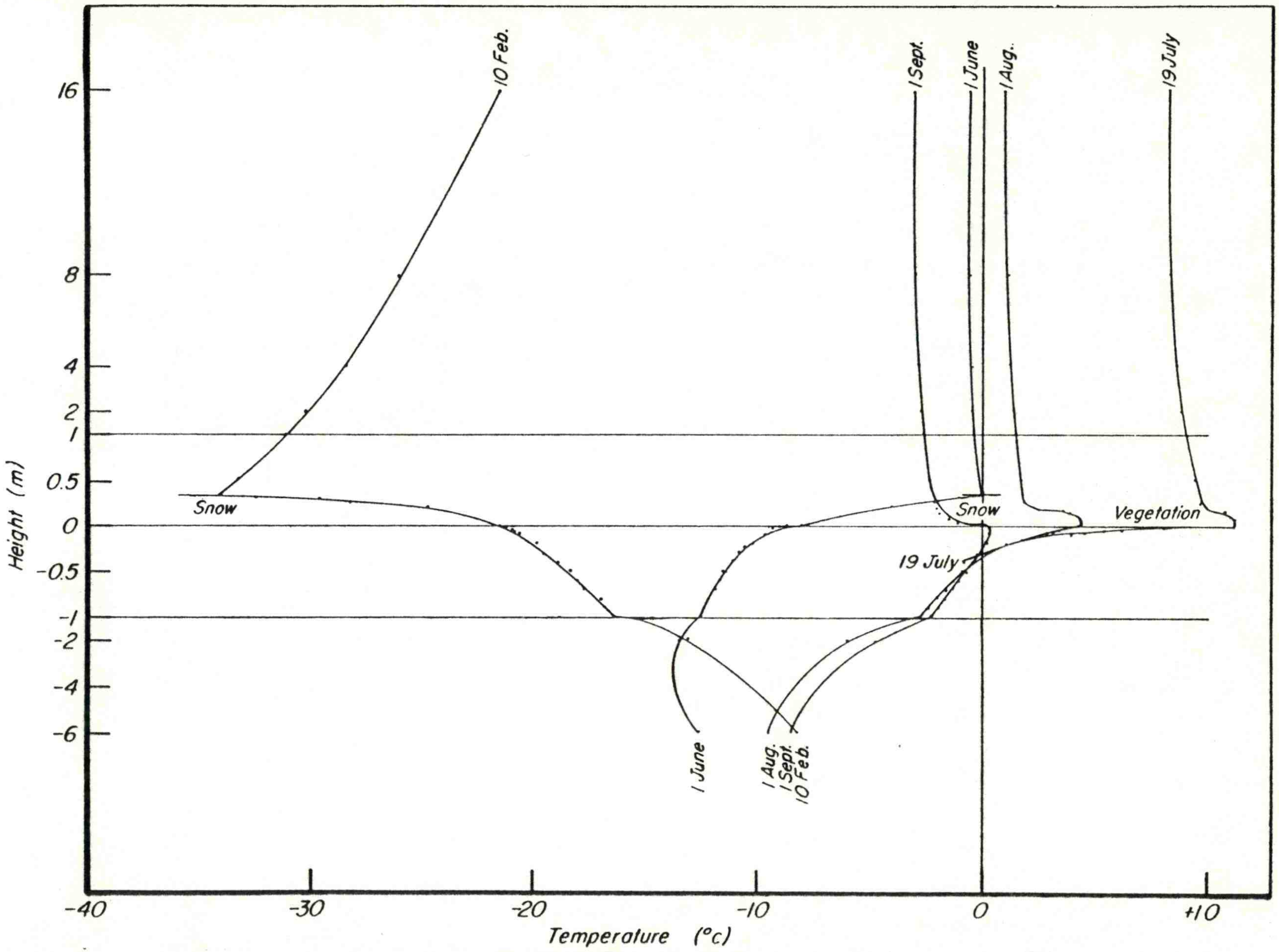


Figure 3. Typical temperature gradients in the air, snow, vegetation and soil of arctic tundra at Barrow. The vertical height scale from +1 to -1 m is expanded.

ships between blowing snow and visibility reduction for Barrow and Barter Island and analysed the occurrence of blowing snow at both locations. Since this phenomenon occurs in the lowest layer of the atmosphere, the visibility reduction associated with blowing snow also has an impact on military ground maneuvers. On the other hand, if it is possible to deposit blowing snow where it is needed, it can augment local water supplies and can be used for constructing roads and airfields.

Another study which started only recently evaluates the microclimatic changes which are caused by a body of water. Preliminary work carried out so far at Lake Minchumina in interior Alaska showed that extreme temperatures are suppressed by the presence of the relatively large water mass. This results in a delay of the frost in autumn, thus prolonging the growing season. The potential importance of this for agricultural productivity is significant. Tree rings analyzed downwind of the lake showed approximately twice the growth rate as trees growing under similar soil, precipitation and slope conditions several kilometers from the lake.

Another study compared the microclimate of two greatly contrasting sites at Ft. Greeley in Interior Alaska (de Percin 1960); one of them located in a coniferous forest (taiga), the other one in an adjacent clearing. It was found that the spruce forest effectively decreased solar radiation received at or near the ground surface in both summer and winter. In June total daily solar radiation in the forest was only one third of that received in the clearing. The vegetation of the forest and, in winter, the deeper snow in the clearing, caused very large differences between the stations in soil temperature, ground

surface temperature and above-surface temperature in the lowest 50 cm. Wind speeds were, of course, much lower in the forest when compared with the clearing.

In winter the net radiation is negative for most places in Alaska, most of the time; in other words, there is a net loss of radiation by infrared emission. This cools the surface and creates strong surface inversions. In urban communities in Alaska these strong inversions trap the available water vapor, and ice-fog results when the temperature drops below -30°C. This phenomenon is discussed by Benson elsewhere in this report. Another kind of inversion not caused by radiation is found in summer over fjords in southern Alaska; especially in those fjords into which glaciers calve icebergs. The ice cools the water steadily and the air layer directly above the water is much warmer; i.e. the air is stably stratified. These conditions are conducive to trapping the pollutants from the smoke stacks of the numerous cruise ships which are sailing into Glacier Bay National Monument near Juneau, Alaska (Fig. 4).

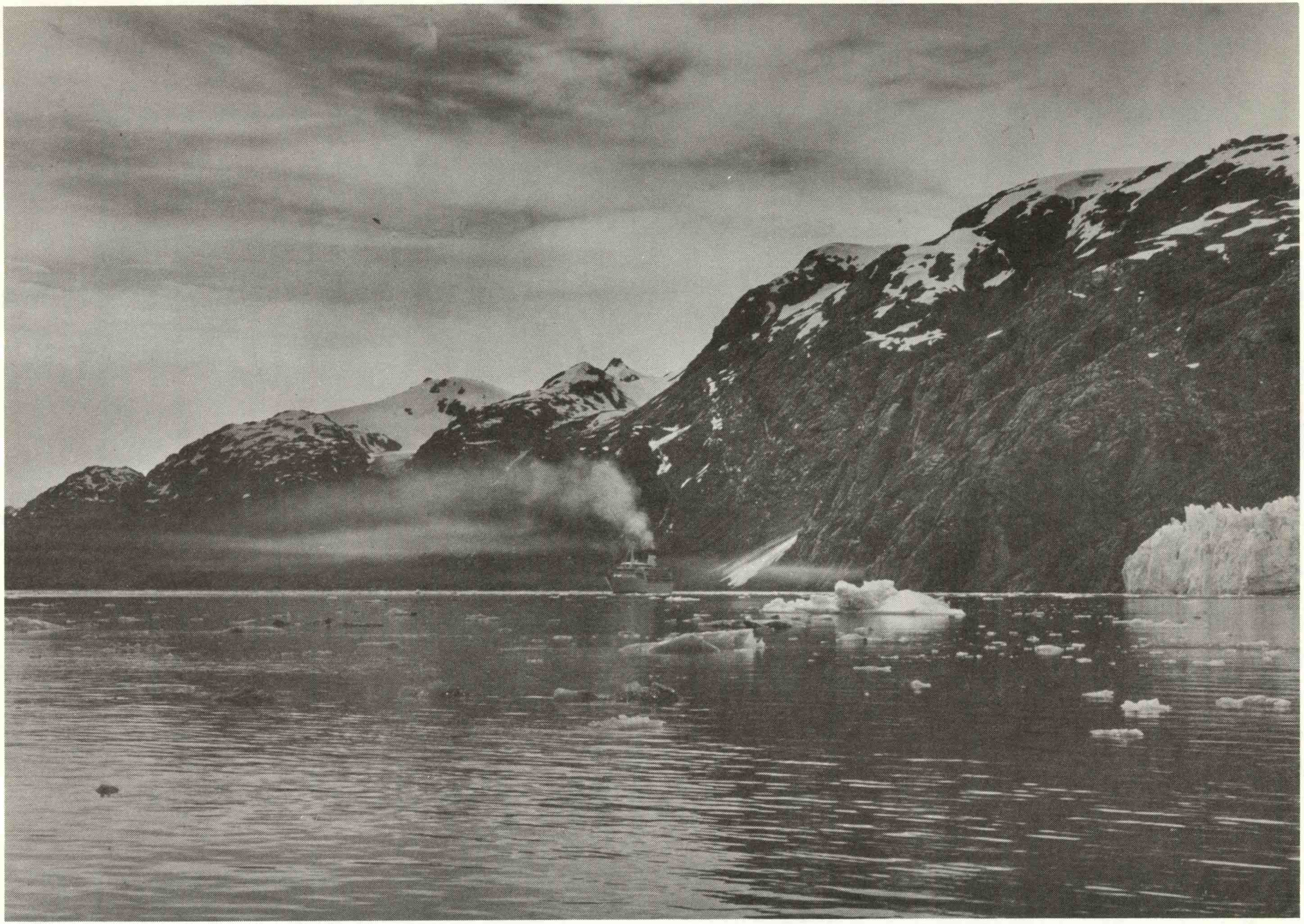


Figure 4. Photograph of pollution from a cruise ship, trapped over cool water in a fjord of Glacier Bay National Monument.

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AIR POLLUTION IN ALASKA

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Arctic and subarctic regions are famous for vast unpopulated expanses with crisp, clean air and beautiful mountains. The population density, when expressed in terms of people per km^2, is very low. For example, Alaska's population density of 0.23 people per km^2 is about 100 times less than the overall U.S. average. It is 1450 times less than that of New Jersey and 8500 times smaller than that of the New York metropolitan area. The population density in the adjacent Yukon territory is 7 times less than in Alaska, and in the Northwest Territories it is three times less than that of the Yukon Territory. It is no wonder that most people in the U.S. properly think of Alaska and northern Canada as wilderness areas, to speak of air and water pollution there brings forth smiles, raised eyebrows, and gentle head shaking at first thought. But, on second thought it becomes clear that local problems exist, just as beautiful, wide-open Montana has air pollution in some of its cities. Additional thought reveals that the apparently rugged arctic and subarctic environment tends to minimize one's conception of the effects man's activities may have on this environment.

If Alaska's population were distributed evenly, so that each person had his or her four square kilometers to live in, there would be little reason to talk about pollution problems. However, nearly three quarters

of the people in the Yukon Territory live within 30 km of Whitehorse and more than three quarters of Alaska's population is associated with Anchorage and Fairbanks. In these centers, natural regenerative processes provide a slow recovery from man-made disturbances. Impermeable permafrost, by complicating the dispersal of human waste products facilitates pollution of water and land. And, the air masses over some population centers have proven to be exceptionally stable, prone to stagnation, and vulnerable to air pollution, even with a relatively low input of pollutants.

Temperature Inversions

The high potential for air pollution in arctic and subarctic lowlands results from a combination of low winds and strong temperature inversions, which are frequent and persistent in the near surface air layers. The inversions trap pollutants below them and prevent effective dilution of the pollutants by mixing with higher air layers. The problem is best introduced by discussing the Fairbanks case as the type locality for low temperature air pollution. To gain perspective, it is useful to compare Fairbanks with Los Angeles because temperature inversions are involved in both cases, and the Los Angeles pollution problems are well known. Air temperature inversions can occur at different heights and may be caused by different processes. For example, temperature inversions form in the boundary layer between two air masses when a warmer mass flows over a cooler one. This occurs in the Los Angeles area where cool air from the Pacific Ocean is overridden by warmer air, sometimes from the Mojave Desert. In this example the air temperature is relatively constant from ground level up to 400-500 m through the oceanic air, and then increases sharply as the warmer air mass is encountered aloft. Temperature inversions

can also form at ground level when there is a net heat loss from the earth's surface by outgoing longwave radiation. The radiative balance between the surface and the air aloft causes this cooling to proceed upward in the air as described in detail by Wexler (1936). These inversions occur during both day and night in parts of the Arctic and Antarctic, and are extremely common in Fairbanks. Typical inversions over Fairbanks and Los Angeles are compared in Figure 1.

Both Fairbanks and Los Angeles are partially surrounded by hills which permit stagnant air to form at low levels when inversions are present. However, air stability over Fairbanks is far greater than over Los Angeles for two reasons: (1) Since the Fairbanks inversion begins at ground level there is no air available to mix with, dissolve, and carry away pollutants beneath the inversion layer. (In the Los Angeles case the air below the inversion layer is 400 to 500 m thick.) (2) Fairbanks inversions are usually steeper (up to three times steeper) than those in the Los Angeles inversion layer. In fact, the wintertime surface inversions in Fairbanks are considered to be among the strongest in the world, even when measured by Arctic and Antarctic standards.

Bilello (1966) compared air temperature inversions at eleven arctic and subarctic stations in Canada, Alaska and Greenland. All eleven stations showed frequent inversions with base at the surface, and Fairbanks was one of the extreme cases. Throughout the nine year record, the frequency of surface inversions at Fairbanks exceeded 60% during <u>all</u> night-time soundings the year round; it exceeded 70% for 8 months of the year and 80% for 5 months. During the dark months of December and January, when diurnal changes in the weather are virtually non-existent, surface inversions were present in more than 80% of <u>both day and night</u> soundings.

