Polar Regions Snow Cover*

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Abstract

The Earth's surface in polar regions is typified by snow cover, both seasonal and perennial. This snow cover is a sedimentary veneer which is simultaneously a product, and a sensitive indicator, of climate. Several meteorological and geological concepts are combined to study some problems involved in the origin, or deposition, of the snow, and post-depositional, or diagenetic, changes which occur in it. The similarities of the North and South Polar Regions are stressed with specific causes sought to explain their differences. Comparative studies between and within the Polar Regions are useful and improvement in our understanding of present conditions along the lines discussed here should help us visualize Pleistocene conditions.

Air masses transport dissolved water vapor which, under proper conditions, is precipitated and deposited as snow. At this stage snow is an aeolian sediment and surface features of wind erosion and deposition such as sastrugi, barchans, dunes and ripple marks are formed during and after deposition. Stratification results from variations in the conditions of deposition and diagenesis. Critical diagenetic variables are rate of accumulation, mean annual temperature, range of temperature, steepness of temperature gradients in the snow, and wind action. The diagenetic environment extends downward for an unspecified depth, but always includes the entire seasonal snow cover and most of the sedimentary veneer of snow on glaciers. Because the ice grains comprising snow are generally within 50°C of their melting point, and because the air within the interstices is saturated with water vapor, diagenetic processes proceed at very rapid rates.

Certain parts of the earth represent laboratories for specific studies—Greenland is perhaps the finest example. Alaska provides an excellent laboratory for the study of seasonal snow cover because maritime, continental and arctic zones are in close proximity and differences in the snow cover from one zone to the next are striking.

Among physical measurements made on snow strata, *i.e.*, temperature, density, grain size, hardness, etc., stable isotope ratios are of special interest because they vary: seasonally, from one climatic zone to the next, and within a zone as a function of distance along an individual storm track. In zones which receive precipitation from two major sources, isotope measurements may allow determination of the amounts from each. Whether or not melt occurs, the annual climatic cycle sometimes produces enough variation in snow strata on glaciers to allow annual units to be identified. In general, meteorology, sedimentation and diagenesis are so closely interrelated that the stratigraphy provides a climatic record which varies in length from several months in seasonal snow cover, to many years in glaciers.

I. Introduction

The Earth's surface in polar regions is typified by snow cover, both seasonal and perennial. This snow cover is a sedimentary veneer which is simultaneously a product, and a sensitive indicator, of climate. In this paper several meteorological and geological

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concepts are combined to study some problems involved in the origin, or deposition, of the snow, and post-depositional, or diagenetic, changes which occur in it.

The processes involved in a snowstorm may be compared with underwater transportation of sediments and subsequent deposition. The suspended load in a river is deposited when the speed of the water decreases. Also, the dissolved load is drastically affected, and partly precipitated, by chemical reactions which occur when a river enters the ocean. Similarly, masses of air transport a load of dissolved water vapor which, when the air is cooled, condenses, crystallizes and is deposited as snow. During and after deposition, snow is an aeolian sediment and surface features of erosion and deposition such as sastrugi, dunes, (often well-formed barchans) and ripple marks are formed by continuing wind action.

Stratification of snow results from variations in the conditions of deposition and is altered, sometimes emphasized, and sometimes destroyed, by subsequent diagenesis. The diagenetic processes proceed at rapid rates because the ice grains comprising snow are generally within 50 °C of the melting point. These relatively high temperatures imply high surface activity (Kingery, 1960) which is especially important for snow because it has a large amount of specific surface (surface area per unit mass). Furthermore, the air which moves through the pores of this permeable material is saturated with water vapor relative to ice. The amount of water vapor which saturated air can hold varies markedly with temperature and the temperature gradients within the first meter below the snow surface are often very steep (up to and exceeding 1°C per cm). Therefore, movement of this saturated air in the snow is accompanied by sublimation of water molecules from ice to air and vice-versa causing significant structural change and redistribution of mass within the snow pack. The extreme end products of these diagenetic processes are hard fine-grained wind slabs and soft, coarse-grained, loosely consolidated depth hoar layers.