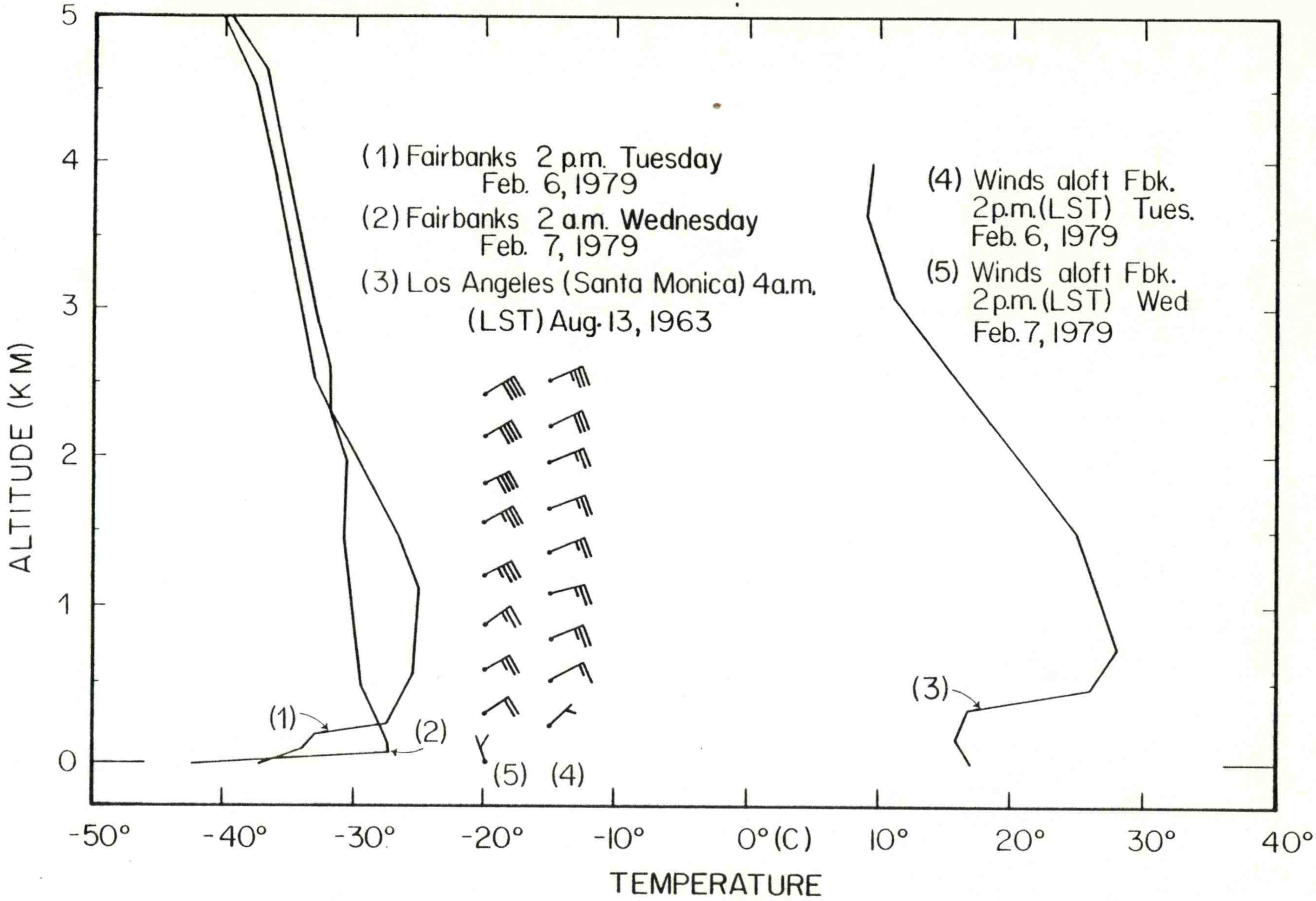


Figure 1. Typical radiosonde profiles measured over Fairbanks and Los Angeles at times when their respective air pollution problems are most pronounced.

Ice Fog

The most serious problems in Los Angeles come with high temperature and lots of sunlight, in Fairbanks they come at low temperature with very little sunlight. Automobile exhausts play important but very different roles in the two cases. Under the low temperature conditions of winter in Fairbanks, the most visible type of air pollution is ice fog (Figure 2). Ice fog occurs at -35°C or below when all water vapor emitted into the atmosphere (from car exhausts, houses, industrial plants and open water surfaces) condenses and freezes into ice crystals. Ice fog was first recognized as an air pollution problem by V. J. and M. B. Oliver (1949), but has since been studied by many researchers including Benson (1965, 1970) and Ohtake (1970).

Ice fog can get very dense because one major effect of low temperature is to severely reduce the amount of water vapor which air can hold at saturation, i.e., before condensation occurs. For example, air at +25°C can hold 255 times more water vapor at saturation than air at -43°C. Thus, a large amount of water vapor introduced into the air at +25°C can be absorbed with no condensation, and no effect on visibility, while less than 1% of that amount of water vapor may make a dense fog of super-cooled water droplets and ice crystals at -40°C.

Man's activities contribute about 6000 metric tons of water vapor per day to the air over Fairbanks which come from many different sources. Water is a major component of exhaust gases in all combustion processes and usually exceeds the mass of fuel which was burned. For example when gasoline is burned in an automobile the complete combustion equation would be:

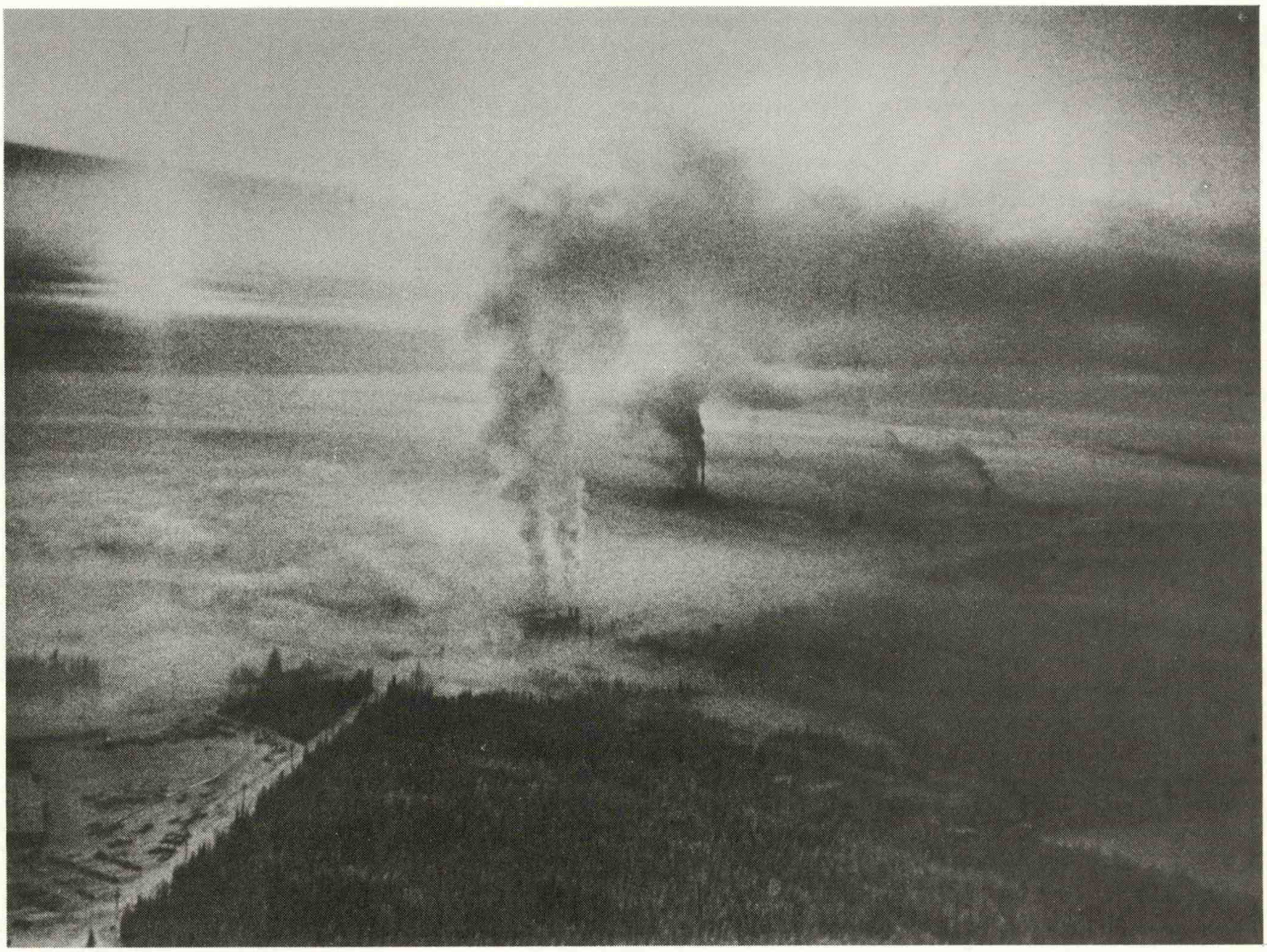


Figure 2. An aerial photograph taken on 3 Janaury 1975. Fairbanks lies under its home-made ice fog to left of center. The exhaust plumes from power plants go in different directions at different altitudes indicating the presence of shear winds. (Photograph by C. Benson).

$$C_8 H_{18} \xrightarrow[\text{oxidized}]{} 8\ CO_2 + 9\ H_2O$$

The resulting mass of CO_2 and water in this reaction is 4.5 times the mass of the gasoline which was burned; specifically, 1 Kg of gasoline yields 3.1 Kg CO_2 and 1.4 Kg H_2O. Generally this reaction is far from complete and carbon monoxide (CO) is a major product which increases in quantity with decreased efficiency of the combustion; its mass commonly approximates the mass of gasoline burned.

Low-Temperature Air Pollution

Most of the air pollution problems discussed in the Fairbanks area have been thought of primarily in association with the ice fog itself. However, ice fog is only the visible manifestation of low temperature air pollution in all regions of the Arctic and sub-Arctic. In that sense it is useful because it increases public awareness of the basic air pollution problems which exist. It enters into the overall air pollution problem in at least three ways: (1) it restricts visibility, (2) it affects the overall thermodynamics of the air mass, and (3) it is an important chemical component. Clearly, water itself must be considered as a pollutant under these low temperature conditions. At higher temperatures, the same amount of water would remain as uncondensed water vapor in the air and would not constitute a problem.

In addition to water (which is so obvious because of its effect on visibility), all other components of air pollution, including lead, SO_2, NO_x, CO and total hydrocarbons, are present as well, and often in high concentrations. The carbon monoxide values are exceptionally high

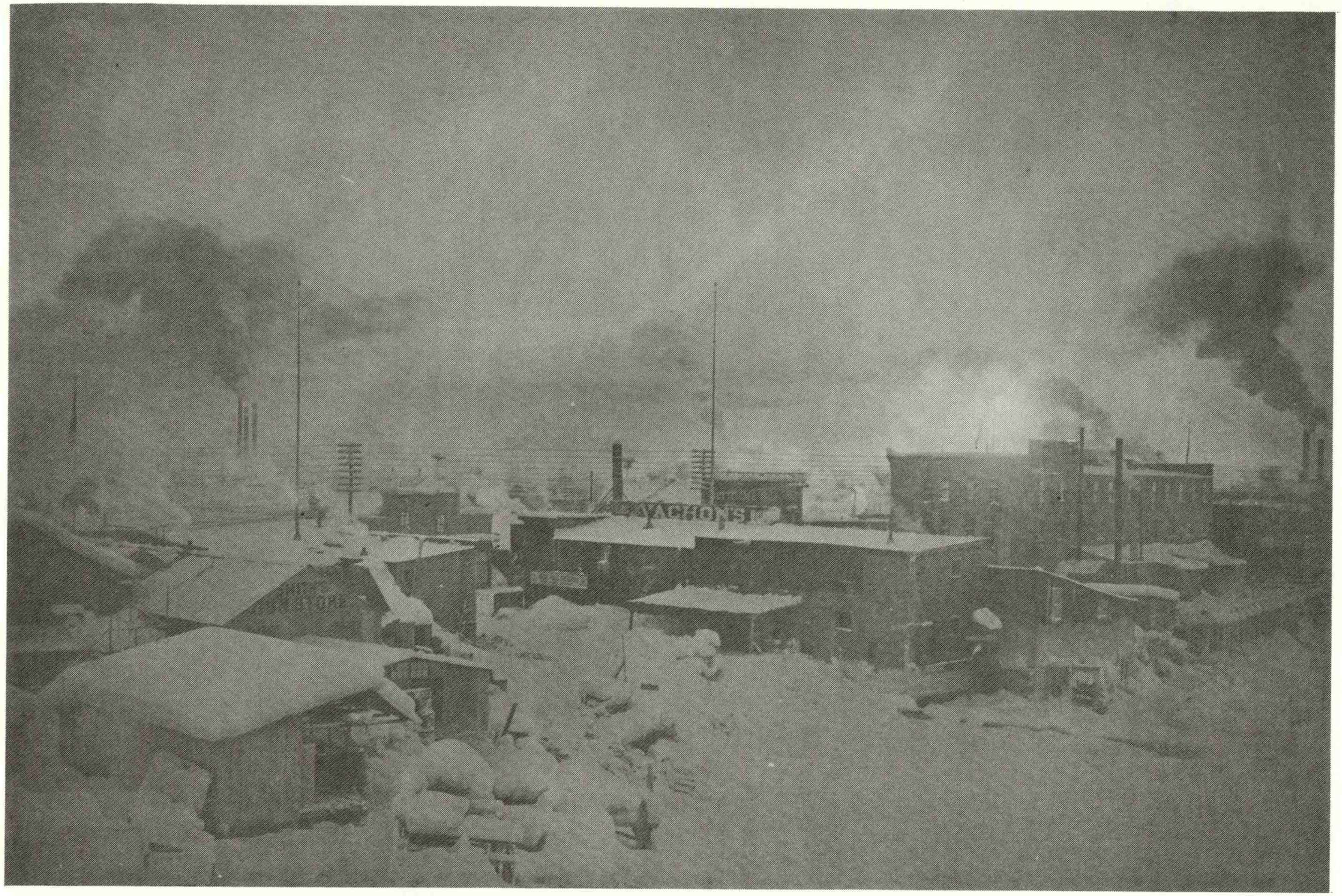


Figure 3. Fairbanks at -50°C (-58°F) on 24 January 1911. Although condensation plumes are visible from power plants and chimneys, the street level visibility is excellent because no automobiles were in operation. Such a photo could not be taken today.

Lithograph Reproduction by Commercial Printing Co., Fairbanks, Alaska

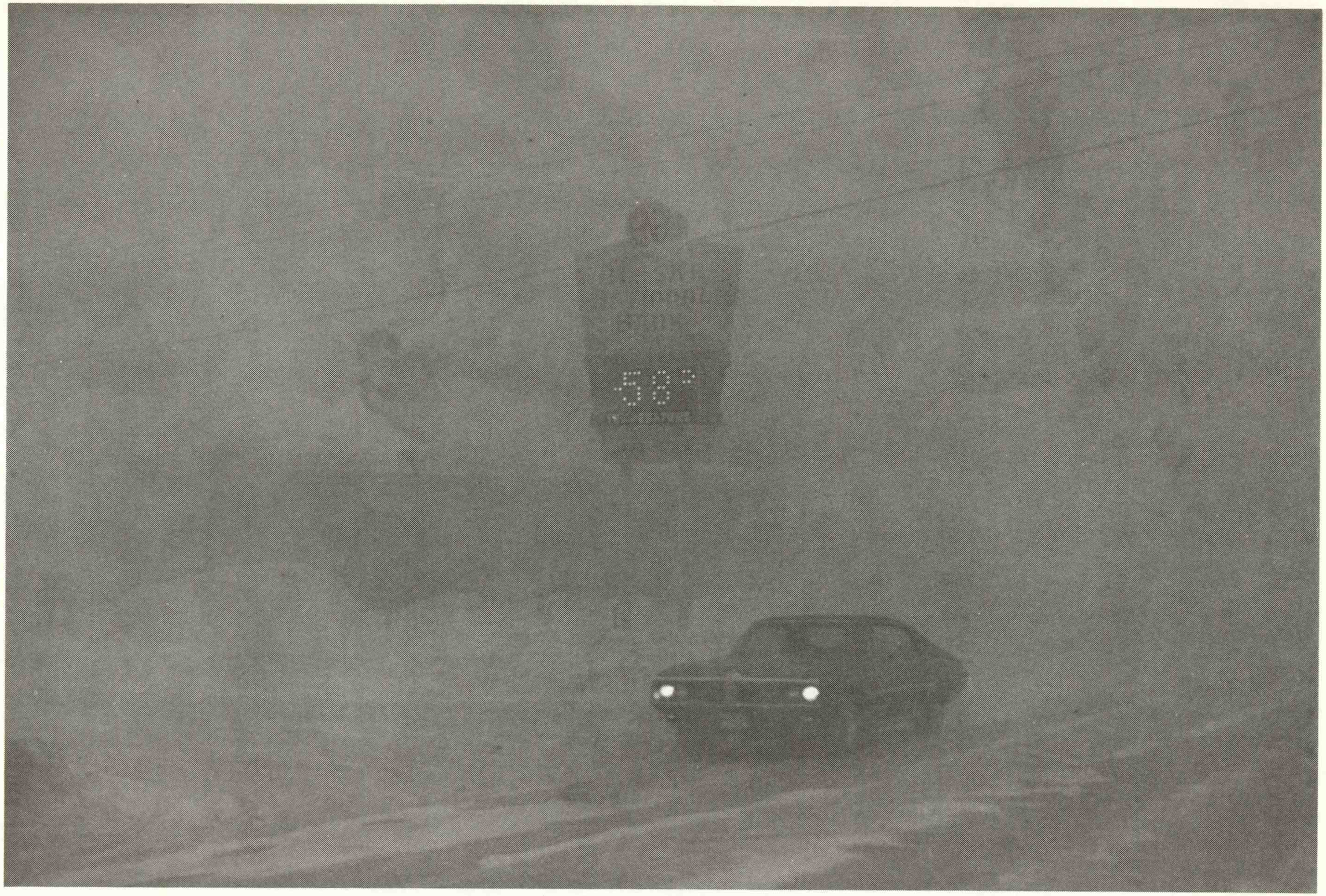


Figure 4. Fairbanks at -50°C (-58°F) on 3 January 1969. This photo, taken at about the same time of day and at the same temperature as the 1911 photo of Figure 3 shows how a plethora of automobiles affects street-level visibility. (Photography by T. Ohtake).

in Fairbanks, especially when one considers the small population involved; this has caused the U.S. Environmental Protection Agency (EPA) to pay special attention to Fairbanks. During a recent study when relatively clear and calm weather prevailed, a total of 201 eight-hour samples of CO were obtained, of which 137 (68%) exceeded the EPA standard of 9 ppm and 58 (29%) were at "alert levels" of 15 ppm or higher (Fisher and Joy, 1977).

Daily values for lead in the air have ranged from 1 to 12 μ m^{-3} during the past decade with monthly values at 6 μg m^{-3} during December 1976 and January 1977 (Coutts, 1979; Winchester et al., 1967). Holty (1973) found that pollution levels doubled or tripled during ice fog episodes. Jenkins et al. (1975) pointed out that pollutant concentrations in the Fairbanks area are as high as in cities such as New York, Detroit and Los Angeles. They considered this surprising when one considers the small size of the Fairbanks community.

Combustion products exhausted at street level, including H_2O, CO_2, and Pb, are primarily produced by automobiles. Because the water vapor condenses to form dense ice fogs it provides an excellent "tracer" by making visible one part of the overall pollution. This is well illustrated by comparing the photographs of Figures 3 and 4 which show Fairbanks at -50°C with and without automobiles.

Of the oxides of nitrogen, NO_2 is generally more common than NO, because the NO formed in combustion is rapidly oxidized to NO_2. In Fairbanks during mid winter, NO is the dominanat species. This is a consequence of the low solar radiation and the lack of photochemically produced ozone. The combustion processes and abundant NO overwhelm what ozone is present (Coutts, 1979).

It is again instructive to compare Fairbanks and Los Angeles. The pollution levels are the same, yet Los Angeles has more than 200 times the population of Fairbanks. Thus, the per capita air pollution potential of the Fairbanks area is higher than that of the Los Angeles area. This emphasizes the extreme stability of air masses in the cold lowlands of interior Alaska.

Thermodynamics and Dynamics

The thermodynamic and dynamic aspects of air pollution episodes are complex and we are just beginning to understand them. The strong ground inversions typical of clear winter weather in interior Alaska are due to surface cooling by outgoing infrared radiation. Ice fog blocks part of the outgoing radiation from the snow surface; this weakens the ground inversion and moves the base of the inversion to the top of the fog layer. Once a thermal structure of this type is established, pollutants entering the air at any height will be distributed throughout the depth of the fog. Pollutants normally emitted at low altitudes, such as CO, can mix upward and those normally emitted at higher altitudes, such as sulfur oxides from exhaust stacks apparently mix downward during ice fog. Under severe and prolonged cold spells, when air temperatures remain below -40°C for more than a week, the ice fog thickness may extend up to and above the highest high-moisture plume in the area, this produces the worst possible dispersion conditions for elevated sources (Figure 2).

Another prominent thermodynamic aspect of the problem is the development of a pronounced "heat island" over Fairbanks (an increase of temperature due to heat losses in built-up areas). The Fairbanks heat island has on occasion been measured as high as 14°C. For a city the size of Fairbanks this is exceptionally high!

In general, air quality and visibility in Fairbanks are very sensitive to synoptic scale phenomena of the atmosphere over interior Alaska. A recent example of this occurred on 6 February 1979 during a 20 day cold spell. Air temperatures, in the range of -35 to -40°C, were punctuated by a few hours of "warm" windy conditions that accompanied the movement of a

short wave trough over Fairbanks. Northerly flow at the 500 mb level had persisted over Alaska for 5 days prior to the passage of the trough; it continued for another week after the momentary relief provided by the trough's passage.

The effects produced on temperature, visibility and carbon monoxide concentrations by passage of the trough are strikingly shown in Figure 5. The carbon monoxide levels abruptly fell to half of their previous value and remained low until the next morning. Simultaneously, the visibility increased and remained greater than 6 km until morning. The temperature did not increase at the airport which lies at the valley bottom, but reports of increased temperatures were common from higher levels. This is consistent with the extremely strong surface inversion which existed at the time the trough passed over the area (Figure 1, curve 2). The progression of the trough across Alaska is shown in Figure 6.

Immediately following passage of the trough, a surface high pressure ridge, rotating clockwise, extended from the northern Bering Sea eastward into Alaska on 2 pm Tuesday, 6 February and the Fairbanks sounding at this time (Figure 1) exhibited a typical strong winter inversion. Low pressure persisted in the Gulf of Alaska and was strengthened by movement of the short wave trough into the gulf. Strong gradient flow associated with the low pressure system in the gulf existed immediately south of Fairbanks. On Tuesday afternoon surface pressure began to rise rapidly (2.4 millibars in 3 hours) because of the approaching high pressure center from the north.

By 2 am on Wednesday morning the sounding revealed the atmospheric mechanisms responsible for the sudden changes shown in Figure 5. The high pressure system moved southward into the interior, and strengthened the gradient winds over Fairbanks (Figure 1) because of the nearly

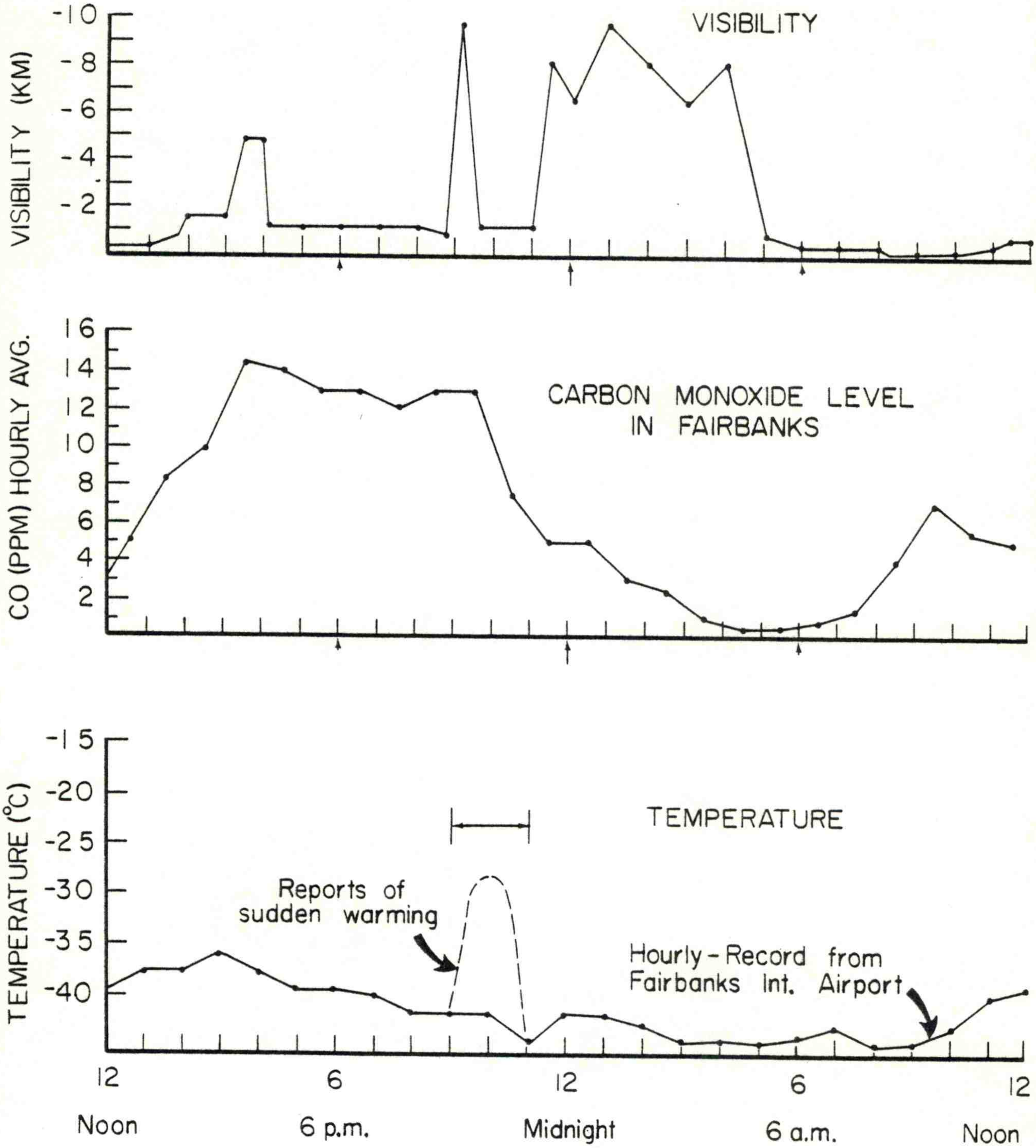


Figure 5. Changes in visibility, carbon monoxide content and air temperature produced by the passage of a short wave trough over Fairbanks. Temperature at the airport did not rise because the measuring site remained under the strong inversion (see Figure 1) but many residents on hill slopes reported the warming as shown.