The purpose of this paper is to discuss some field relations of these extreme snow types as products and indicators of climate. Mechanisms* of origin will be treated only briefly as required in the discussion since our emphasis will be on using the snow types as tools to interpret climatic conditions. We begin by considering the snow cover of Greenland, which varies from perennial on the ice sheet to seasonal in the coastal region and serves as an excellent introduction to the subject. Our questions concern the physical properties of this snow, and the rate and sources of its accumulation. Individual strata record specific meteorological events and climatic conditions. In general, meteorology, sedimentation and diagenesis are so closely interrelated that the stratigraphy provides a climatic record which varies in length from several months in seasonal snow cover, to many years in glaciers.

II. Stratigraphy and Accumulation

The following brief summary is based on Chapters III, IV, and VII of SIPRE Research Report 70 (Benson, 1962 a) where it is demonstrated that classical methods of

* The mechanism of depth hoar formation as dependent on diffusion rates of water vapor in the snow pack is treated by Giddings and LaChapelle (1962).

stratigraphy and sedimentation can be successfully applied to the snow and firn layers of the Greenland ice sheet. A meaningful stratigraphy exists in these layers, it is preserved for several decades, it is subject to detailed interpretation, and it has been correlated between pit studies spaced 10 to 25 miles apart. Identification of seasonal layers gives a measure of annual accumulation, and the correlation of annual units at 434 test sites along 1 100 miles of traverse has yielded a detailed picture of accumulation and precipitation for the years 1933 to 1955 on the west slope of the Greenland Ice Sheet.

Interpretation of snow strata is based more on similar layered sequences than on positive identification of a specific layer, such as a particular wind crust or ice stratum. The stratigraphic sequence represents a response to environmental changes. Since these changes occur over an annual cycle, similar sequences are produced each year. Summer strata are generally coarser-grained and have lower density and hardness values than winter layers; they may also show evidence of surface melt. Also, there is usually more variability in summer layers, with coarse-grained, loose layers alternating with finer-grained, higher density layers or even wind slabs of variable thickness. Winter layers are generally more homogeneous, with higher density and finer grain size than summer layers. To measure annual accumulation by integrating depth-density profiles, a specific reference datum in this annual sequency is needed. It should form within a short time interval and must be recognizable in all facies.

Selection of a reference datum

The short "fall season" produces a unique record in the snow. It is represented by a stratigraphic discontinuity in the form of a coarse-grained, low-density layer—often containing depth-hoar crystals—overlain by a finer-grained, harder layer of higher density. In regions where surface melt occurs, the discontinuity lies slightly above the topmost evidence of surface melt. The discontinuity at the base of the first year's snow is the base of the *Jungschneedecke* of Koch and Wegener (1930, pp. 371–372).

The time of formation of the sequence containing the discontinuity was determined by the exact dating of specific layers. In 1953 the low density layer formed during middle and late August; and the discontinuity between it and the overlying higher density layer was fully developed by early September. Specifically, at 77°N latitude, on the west slope of the Greenland ice sheet, the discontinuity forms between 20 August and 10 September.

The discontinuity is easily recognized and may be traced continuously over long traverses. In 1953, 1954, and 1955, it was recorded at each of 26, 68, and 27 stations made along a 220 mile segment of trail in northwest Greenland during the respective years. Sometimes the upper layer is a wind slab, sometimes it contains a series of thin, closely-spaced wind crusts, but always it is harder and of higher density than the layer below. At some stations ice masses in the snow make it difficult to obtain accurate density measurements. In these cases the ram hardness profiles are helpful in locating the discontinuity. Below the saturation line the discontinuity may be largely obscured during mid-summer by melt water wetting, soaking and percolation.