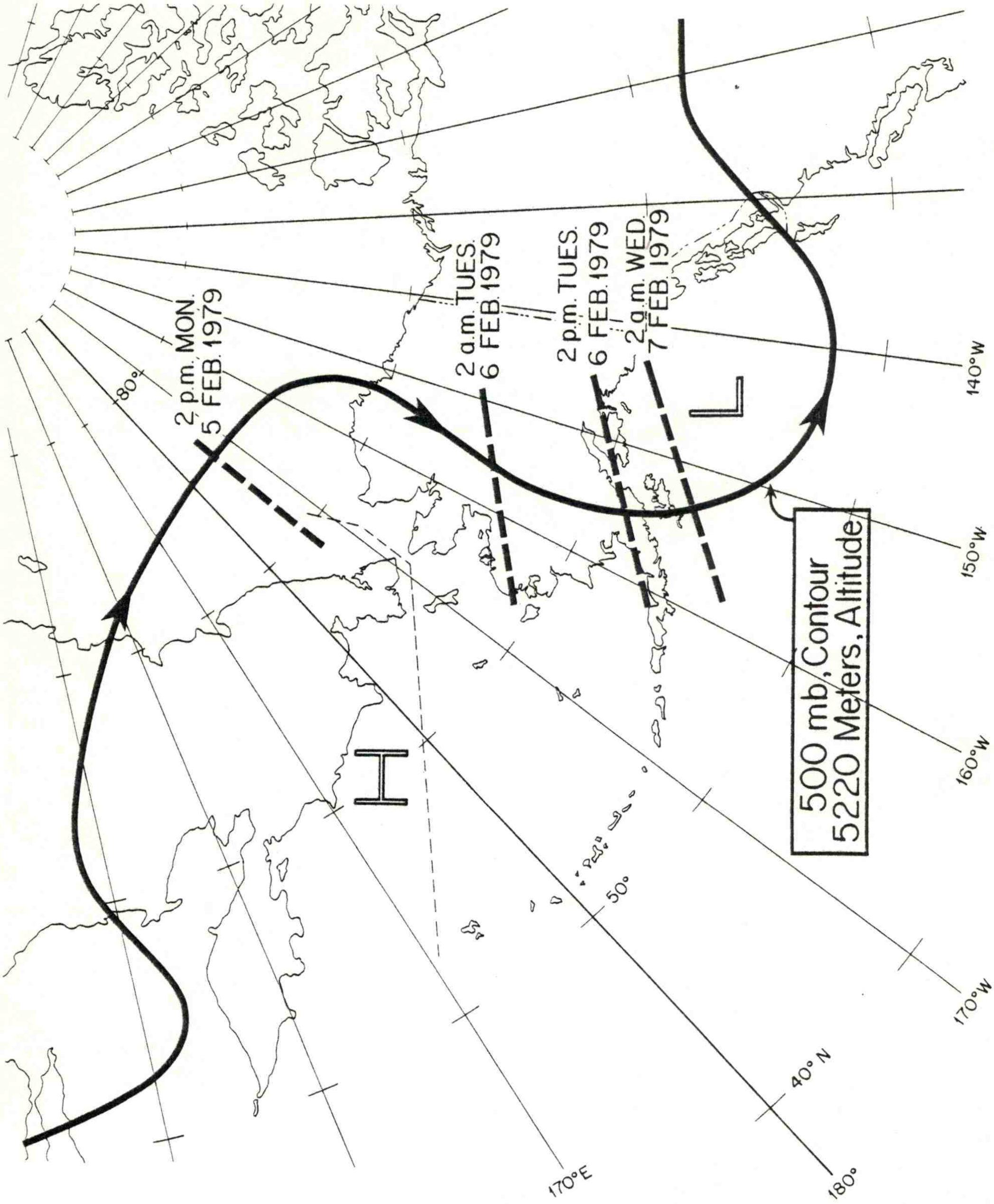


Figure 6. The general flow of cold air over Fairbanks during February 1979 is illustrated by this contour of the 500 millibar level. The location of the short wave trough which moved across Alaska in 24 hours is indicated by the dated, dashed lines.

stationary low pressure system in the Gulf of Alaska. Initially the stronger winds above the inversion increased turbulent mixing. This mixing of warmer air to lower levels slowly eroded the top of the inversion, thus producing the extremely strong shallow inversion shown in curve 2 of Figure 1. The surface inversion increased in strength from 4.8°C/100 m to 21°C/100 meters and it was only 72 meters thick!

Due to the stronger gradient wind areas above the valley bottom experienced enough turbulent mixing to briefly break the inversion. This brief period was long enough to flush away the high concentrations of carbon monoxide and ice fog. The inversion quickly re-established itself, but visibility remained above 6 km and carbon monoxide level below 5 ppm until morning traffic began.

Life Style

The exceptionally high concentrations of pollutants in the Fairbanks area are due to the combined effects of extreme air stability and the attempt to force wasteful, temperate-latitude U.S. life styles into the sub-Arctic. We put an unusually large amount of pollution into an exceptionally stable air mass.

Compared to other cities, Flohn (1971) found that during the period 1955-1965 the energy consumption per capita in the small town of Fairbanks was among the highest in the world. This has not decreased; instead, the commercial electricity used during the coldest months in the Fairbanks area increased from 8 to 15 million Kilowatt hours (KWHr) between 1957 and 1967; and by 1976 it had grown to 53 million KWHr. Also the amount of gasoline burned quadrupled between 1967 and 1976. The heat released to the Fairbanks atmosphere per capita in 1977 was about 10 KW per person

in winter and 5 KW per person in summer; the winter value is comparable to that of the heavily urbanized strip along the east coast of the U.S. (Bowling and Benson, 1978).

In addition to the high per capita use of energy in winter, the mode of operating vehicles at low temperatures, in fog with restricted visibility, and on crowded, snow and ice covered streets plays a role in itself. Automobiles are started when cold, and according to Leonard (1975), a "cold start" releases an order of magnitude more CO than a "warm start." The slow speed, stop-and-go style of driving on icy streets in the fog also aggrevates the problem, because automobile engines are operating in their least efficient mode. To add insult to injury, thoughtless people leave their cars idling unnecessarily for extended periods while shopping. As temperature decreases in the Fairbanks area, the need for increased heat and power, combined with decreased efficiency of vehicle use, increases all sources of pollution. Unfortunately, it is in these times that the stability of the air mass becomes most extreme. Thus, natural and man-made factors reinforce one another in ways which invariably lead to intensification, never mitigation, of the air pollution problem in the Fairbanks air shed.

Conclusions

The Fairbanks type of low temperature air pollution also occurs in Whitehorse, Yukon Territory and in the areas of increasing development of Canadian oil sands in northern Alberta. Another type of air pollution exists in the fjords of southeastern Alaska as in Glacier Bay National Monument (see article by Wendler and Eaton in this volume). Air pollution has even surfaced on the arctic coast of Alaska with the petroleum development

at Prudhoe Bay. One of the most striking examples of this is the clear presence of exhaust plumes extending for 30 km or more across the arctic tundra and adjacent Arctic Ocean, which can be seen on satellite images (Figure 7). The fact that air pollution problems are recognized in Alaska and other arctic and subarctic regions is another reminder that industrial society interacts significantly with the overall environment of our planet Earth.

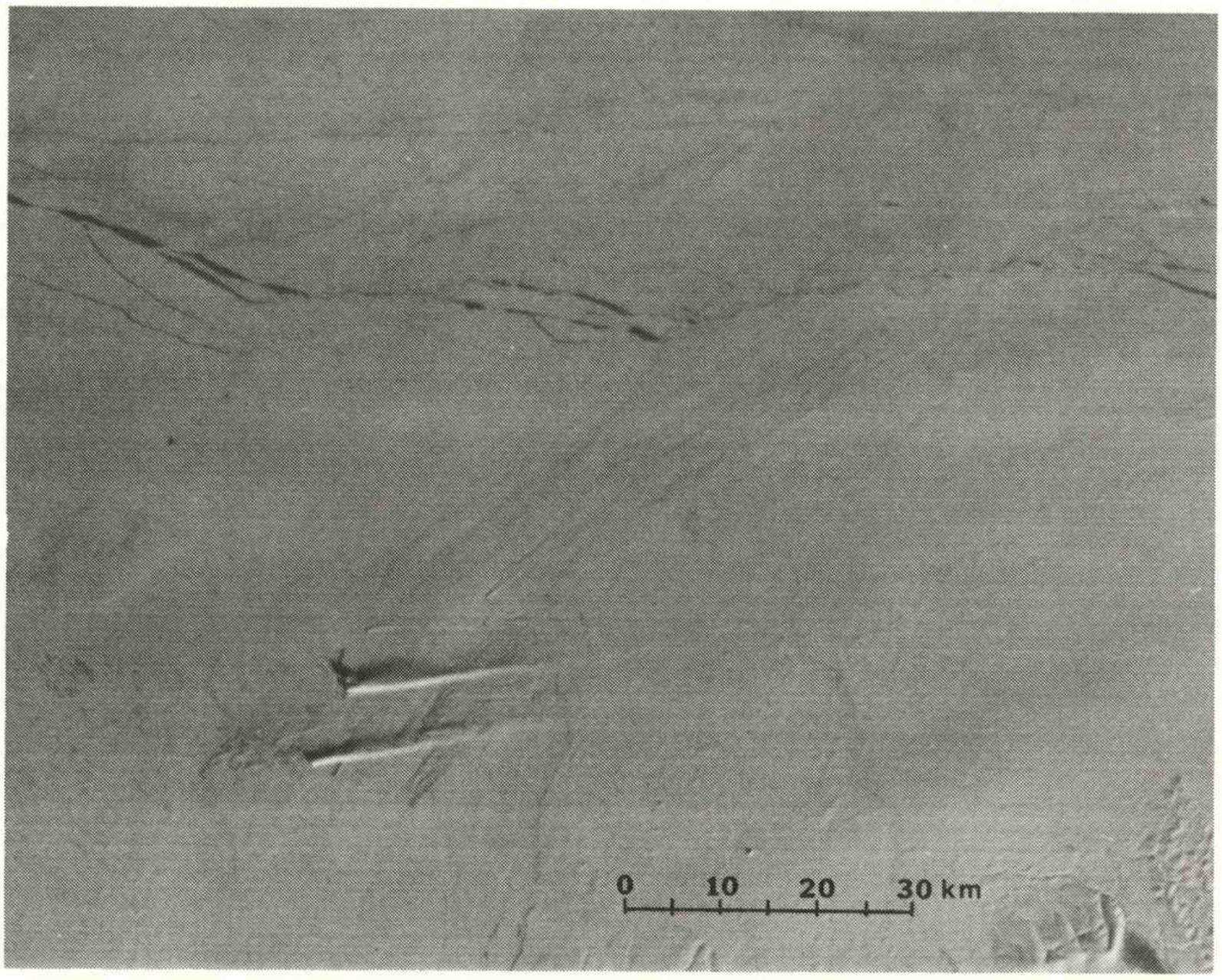


Figure 7. The impact of man of the Arctic atmosphere as seen from space. LANDSAT Satellite Image No. 2752-20505, 12 February 1977 over Prudhoe Bay, Alaska. The two large condensation plumes can be traced for 20 to 30 km across the snow covered tundra and adjacent Arctic Ocean. Many smaller plumes can be seen trending at in a different direction at lower altitudes. The shoreline is not clearly visible because during winter there is little difference between snow covered sea ice and tundra. Open leads and large cracks in the sea ice are visible near the top of the picture.

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ARCTIC HAZE

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Almost invariably those coming to the Arctic for the first time are impressed enough with the great clarity of the polar air to comment on it in journals--and sometimes are fooled enough by it to end up travelling further than they would imagine to reach objects that "appear" to be relatively near by. Indeed, on a fine summer day in the Arctic one can perhaps not see quite forever, but one can certainly see mountains and landmarks two or three hundred kilometers away. Arctic air can and has been described as being crystal clear--at least in summer.

In the winter months, surprisingly, the situation changes and the northern polar air becomes filled with a pervasive haze (Fig. 1). What is this haze and where does it come from?

A Brief History of Arctic Haze

Arctic haze was first described in 1956 by J. M. Mitchell; he used slant-path visual range determinations from B-29 aircraft flying on USAF "Ptarmigan" weather reconnaissance missions. Mitchell said that the haze could be seen even at latitudes as high as 85°N. From its color he surmised it was composed of submicron-sized particles and he rejected the hypothesis that it was caused by ice crystals since no halo phenomena were observed.



Figure 1. Arctic haze as photographed near Barrow, Alaska in April, 1978.

In the early 1970's turbidity* at several arctic stations was found to be higher than expected and to have a seasonal variation opposite to that at mid latitudes (Shaw and Wendler, 1972; Holmgren, 1971). This was entirely unexpected and puzzling. Subsequent investigations from 1973-1977 confirmed that the haze was strongest in late winter/early spring.

More detailed investigations made with a research aircraft at Barrow showed that the haze concentration usually increased from the surface and reached maximum concentration at several thousand meters altitude (Fig. 2). From this, we were able to infer that the haze is not produced from nearby surface sources. The altitude of the haze, its horizontal extent covering hundreds of kilometers, and its association with air masses coming from the north have made it difficult to understand exactly where the haze originates. Until recently all that could be said was that the particles were either being produced *in situ*, perhaps from trace gases, or that the particles had been carried in from very distant sources.

In 1976 studies of arctic haze were supplemented by a series of chemical sampling experiments carried out by Kenneth Rahn and Randolph Borys at the University of Rhode Island. These experiments were aimed at determining the composition of arctic haze and the results are quite surprising. They tell us that arctic haze is rich in those elements associated with industrial pollution such as vanadium and manganese. This was the first of several indications we now have that arctic haze is not natural in origin but is associated with industrial air pollution. Just where this pollution comes from is not yet known with certainty, but we think its source is likely to be Europe and, at times, the northeast United States.

*Turbidity is a term used to express the haziness of the atmosphere. It is usually expressed in terms of how much sunlight is lost on its way through the atmosphere when no clouds are present.

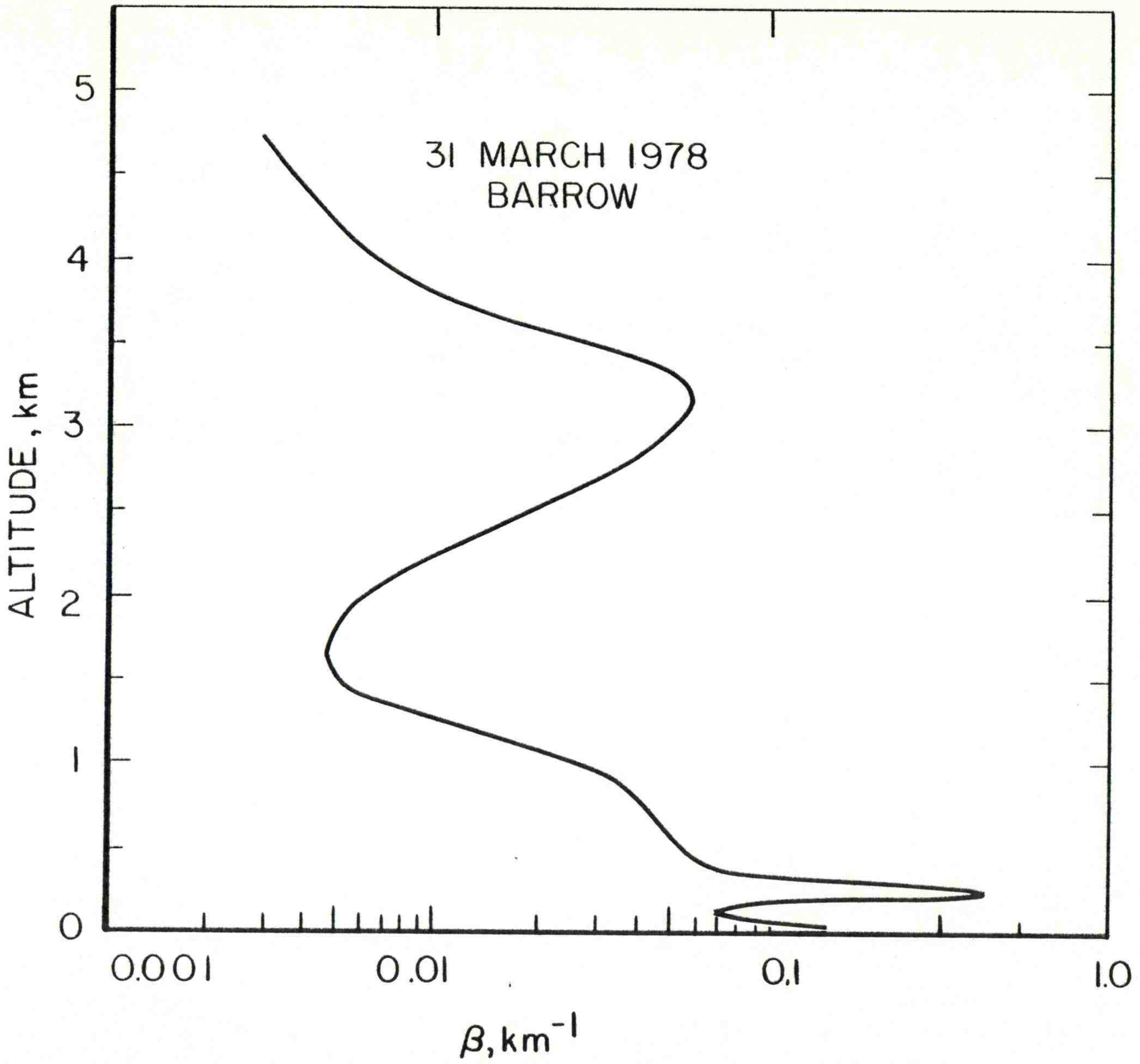


Figure 2. The vertical profile of arctic haze over northern Alaska. This figure represents the profile of optical extinction; it is more or less proportional to the particle mass concentration.

The Summer-Winter Contrast

The presence of haze in the arctic atmosphere during winter is as surprising as its absence in the summer. Why is it that there should be a seasonal variation exactly opposite from that found at the mid latitudes?

During summer, at least near the coastal regions of the Arctic, one can imagine a suite of local particle sources: organic vapors, combustion products from wildfires, windblown material from loess deposites, particles from on the arctic seas, etc. These sources disappear in the winter and, in addition, air convection currents also disappear or are greatly reduced in strength in winter. That is why it is so surprising to find a decrease, rather than an increase, in air turbidity as winter changes to summer in the arctic.

Turbidity cannot be measured in the usual way, by measurements of sunlight with a sun photometer, during the polar winter months when the sun is below the horizon. When the sun is up, the turbidity shows a strong spring maximum but it probably is also high in middle winter. We are presently developing a starlight photometer to extend the arctic turbidity measurements into winter. In the meantime, various chemical constituents of aerosols collected in the Arctic, including sodium and aluminum which are derived respectively from ocean and soil, both show a winter peak in their concentrations. The same is true of sulfate and vanadium, as shown in Figure 3.

The winter-summer contrast is apparently Arctic-wide; we've found that sampled particles from the air at Thule, Greenland, at Spitsbergen, and near Fairbanks (representative of interior Alaska), all show pronounced winter maxima. In general it seems one can say that arctic air is clean in summer, but dirty in winter.

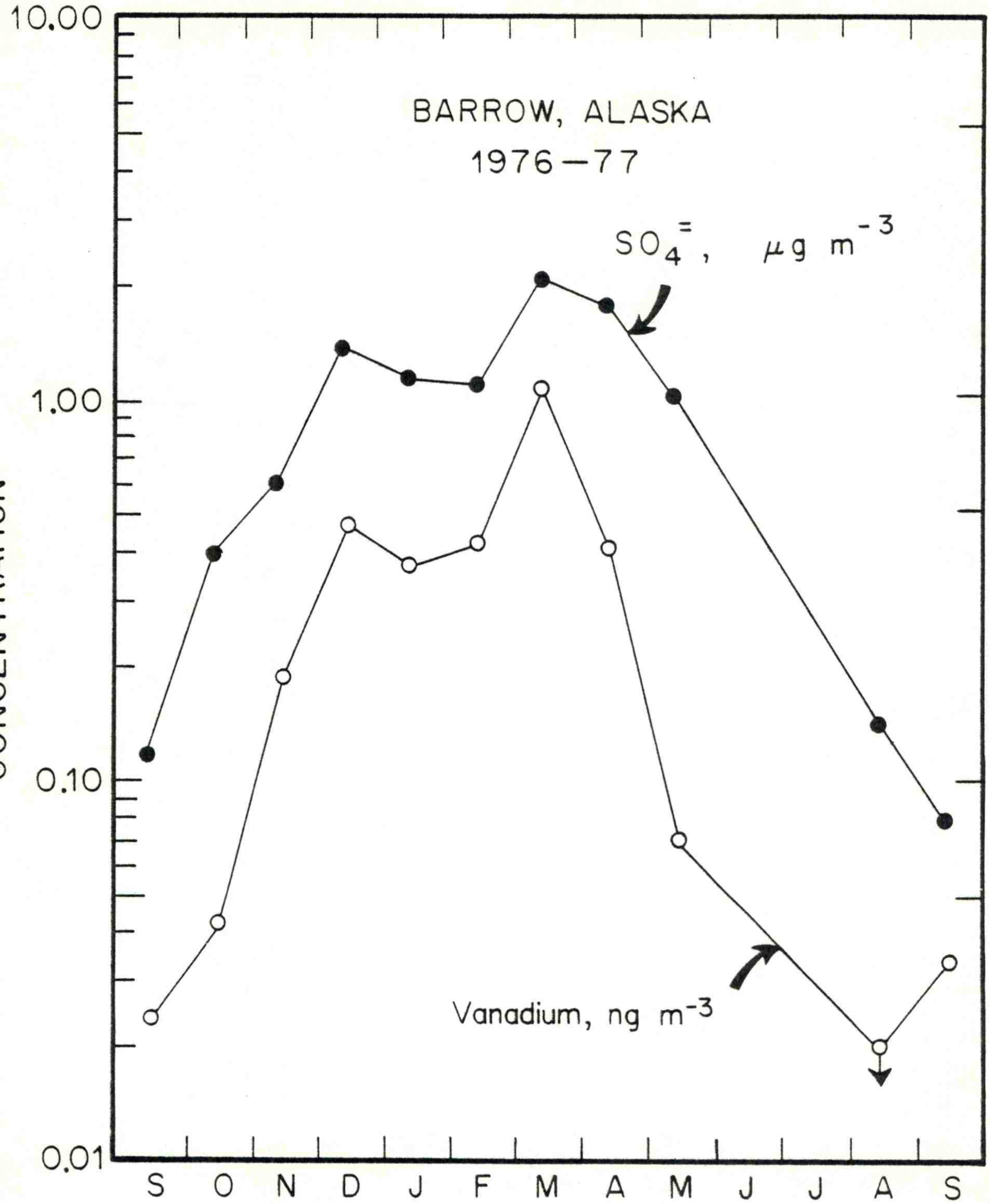


Figure 3. The seasonal variation of sulfate and vanadium in particles collected on filters at Barrow, Alaska. Both $SO_4^=$ and V are generated in large quantities by industrial processes.

The Pollution Component of the Arctic Winter Aerosol

The conclusion that arctic haze is mainly pollution-derived can be supported by studies of individual strong haze episodes that occur during winter in relation to the meteorological situation that preceeded the haze build-up. In this way we have identified, at least tentatively, some of what seem to be the main pathways to the Arctic for pollution products (Figure 4). These pathways are tentative inferences that have been based on air trajectories for several individual episodal haze events seen at Barrow and at Spitsbergen.

A strong indication that arctic haze is pollution-derived comes from the high concentrations of pollution-associated elements, like vanadium. A useful way of expressing the relative increase in concentration enrichment (or for that matter depletion) of an element is to compare its concentration with that normally found in the earth's crust. The strong enhancements seen in vanadium and manganese in the winter months implies that arctic haze consists of particles from industrial pollution.

Besides enrichments of the elements like vanadium which indicate a pollution source, another indicator of pollution is the gray coloration of winter filter samples. The color comes from the presence of sooty, unburned carbon, an element likely to have originated from man-made combustion processes. Perhaps the most convincing evidence for a man-made source of the haze is the high concentrations of sulfate in the haze.(fig. 5). Sulfate accounts for more than one-half of the mass of atmospheric particles over Barrow. By considering the transport and evolution of sulfate aerosol Rahn and McAffery (1979) have found that the majority of it is likely to come from conversion of gaseous sulfur dioxide. The major source of this SO_2 is from midlatitude pollution!

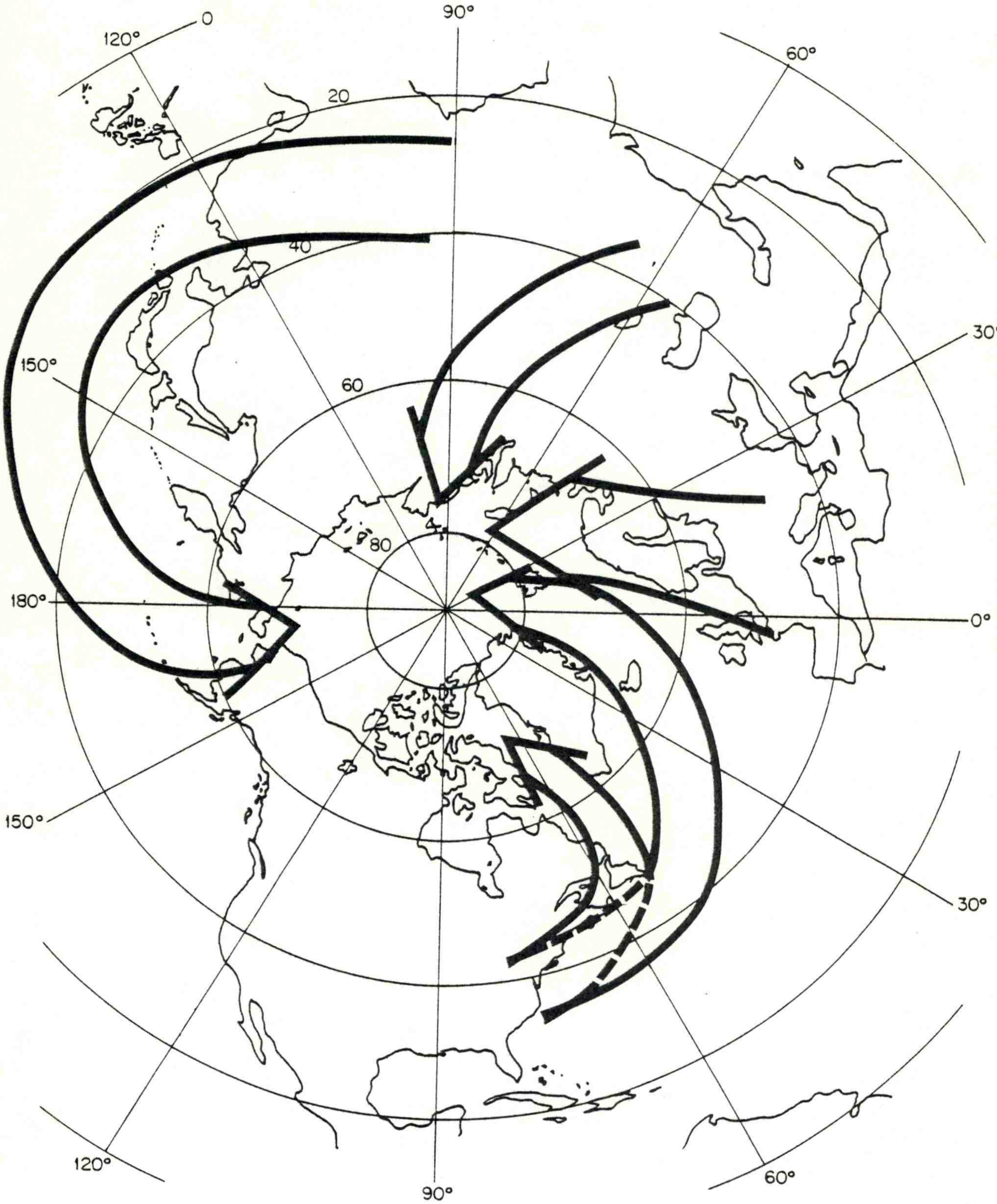


Figure 4. A schematic representation of the principle pathways of industrial pollutants being injected into the Arctic.