Mechanism of forming the reference datum

At the end of August and during early September the temperature gradients in the upper snow layers are steep (0.3 °C·cm^{-1} or more) and the accompanying vapor pressure gradients cause *upward* diffusion of water vapor and possible loss of mass to the atmosphere. By assuming an idealized situation with no movement of air through the firn, Bader (1939) computed the amount of moisture transferred by diffusion alone. His result was of the order of milligrams per square centimeter per day for the temperature range 0 to −10°C. Because this is so small, he concluded that movement of air through the snow is essential to significant transport of moisture within or between layers. Actually, it is impossible to eliminate convection of air in the snow when the temperature of the upper layers is lower than that of the layers below, because the highest density air is on top producing a turnover of air in the snow. This occurs even in the absence of winds in the air above the snow. When the temperature gradient is reversed a stable stratification of the air layers occurs and convection in the absence of winds is eliminated. Thus, all significant vapor transfer is directed upward. Winds, especially very gusty ones, produce rapid fluctuations of air pressure within the upper snow layers, and increase the rate of vapor transfer. As a result, material is redistributed within the upper layers, and some material is removed. Actual removal of some water vapor from the snow by wind was observed by Seligman (1936, p. 107).

A combination of strong wind and steep temperature gradient in the fall accounts for the observed formation of the low density layer. It also accounts for the observation theat this layer is thickest in the wetted and lower part of the percolation facies*. Temperatures in the near-surface layers are highest in these facies. This produces maximum transfer of mass by sublimation and evaporation because the difference in vapor pressure between two snow levels depends on the range of temperature involved as well as on the temperature gradient. For example, the temperature range 0 to −10°C produces more than twice the difference in vapor pressure produced by the range −10 to −20°C (3.5 as compared with 1.6 mb).

The ideal development of the discontinuity involves both extreme snowtypes; depth hoar and wind slabs. Their origins are often related because, although part of the upward migrating water vapor escapes to the atmosphere, the remainder is redeposited within the upper layers. According to Bader's computation for case with no wind, these amounts are nearly equal. If new or drift snow is being deposited on the surface, it will be indurated because some of the upward moving vapor is deposited in it. The sublimed vapor will first fill in the cracks to form "necks" between grains, because vapor pressure is lowest there. This process strengthens grain bonds and in the extreme case forms a wind slab; it was described as wind packing by Seligman who concluded:

> "… that wind-packing consists of the compacting of snow grains by the condensation of water vapor among them when subjected to the action of a moisture bearing wind. It is practically certain that at any rate some of this moisture is derived from the grains themselves. We can therefore define wind-packing as a special form of firnification accelerated by a wet wind. The mechanism of the processes is probably one of wind-accelerated diffusion which may or may not be influenced by the pulsations or pressure variations of the

* See data sheets 1 through 6 (Benson, 1962 a).

wind." (Seligman, 1936, p. 200).

Indeed, much of the moisture referred to as coming "from the grains themselves", comes from the low density layer below the wind slab.

The best example of wind-slab formation observed during the course of the writer's Greenland research formed on the first and second of September, 1952. The meteorological conditions which produced it are summarized here from observations made on an expedition in the vicinity of 77°N; 46°W, 2 150 m above sea level. Low clouds and light snowfall on 1 September turned into a blizzard during the night which lasted throughout the next day. Winds exceeded 40 mph, new snow was combined with blowing and drifting snow, and the air temperature remained near or above −5°C during the storm. On the morning of 3 September the sky was clear and the air temperature was below −25°C. The snow temperatuse was −25°C at the surface and −15°C at a depth of 20 cm. This produced an upward-directed vapor pressure gradient of 0.05 $mb \cdot cm^{-1}$, amounting to a total vapor pressure difference of 1 mb in the top 20 cm of snow. The weather remained cool and clear until 5 September and a hard wind slab covered the ice sheet inland from Thule Peninsula (Benson, 1962 a, p. 32).