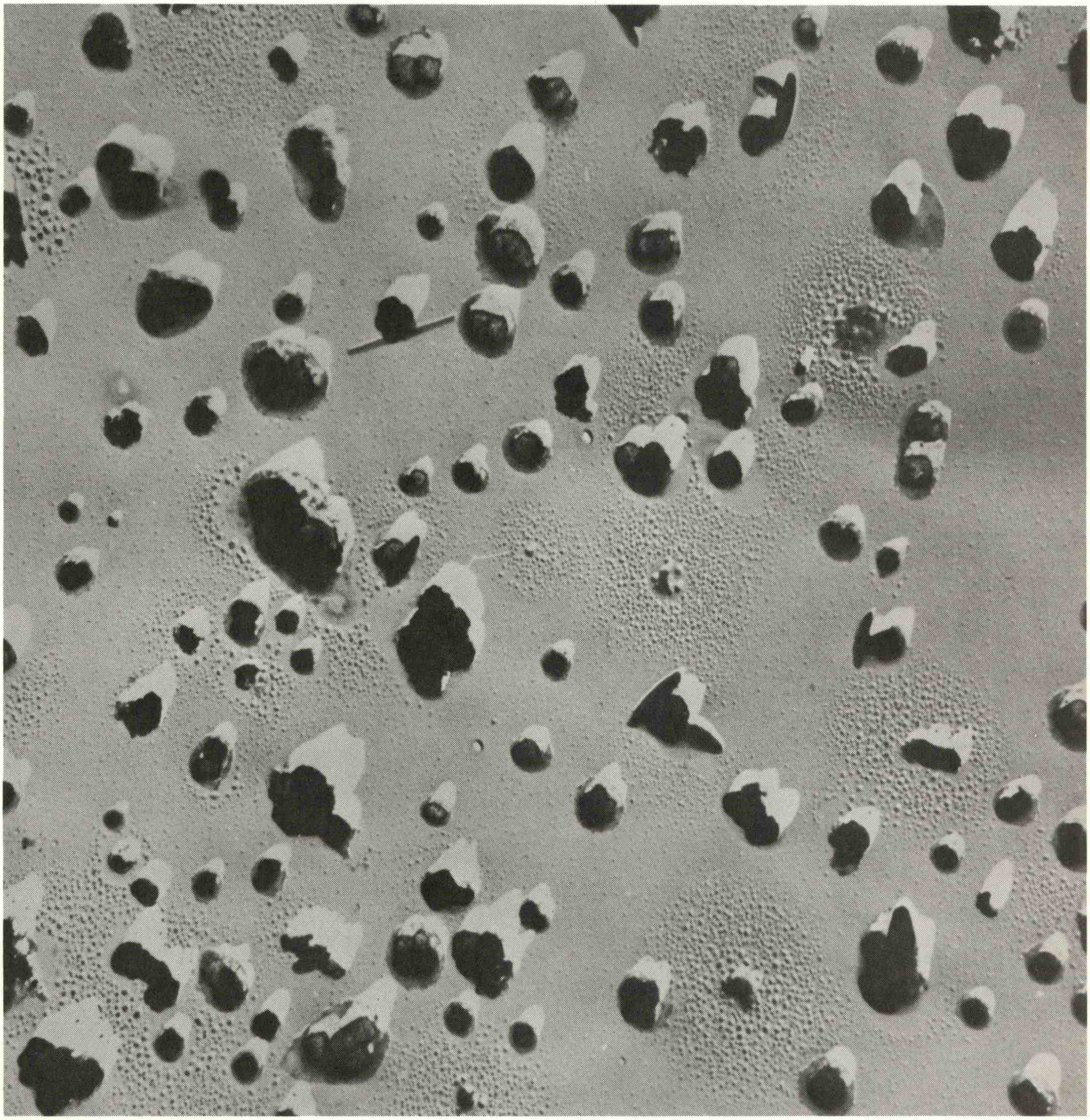


Figure 5. Electron microscope photograph of arctic haze particles collected at Barrow, Alaska. The majority of the particles contain sulfur. The particles have been overcoated with a thin layer of Barium monoxide and the electron beam illuminating the sample has been adjusted at a slant angle to cast a shadow of the particles. Photo courtesy of Dr. E. K. Bigg.

Why is the Arctic Polluted?

The concentration of haze and gases such as SO_2 in the antarctic region is about a tenth of that found in the Arctic. This is not surprising if the hypothesis is true that arctic haze comes from industrial activity, for most industrialization is concentrated in the northern hemisphere. Strong convective storms in the Intertropical Convergence Zone (ITCZ) near the equator probably prevent most northern pollutants from passing into the southern hemisphere.

The arctic regions in general are distant from aerosol sources. Within the northern hemisphere the meridional transport (by mean flow and by eddy diffusion) is considerably less efficient in summer than it is in winter. In summer the Arctic regions are effectively cut off from the lower latitudes due to the highly zonal airflow; this may be part of the reason for the low summer turbidities found at places like Barrow. In summer the arctic air is almost as clean as the antarctic air. In winter the arctic atmosphere becomes turbid. In the Arctic there are three possible reasons for the fact that the haze is most pronounced in winter: (1) increased winter emissions of pollutants, (2) more rapid and efficient poleward transport in winter, and (3) longer residence times of the haze particles in winter. The third reason is based on the fact that clouds and precipitation are quite efficient at removing ("scavenging") pollution particles from the atmosphere and both clouds and precipitation are less prelevent in the Arctic during winter than they are in summer.

It also is probably true that particle removal rates decline substantially in winter due to the high convective stability of arctic air.

This stability limits the strength and number of turbulent eddies. The lesser turbulence means that it takes longer for particles to be removed by being carried close to and impacting or diffusing onto the surface.

Based on analytical models of the transport of material from mid-latitudes to the Arctic it has been learned that arctic haze particles are mainly secondary products, having been formed from the conversion of trace gases, especially by the oxidation of sulfur gases (Rahn and McCaffrey, 1979). The model indicates that the transit time to the Arctic is about 20 days. Thus it appears that arctic haze is caused by particles and gases injected from the lower latitudes, which accumulate for about 3 weeks before being removed. During this time the haze cloud diffuses to a size larger than individual cyclones and anticyclones and a uniform mix eventually covers the entire northern polar regions. An "instantaneous" injection of particles into the arctic, say by northward-flowing wind over Europe, creates an episode lasting about 3 weeks in length.

Significance of Air Pollution Aerosol in the Arctic

The mean size of particles making up arctic haze is around a half micrometer, a size comparable to the wavelength of visible light. Particles like these interact strongly with sunlight and redistribute the radiation fluxes in the atmosphere and at the surface. In addition, radiative interactions occur between the particles and the infrared terrestrial (heat) radiative fluxes in winter, when the sun is down. What, if any, effect does the redistribution of radiant energy have on climate? Unfortunately a complete answer is not yet available.

Preliminary calculations indicate that in springtime the haze introduces slight cooling at the surface and a relatively strong heating within the haze layer itself. In winter, the haze would warm the lower layers of the atmosphere and the surface (Shaw and Stamnes, 1979). The response of weather, and ultimately climate, to such forced cooling and heating is difficult to predict because of complicating positive and negative feedback mechanisms. For example the cooling near the surface and the heating aloft alters the dynamical stability of the atmosphere and introduces air subsidence which could affect cloudiness. The haze may also affect the nucleation properties of the Arctic atmosphere and through this mechanism change cloudiness and precipitation. Very little is known about the extent of these effects.

Arctic haze particles will deposit on the surface, though in relatively low concentrations, and the environmental consequences of deposited acidity, sulfate, nitrate, heavy metals and organic substances on the pack ice and tundra are not yet known.

The most significant result of studies of arctic haze is the recognition that huge areas of the world are being affected by pollution aerosols and gases. On a monthly basis, the Arctic and Euro-Asian sub-arctic are all nearly equally polluted by sulfate aerosol, at least in winter. Clearly arctic haze is a multi-national, even a multi-continental phenomenon. Future research should clarify many questions about arctic haze that are presently unanswered.

Acknowledgements

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Noctilucent Clouds in Alaska

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It's almost midnight in the Alaskan interior on a cool crisp August night. A late evening thundershower has cleared the air and patchy cirrus clouds are the only remnants of a vigorous thunderhead that only hours ago dropped 15 mm of rain and generated 50 km/hr winds. After a beautiful sunset, with cirrus clouds underlit by the waning orange rays of the sun, near darkness has set in. The remaining filaments of cirrus have now become diffuse silhouettes against the background of the twilight sky. A few of the brighter stars appear overhead as the sky gradually darkens. It's almost too dark to see clouds...or is it? A scan across the northern horizon brings dimly-lit clouds into view. High above the ground, they resemble cirrus clouds with delicate formations, and glow with an eerie silvery-blue color. It's now one o'clock in the morning. The clouds have become brighter and cover almost one quarter of the sky. A noctilucent cloud display...one of the most interesting sights in the Alaskan sky, has made its appearance.

Noctilucent clouds are among the most beautiful of all cloud formations in the earth's atmosphere. The highest clouds known to exist in the atmosphere, they form at an average height of 82 km. Noctilucent means "night-luminous". The clouds do not emit light of their own but at their great altitude are often illuminated by sunlight so we can see them against the background of a dark sky when the sun is well below the horizon.

Noctilucent clouds are believed to consist largely of ice particles, formed from water vapor in the earth's atmosphere around nuclei of uncertain origin - probably grains of meteoric dust. The particles are microscopic in size with an average diameter of .00003 cm. Typical noctilucent clouds may contain from 10,000 to 100,000 particles per cubic meter.

Noctilucent clouds are rather thin and tenuous in appearance, often resembling cirrus clouds. Typically they are about 1 to 2 km thick. The clouds are usually bluish white in color, although near the horizon they may appear silvery white with a tinge of yellow or orange.

Usually, noctilucent clouds can easily be distinguished from more ordinary clouds. For one thing, noctilucent clouds always appear brighter than the background sky. Clouds that are darker than the sky background cannot be noctilucent clouds. Also, noctilucent clouds remain visible when the sun is much too far below the horizon to illuminate any clouds visible at lower altitudes, and their silvery or bluish-white color is quite different than the pinks, yellows, and oranges of ordinary cirrus clouds seen soon after sunset or before sunrise.

In order for noctilucent clouds to be visible, the sun must be between 6° and 16° below the horizon. The angle formed by imaginary lines connecting the sun, the observer, and the horizon is known as the solar depression angle when the sun is below the horizon. When the solar depression angle is less than 6° the background sky will be too bright for noctilucent clouds to be easily visible. Very bright noctilucent clouds can occasionally be seen at solar depression angles between 2° and 6°, but under such circumstances it is difficult to distinguish them from ordinary cirrus clouds. When the solar depression angle is more than 16°, the very high atmospheric level at which noctilucent clouds form will lie in the earth's shadow where no sunlight can reach them, and like all lower clouds they become invisible.

In the northern hemisphere, noctilucent clouds have been sighted as far south as latitude 45°N and as far north as latitude 80°N. However, the clouds are most frequently observed between latitudes 50°N and 65°N with the optimum being 60°N. While noctilucent clouds are quite commonly seen every summer in most parts of Alaska, they are seldom seen in the "lower" 48 states. For the most part noctilucent clouds are seen between mid-June and mid-August and are especially often seen in mid-July. An observer at latitude 60°N might expect to see noctilucent clouds on as many as three of four nights during July. A few sightings of noctilucent clouds have been reported in the southern hemisphere during the summer months in that hemisphere.

At latitude 65°N, approximately the latitude of Fairbanks, the best month for observing noctilucent clouds is August. From about mid-May through the third week in July noctilucent clouds cannot be observed at this latitude since the sun does not drop far enough below the horizon, rendering the background sky too bright. Farther south in the state, at latitude 60°N, it is possible to observe noctilucent clouds during the entire summer. Unhappily, during much of the summer in Alaska overcast skies intervene to block noctilucent clouds from view except from high-flying aircraft. Extended periods of clear weather do occasionally occur in summer, however, when the clouds are readily visible.

The duration of noctilucent cloud displays can range from a few minutes to more than five hours. The lifetime of individual parts of a display may often vary from a few minutes to perhaps a half hour. While noctilucent clouds can often be first seen before midnight, the brightest and most extensive displays are observed after midnight. Using binoculars, one can often detect the presence of noctilucent clouds about a half hour before they become visible to the unaided eye. The area covered by noctilucent cloud displays can range from about 10,000 square kilometers to more than 4,000,000 square kilometers.

Noctilucent clouds take on a variety of forms: veils, bands, billows, and whirls. The veil form is the one most often seen; it is generally tenuous in nature and frequently serves as a background for the other forms. The veil form often precedes the appearance of noctilucent clouds having well-defined structure. Noctilucent cloud bands tend to be persistent and may be as long as several hundred kilometers. Edges of the bands may be either diffuse or well-defined. Frequently, bands may appear interwoven at small angles. Sometimes small streaks with twists or bends may lie across the bands or branch out from them, giving a featherlike appearance. Very closely spaced and roughly parallel streaks, called serrations, can occasionally be seen in the veil background, often appearing to the unaided eye as an almost continuous cloud mass. Billows are closely-spaced, roughly parallel short bands, sometimes exhibiting a wavelike structure with undulations. The life span of billows is considerably shorter than that of the long bands; cloud elements may appear and disappear or change shape within a few minutes. Whirls are sometimes seen in the band, veil, and billow formations. The presence of whirls is indicative of considerable turbulence. The band and billow structures are indicative of wave motion in the upper atmosphere.

The author has observed noctilucent clouds at Fairbanks, Alaska, several times each summer during the last five years. All of the sightings occurred during the month of August. Two of the more interesting displays occurred on the nights of August 6-7, 1976 and August 6-7, 1977 (see Figures 1-4).



Figure 1. Noctilucent clouds seen from Fairbanks, Alaska. August 7, 1976, 1:30 am Alaska Daylight Time (ADT). The camera is facing ENE. Solar depression angle 8.3°. Photo taken with 50mm lens, exposure 1 second at f/1.8 on Kodachrome 64 film. The angular field of view seen in the picture is 39° x 26°. Photo by Ken Rizzo, National Weather Service Forecast Office, Fairbanks, Alaska.



Figure 2. Noctilucent clouds seen from Ester Dome, Alaska (16 km west of Fairbanks). August 7, 1977, 1:24 am ADT. The camera is facing N. Solar depression angle 8.3°. Photo taken with 28 mm lens, exposure 1 second at f/2.8 on Ektachrome 200 film. The angular field of view seen in the picture is 42° x 63°. Photo by author.

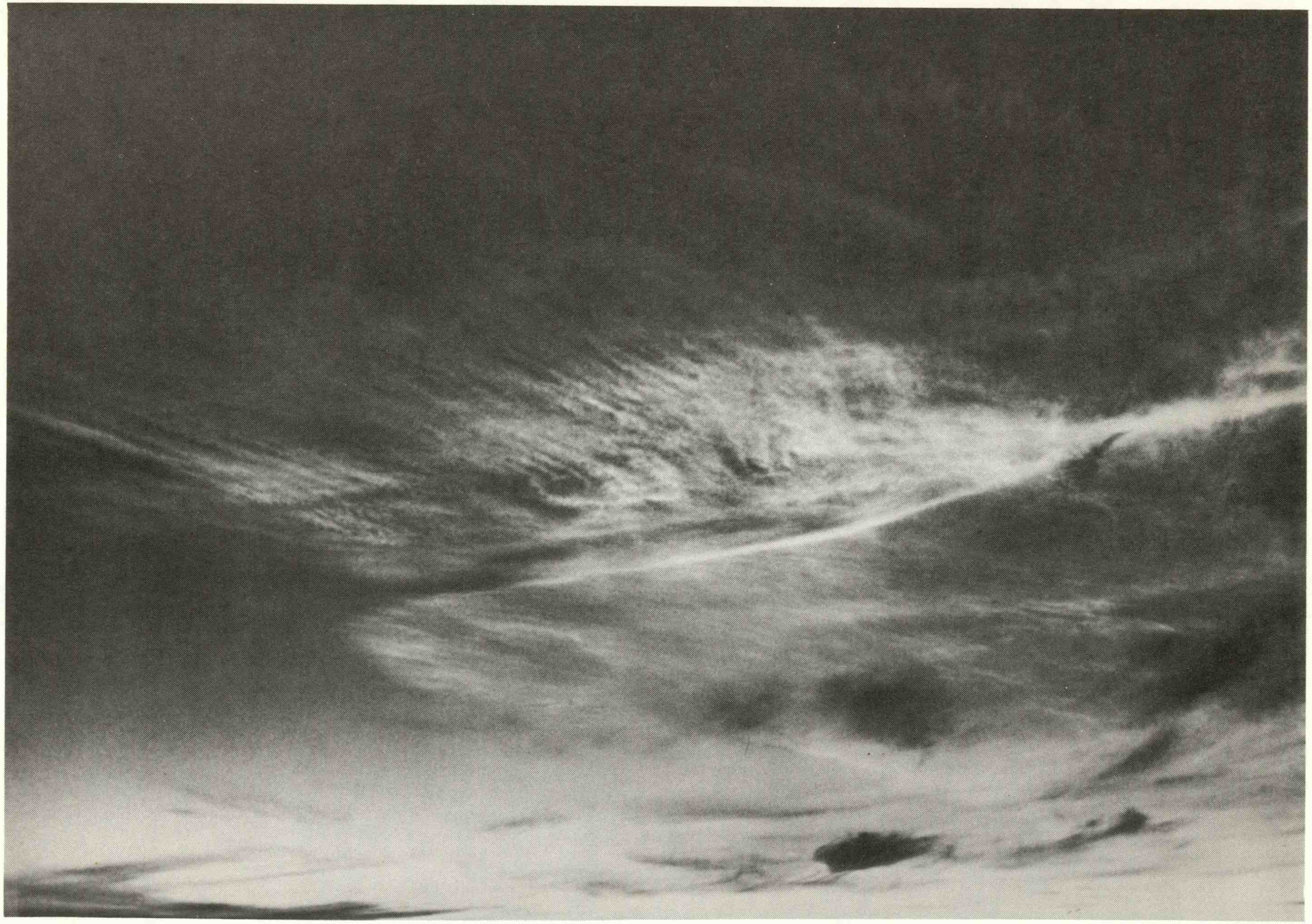


Figure 3. Noctilucent clouds seen from Ester Dome, Alaska. August 7, 1977, 12:53 am ADT. The camera is facing NNW. Solar depression angle 8.6°. Photo taken with 28 mm lens, exposure 1 second at f/2.8 on Ektachrome 200 film. The angular field of view seen in the picture is 42° x 63°. Photo by the author.



Figure 4. Noctilucent cloud band seen from Ester Dome, Alaska. August 7, 1977, 1:01 am ADT. The camera is facing NNW. Solar dpression angle 8.5°. Photo taken with 80 mm lens, exposure 2 second at f/4.5 on Ektachrome 200 film. The angular field of view seen in the picture is 17° x 25°. Photo by the author.

Both displays were visible for almost 4 hours. In Figure 1, billow and veil forms can be seen. The center of the prominent billow formation is 17-18° above the horizon. The distance between adjacent billows is estimated to be about 8 km. The prominent cloud form in Figure 2 and 3 is the veil form. A finely serrated pattern can be seen in Figure 2 just to the right of center. A single thin band can be seen in Figure 3 just below the center. This particular feature became more well-defined several minutes afterward (Figure 4) and had a lifetime of about 30 minutes in all. Cirrus clouds can be seen in the lower half of Figures 2 and 3; note that they appear darker than the background since they are no longer illuminated by the sun.

Photographing noctilucent clouds is not difficult. For relatively bright clouds, exposure times of 1 to 4 seconds at f/2.8 with Ektachrome 200 film should generally provide good results. Dimmer clouds will require longer exposures. Due to the slow shutter speeds required, use of a steady tripod and a cable release is necessary. Many films may require longer exposure than indicated by a light meter due to a property known as reciprocity failure. To be sure of obtaining good results, several photographs can be taken, each with a different exposure to bracket the correct one.

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THE AURORA

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The sun sends out a vast amount of energy in a variety of forms, including: (i) black body radiation, (ii) the extreme ultraviolet (EUV) and X-rays, (iii) the solar wind particles and (iv) subcosmic rays. Traditionally, the atmospheric sciences deal with effects of (i) and solar-terrestrial physics (including aeronomy) deals with (ii) the very short wave radiations, (iii) the solar plasma and (iv) very high energy solitary particles.

The aurora indicates, among other things, the location where a part of the solar wind energy (iii) is finally deposited after undergoing several conversion processes. Figure 1 shows a color photograph of an active aurora. Note that the aurora has a thin curtain-like form with folds. In the first part of this article, the present understanding of the cause of the aurora is described. In the second part, however, we shall refer to possible effects of (ii), (iii) and (iv) on the mesosphere and below.

The earth and its magnetic field are always embedded in the streaming solar corona called the solar wind, which is a fully ionized gas. The earth's magnetic field is an obstacle in the solar wind and thus the latter tends to flow around the former. As a result a cavity is formed around the earth. This cavity, called the magnetosphere, has a comet-like outer shape. The solar wind and the magnetosphere constitute a generator, called the solar wind-magnetosphere generator. The current



Figure 1. A color photograph of the aurora taken from Fairbanks, Alaska (Photo credit: Lee Snyder)

is generated on the magnetopause; see Figure 2. The aurora is an electrical discharge phenomenon surrounding the earth, powered by this natural generator.

The power of this generator is of order 10^{12} W. The voltage generated by the dynamo is estimated to be of the order of 50 kilovolts, and thus the total current is of the order of 2×10^7 amperes. A small part of the current (about $2 - 4 \times 10^6$ amperes) generated by the solar wind-magnetosphere generator flows along magnetic field lines, from the morning side boundary of the magnetosphere (called the magnetopause) to the morning upper atmosphere and flows back from the evening upper atmosphere to the evening side magnetopause. Figure 2 illustrates this portion of the current which may be called the auroral current, since it is this part of the circuit which causes the aurora in the polar upper atmosphere. In fact, the upper atmospheric particle density is similar to that in a typical discharge tube.

The current-carrying electrons ionize and excite upper atmospheric molecules and atoms as they collide with them. The auroral radiations are emitted from these atoms and molecules. The most common emission, the greenish-white color, comes from oxygen atoms, since they are one of the most abundant constituents in the lower ionosphere. The wave length of this particular emission is 5577 Å. A very dark red light, which can be seen in mid-latitudes during great magnetic storms, comes also from atomic oxygen. Other emissions come from molecular nitrogen and other constituents.

If one could look down on the earth from well above the north polar region, a faint luminous annular ring around the magnetic pole would be

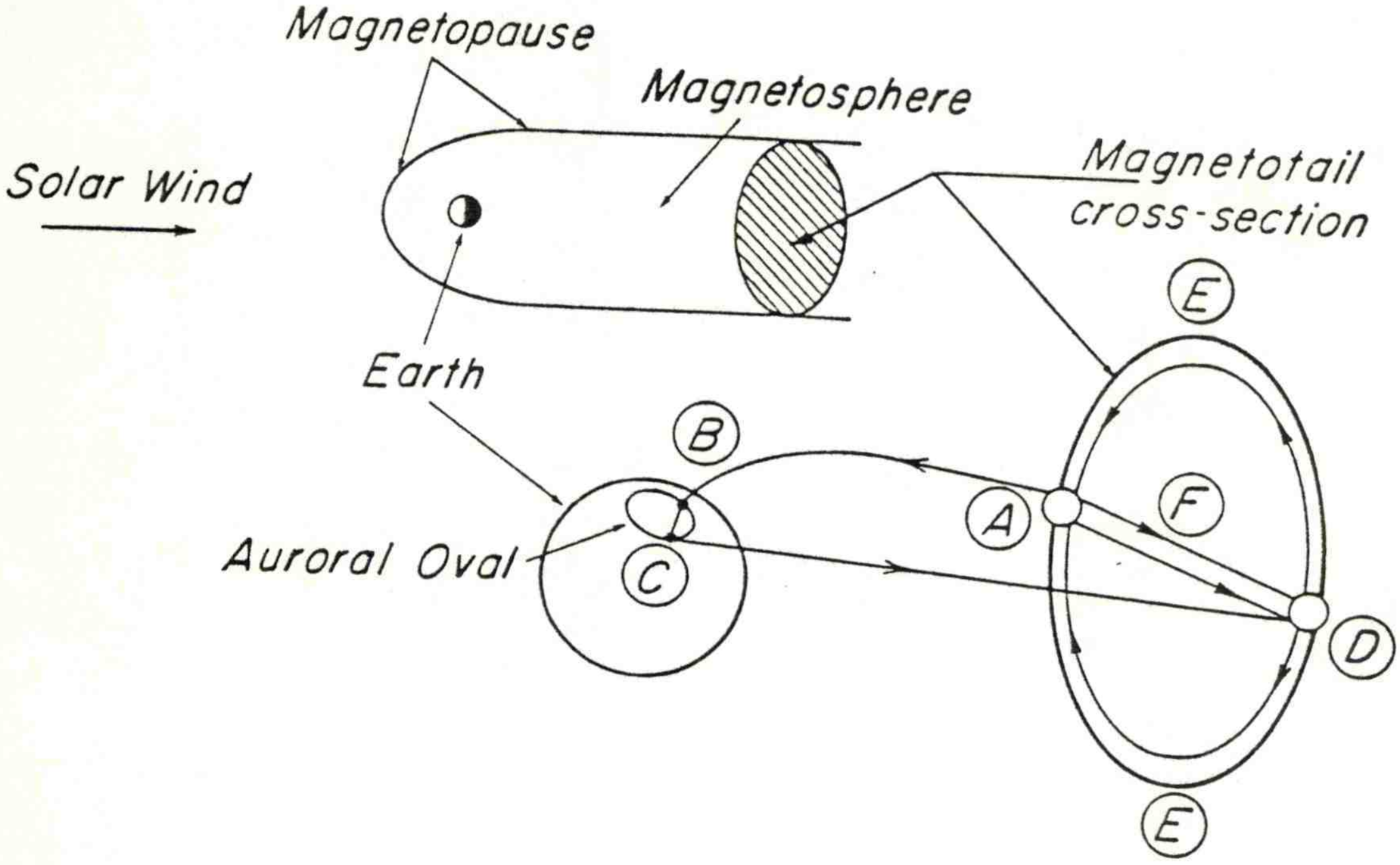


Figure 2. The solar wind and the magnetosphere constitute a generator. The figure shows where the current is generated in one cross-section of a comet-shaped magnetosphere. The current is generated along (D-E-A), flows from (A) to the morning sector of the polar cap (B-C) and finally flows back to the magneto-sphere boundary (C-D); there will be a similar current loop in the southern hemisphere.

observed. This luminous belt is called the northern auroral oval, and there is a similar one surrounding the southern 'magnetic pole'. It is in this belt that one can see the aurora, namely where the discharge currents interact with the upper atmosphere. The aurora has a curtain-like or ribbon-like form, lying in the auroral oval. The thickness of the auroral curtain is of order 1 km, but its length is usually more than a few thousand kilometers. The bottom height of the curtain is about 100 km and the top extends to an altitude of 500 km or greater. This particular shape of the aurora is due to the fact that the discharge current has a very thin sheet form.

A part of the auroral discharge current flows in the ionosphere by connecting the incoming and outgoing field-aligned currents. This portion of the current is called the auroral electrojet, because it is rather concentrated in the vicinity of the aurora. Its current intensity often exceeds 10^6 amperes and causes severe magnetic disturbances (of the order of 1000 $\gamma = 10^{-3}$ gauss) under active auroras.

The direction of the solar wind magnetic field is of great importance in controlling the efficiency of the solar wind-magnetosphere generator. In a generator, the axis of the armature is perpendicular to the magnetic field. However, since the direction of the solar wind magnetic field is highly variable, it may be said that the angle between the armature axis and the magnetic field is variable in the solar wind-magnetosphere generator. As a result, the efficiency of our generator is partially controlled by the varying solar wind magnetic field. The solar wind speed and the magnitude of the magnetic field contribute also to the efficiency. The solar wind energy flux ε coupled to the magnetosphere is found out to be of the form

$$\varepsilon = VB^2 \sin^4 \frac{\theta}{2} \ell_o{}^2$$

where

V = the solar wind speed

B = the magnitude of the solar wind magnetic field

θ = the polar angle of the solar wind magnetic field vector projected onto a plane perpendicular to the sun-earth line.

ℓ_o = constant (= 7 earth radii).

The solar wind-magnetosphere generator attains its maximum efficiency when the solar wind magnetic field is directed southward ($\theta = 180^o$) and its minimum efficiency for $\theta = 0^o$. A high efficiency of the generator produces a larger auroral oval, and brighter and more active auroras. Thus, by monitoring the quantity ε at the front of the magnetosphere, say at the libration point (about 240 earth radii toward the sun), we should be able to predict the occurrence and intensity of auroral displays. Since the International Sun-Earth Explorer C (ISEE/C) satellite is placed at the libration point, it will not be too long before the forecasting of auroral activity will become possible.

When the generator efficiency is greatly increased, the aurora becomes explosively active. This phenomenon is called the auroral substorm. Active auroral displays spread from the midnight part of the oval, and as a result a significant part of the polar night sky can be covered by active auroras. Figure 3 shows a montage photograph of the auroral oval during an intense auroral substorm.