Independent checks on stratigraphic interpretations

The importance of independent checks on stratigraphic interpretations has been, and should continue to be, stressed repeatedly. This is essential because the stratigraphy records climatic factors which vary continuously within climatic zones and sharply between them; such checks are especially important in setting up studies in new areas. This point is emphasized here because broad generalizations have been made comparing a "simple" situation in Greenland with a "complex" situation in Antarctica. It must be remembered that four years of field work combined with several additional years of analysis were required to interpret selected climatic areas of Greenland. Also, the following independent checks were made:

(1) Studies repeated annually for four years at selected points showed the preservation of certain features and the degree of certainty to be expected in interpretation (p. 33 and Figs. 22, 23, 24, and 25).*
(2) Good correlation was demonstrated between the amount of melt evidence in snow strata from 1947 to 1955 in northwestern Greenland with the number of degree days above freezing recorded at Thule (p. 41).*
(3) Specific layers were marked and errors introduced by using accumulation marker poles without marked layers were demonstrated (p. 34 and Figs. 22 and 26).*
(4) Oxygen isotope measurements were made in 1954 and 1955 and they correlated perfectly with the stratigraphic interpretations (pp. 41 to 43 and Figs. 32 a, b, c, and data sheet 4).*

The stratigraphy, graphically recorded and interpreted on 11 data sheets*, constitutes a "fast exposure snapshot" of a very dynamic sedimentary sequence. However, this snapshot gives a good representation of the climate. Furthermore, on the inland ice of

* These page and data sheet references refer to SIPRE Research Report 70 (Benson, 1962 a).

Greenland, this climate appears to have undergone negligible change over the past 50 years because data obtained throughout this time interval are internally consistent; the sources of these data are:

Date	Location of traverses and stations	References
1912	Traverse from 70°N on the West coast to 66°N on the East coast	de Quervain and Mercanton (1925)
1913	Traverse from 76°N on the East coast to 72°30′N on the West coast	Koch and Wegener (1930)*
1930–1931	Eismitte 70°54′N; 40°42′W	Sorge (1935)
1948–1951	French Central Station 70°55′N; 40°38′W	Bauer (1954)
1952–1956	Traverse to center from West coast at Thule (77°N) South to 71°N West to margin at 70°N	Benson (1962 a)

* In particular, the raw data from Koch-Wegener's two deep pits on the crest of the ice sheet, when corrected according to the procedure used by Benson (1962 a, Appendix B), agree well with the isotherm map drawn from the 1952–1955 data. The points are:

Lat.	Long.	Date	Corrected mean annual temperature
74°55′N	38°11′W	5–6 June, 1913	−31.6°C
74°23′N	41°24′W	11–12 June, 1913	−30.4°C

In summary, it has been effectively shown that the annual climatic cycle produces recognizable annual sequences of strata. Measurement of the water equivalent of these annual units on the ice sheet is the *only* satisfactory method of determining the amount of precipitation in Greenland. Precipitation gages at meteorological stations along the coast are strongly influenced by local conditions of exposure, and give readings which are invariably too low and apply only to a single point. In contrast, the ice sheet is an infinite set of automatically recording precipitation gages. It is only necessary to learn how to read the records.

III. Stratigraphy and Facies

In addition to providing a useful record of accumulation the stratigraphy also indicates that diagenetic facies exist in the snow and firn. Physical properties in the upper 50 m of the ice sheet differ significantly from one facies to the next and the boundaries between facies are defined and located by simple physical measurements.

Characteristics of the glacier facies may be summarized as follows:

(1) The *ablation facies* extends from the snout, or margin, of the glacier to the firn line. The *firn line* is the highest elevation to which snow cover recedes during the melt season.

(2) The *wetted facies** becomes wet throughout during the melting season and

* The terms "wetted facies" and "wet-snow line" are introduced here to replace the originally defined terms "soaked facies" and "saturation line" (Benson, 1962 a). The terms "soaked" and "saturated" clearly imply more than is wanted as pointed out by Müller (1962). Müller gives

extends from the firn line to the uppermost limit of complete wetting, the *wet-snow line**. The wet-snow line is the highest altitude at which the 0°C isotherm penetrates to the melt surface of the previous summer.