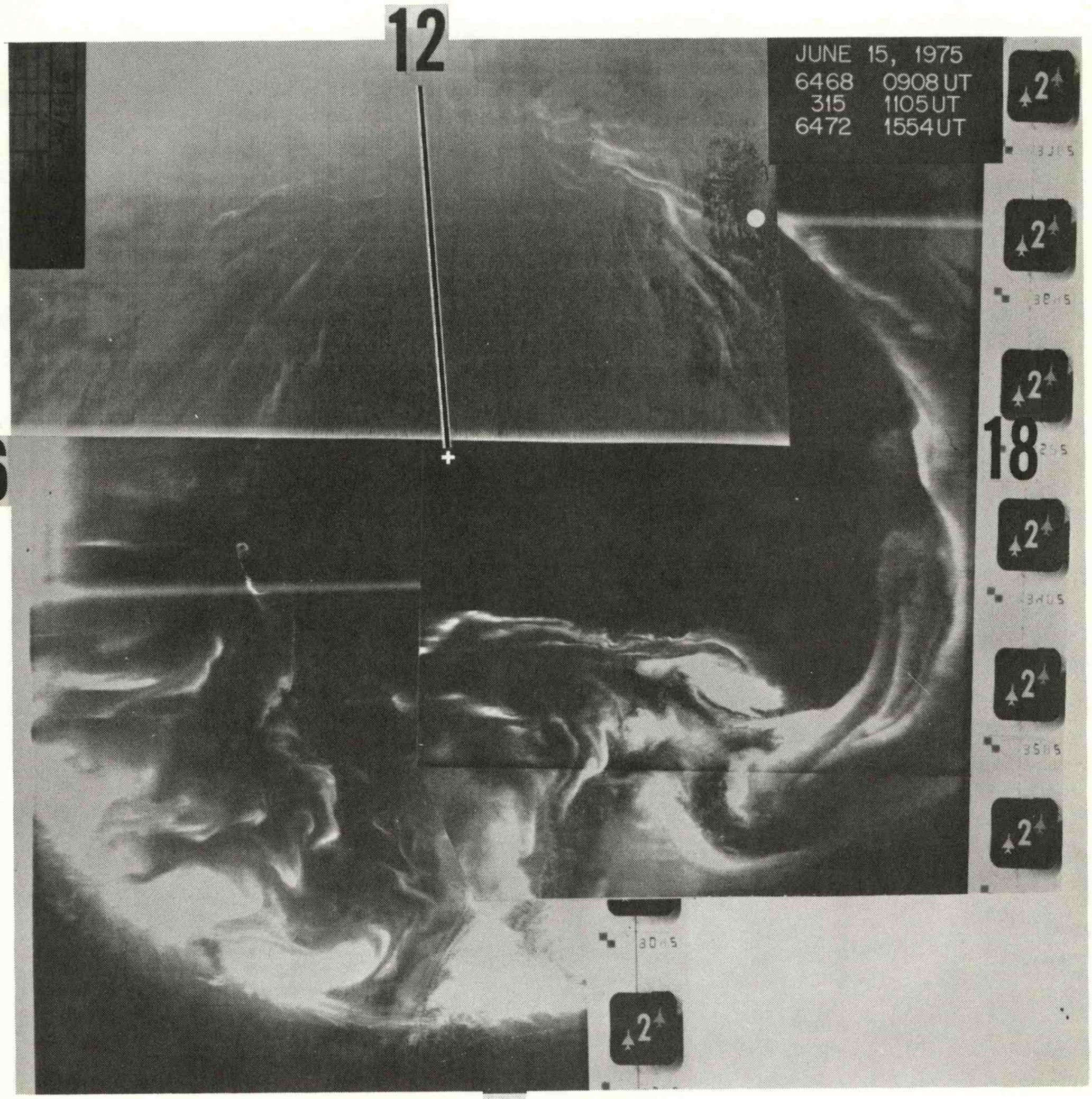


Figure 3. A montage photograph of the auroral oval over Antarctica. An intense auroral activity is in progress, particularly in the dark hemisphere (Defence Meteorological Satellite Program). The (invariant) magnetic pole is indicated by a cross, and the noon meridian by a line from the magnetic pole.

The auroral substorm is, however, only the visual manifestation of what happens in the magnetosphere - the magnetospheric substorm which is one of the major disturbances in the magnetosphere as the efficiency of the solar wind-magnetosphere is increased. Both satellite and ground-based observations are now coordinated to find out plasma processes involved in magnetospheric substorms. One of such efforts is the International Magnetosphere Study (IMS), 1977-1979.

Various solar activities cause the solar wind speed to vary. One example of such solar activity is a solar flare, an intense explosion, which takes place in the vicinity of a complex sunspot group. When the gusty solar wind thus generated 'hits' the magnetosphere, the efficiency varies considerably, and the auroral oval expands unusually equatorward. Such an occasion coincides with the period of a geomagnetic storm, since the discharge currents cause fluctuating magnetic fields. It is on such an occasion that the aurora can be seen even in the southern U.S.

In summary, the magnetosphere may be considered to be a machine which receives the solar wind energy of flux $\varepsilon = VB^2 \sin^4 \frac{\theta}{2} \ell_o^2$ and then converts some of the received energy into heat energy in the upper atmosphere and the rest into particle energies in the Van Allen belt. The aurora can be considered to be the visible indicator of the degree of this solar wind-magnetosphere coupling process.

The auroral effects to the lower thermosphere consist of two processes: (i) Joule heating by the auroral electrojet and (ii) ionization and excitation by impinging auroral particles. Their subsequent effects on the mesophere and below, both confirmed and suggested or speculated, are shown in the block diagram . Note that effects of solar EUV x-rays and cosmic rays (solar protons) are also included(Fig.4).

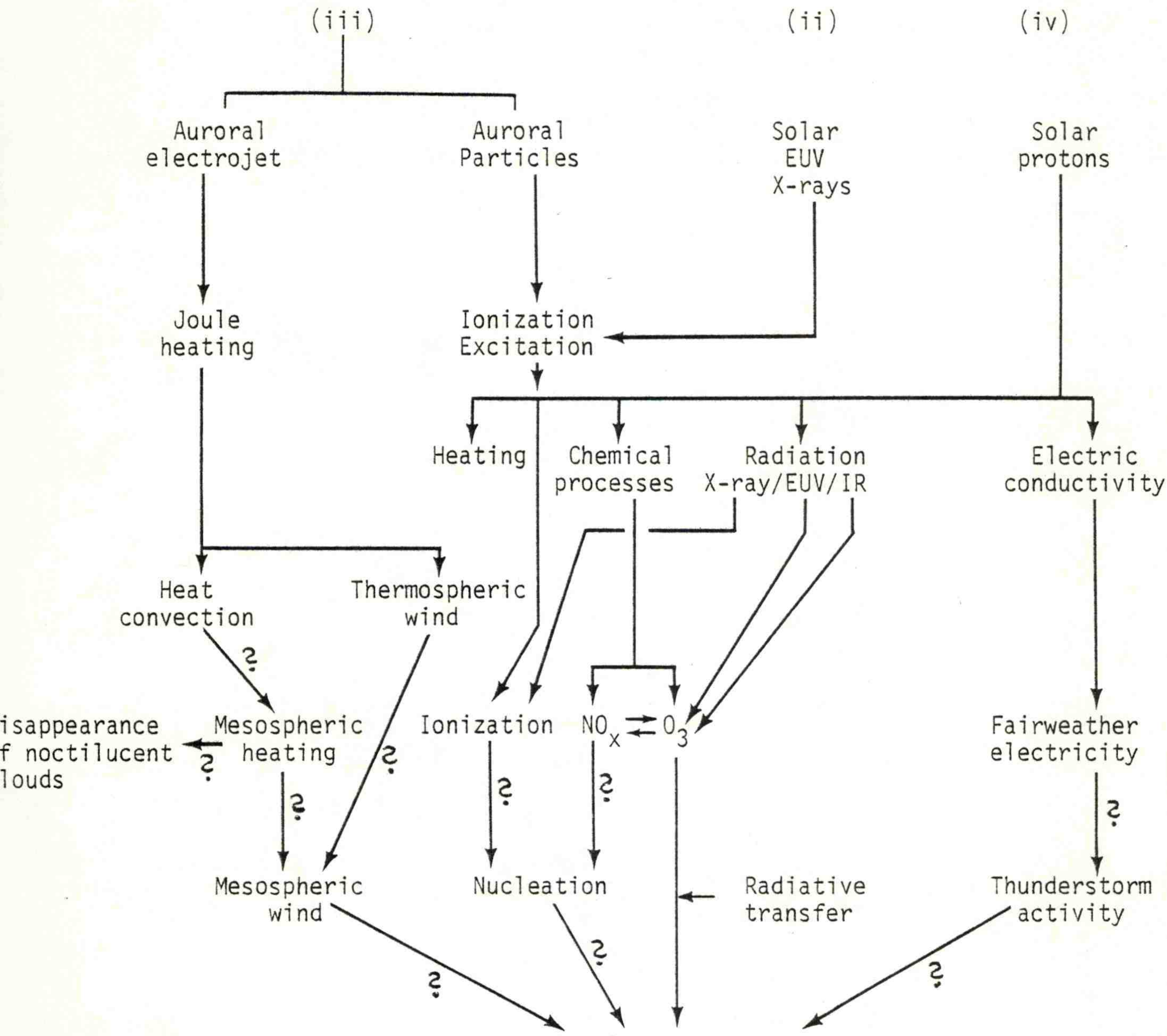


Figure 4. Possible auroral effects on weather (not confirmed)

A few notes may be in order here on auroral effects on the mesosphere. Fogle (1966) noted that noctilucent clouds disappeared during some auroral activity. A new type of radar has just been installed in Fairbanks to study winds in the mesosphere, stratosphere and troposphere (the MST radar) and its potential use in studying the possible dynamic coupling between the thermosphere and the mesosphere is great (Gage and Balsley, 1978). Effects of solar cosmic rays on the mesosphere were most extensively studied by Reid (1974). Holzworth and Mozer (1979) observed a large change of the stratospheric electric field after the great solar flare of August 1972; for modeling a global atmospheric electricity, see Hays and Roble (1979).

There is at present no definitive evidence to confirm any of the suggested specific relationships with the global weather system. It has been shown statistically by Wilcox et al. (1974) that there is a relationship between the so-called 'interplanetary sector boundary crossing' and the 'vorticity-area index'. The 'interplanetary sector boundary crossing' is now found out to be the crossing of the earth between the northern and southern hemisphere of interplanetary space across the wavy solar magnetic equatorial plane. Therefore, as recognized by the authors, the crossing of the earth across the solar equatorial plane itself has no significance in the solar wind-magnetosphere coupling (see also Akasofu, 1975). The crossing days were used only as 'key' days for the superposed epoch analysis. The wavy feature of the solar magnetic equatorial plane is closely related to the distribution of various features on the sun, such as the coronal holes, coronal streamers and active regions. Therefore, it is difficult to distinguish which of the three energy transmission

modes from the sun - (ii), (iii) or (iv) - is actually responsible for the variation of the vorticity area index around the key days.

It may be instructive here to refer to the well-established statistical relationship between the sunspot number and geomagnetic activity. In spite of such a well accepted correlation, there is actually no direct relationship between them. In fact, during the declining epoch of the sunspot cycle, the correlation is negative. This is because sunspot magnetic fields tend to suppress the solar wind flow. The sunspot-free region is called the coronal hole from which an intense solar wind stream flows out. In fact, the correlation becomes positive only when solar activity around sunpots becomes intense enough for the solar wind to blow harder; a solar flare is an example.

Furthermore, in spite of the well-accepted relationship between the solar wind and geomagnetic activity, it is only very recently that the coupling function ε has been found after the last two decades of intensive studies of the magnetosphere and the solar wind. On the other hand, we have not yet uncovered specific plasma processes which couple this solar energy flux ε into the magnetosphere and also which eventually dissipate it in the ionosphere. One can appreciate thus how difficult it would be to identify basic processes involved in possible solar effects on the global weather system.

In dealing with solar-terrestrial phenomena, the most crucial problem we face at the present time is to find out which of the three energy transmission modes (ii), (iii) or (iv) is most important if solar activity affects global weather at all. Without finding the specific energy transmission mode, there is little chance of making headway in

this study. This statement assumes that the black body radiation does not change significantly in a few decades. Most of the present statistical solar-weather relationship studies are not very useful for this particular purpose, and some of them even confuse the issue further.

Here are a few hints for future studies in determining the specific energy transmission mode. For example, when solar flares occur near the western limb of the sun, solar cosmic rays have the easiest access to the earth along the spiral interplanetary magnetic field lines. On the other hand, the associated intense plasma flow tends to 'miss' the earth, so that geomagnetic and auroral activities tend to be relatively weak. By choosing such events, one should be able to distinguish possible effects of (iii) and (iv).

One could perhaps distinguish possible effects of (ii) and (iii) by choosing geomagnetic storms which are caused by solar flares in the eastern hemisphere and are associated with a rather small value of ε. Such storms are characterized by a large storm sudden commencement (ssc), but little main phase and auroral activity. Therefore, possible effects of (iii) are minimum for such events; further solar protons (iv) do not reach the earth for such flares.

A high variability of the global weather system (almost intrinsically indeterministic fluctuations) makes it extremely difficult to correlate global weather changes with specific solar active features. Nevertheless, it is vital to eliminate at least expected changes, on the basis of numerical forecasting techniques (prognostic maps), from the raw data before making the correlation study. The omission of such an effort will only serve to confuse the issue further. We are going to examine,

most likely, whether or not deviations from prognosis maps might be related to particular solar activities, rather than to find relationships which are so obvious that everyone has overlooked them for centuries. Therefore, the task requires also a considerable accuracy of prognosis maps. Such a task would require investigation of individual events with prognosis charts in hand on a daily basis. Alaska is suited for such a daily study, since cyclogenesis in the Gulf of Alaska contributes significantly to the vorticity area index, while auroral activity is extensively monitored along the magnetic meridian through central Alaska.

From the above discussion, it may be clear that statistical studies of the sunspot number and climatic parameters would not tell us specifically the energy transmission mode which is responsible for climatic changes. Fairly systematic geomagnetic data are available in the form of the so-called "C9" index only from 1892 (the C9 index is a measure of the quantity ε in interplanetary space, namely a measure of (iii), but not of either (ii) or (iv)). Records of sighting of auroras in middle latitudes are available for a very extended period ($\sim$ 500 B.C.). Such auroral activity, however, was associated with solar flares, since the auroral oval can expand below geomagnetic latitude 60^{o} only during major geomagnetic storms caused by solar flares. Auroral activity associated with an intense solar wind stream from the coronal hole is limited above geomagnetic latitude 60^{o} and is not seen in middle latitudes, although it can be intense for an extended period, especially during the declining epoch of the sunspot cycle; in fact, it is for this reason that auroral activity continues to be high even during the declining epoch of the sunspot cycle. In any case, it would be a difficult task to find which

of the three energy modes (ii), (iii) or (iv) contributes to climatic changes from auroral data. Some lunar rocks and ice on the Greenland ice cap and in the Antarctic may contain information on a long term change of (iii) and (iv).

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