(3) The *percolation facies* is subjected to localized percolation of melt water from the surface without becoming wet throughout. Percolation can occur in snow of subfreezing temperatures with only the pipe-like percolation channels being at the melting point. A network of ice glands, lenses and layers forms when refreezing occurs. This facies extends from the wet-snow line to the *dry-snow line.* Negligible soaking and percolation occur above the dry-snow line.

(4) The *dry-snow facies* includes all of the glacier lying above the dry-snow line and negligible melting occurs in it.

The wet-snow and dry-snow lines give rise to discontinuities in temperature, density and rammsonde data, and may also be located by examination of melt evidence in snow strata. The altitude of the wet-snow line is dependent among other things, on annual accumulation, as is the firnline (Matthes, 1942, Chapter V, Glaciers) because they both involve the entire annual unit. The altitude of the dry-snow line is independent of accumulation, because it simply separates areas where melt either does, or does not, occur. The mean annual temperature** and snow density measurements together with the rates of accumulation are the most important parameters in defining facies boundaries.

The facies are not restricted to the Greenland ice sheet and a quantitative classification of glaciers based on them has been developed, using glaciers from Alaska, Washington and Greenland, (Benson, 1962 a, Chapter V, pp. 61–75). This "facies classification" is areal in nature and permits subdivision of large glaciers which span the entire range of environments from temperate to polar (Sharp, 1960, pp. 26–27). Ahlmann's (1948) useful distinction between temperate and polar glaciers takes on new meaning in the light of glacier facies. Thus, a temperate glacier exhibits only the two facies below the wet-snow line, whereas one or both of the facies above the wet-snow line are present on polar glaciers.

IV. A Hypothetical "North Polar Ice Sheet"

The search for symmetry in nature entices one to compare facies parameters observed

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a good argument for separating the originally defined soaked facies into two zones: a "slush zone" and a "percolation zone B" which are separated by the "slush limit". His "percolation zone A" is identical with the originally defined percolation facies but he chose not to separate his zones A and B with a labeled "line". However, between the wetted and percolation facies discontinuities occur in temperature, density, and hardness data, and in visually observed features of the snow strata (Benson, 1962 a, pp. 61–75), because the entire annual increment of snow is wetted, or raised to the melting point, in one facies but not the other. The temperature at the snow surface may reach the melting point all the way up to the dry snow line. This, of course, means that there is no surface expression of the wet-snow line. However, the boundaries between facies are not defined in terms of surface features; they depend on physical properties measured through several annual units.

** The snow temperatures measured 10 m below the surface represent mean annual air temperature.

on the Greenland ice sheet with those in other parts of the northern hemisphere, and even to draw comparisons with the southern hemisphere. In particular, it is interesting to speculate by extrapolating the mean annual temperature gradients.

An analysis of available temperature data indicates gradients with respect to altitude and latitude of about 1°C per degree latitude respectively. These gradients* define a set of north-sloping planes which intersect the surface of the ice sheet to form contours of mean annual temperature. Mock and Weeks (1966) used a digital computer to apply multiple regression techniques to all available 10 m snow temperature data from the Greenland ice sheet. One result of their analysis is an explicit representation of the maritime influence on isothermal surfaces observed south of about 67°N lat. This is roughly indicated by the change in latitude gradient observed in records from west-coast meteorological stations with the break occurring near Upernavik (73°N) (Benson, 1962 a, p. 56 and Fig. 39 b). In South Greenland Mock and Weeks also demonstrated an interesting longitudinal component which slightly warps the isothermal surfaces. However, in North Greenland the simple linear model, using the above stated gradients, is good enough for the following speculation.

Let us extrapolate the latitude gradient, determined on the Greenland ice sheet to 90°N, let us also replace the altitude at the North Pole (sea level) by the altitude observed at the South Pole (2 800 m). Thus, we mentally construct a "hypothetical North Polar ice sheet" similar to the Antarctic ice sheet, and arrive at a calculated mean annual temperature of −46.5°C (Benson, 1962 b). This compares well with the mean annual temperature of −50°C measured 2.5 m above the snow surface, and −50.9°C measured 12 m below the snow surface at the South Pole Station (Giovinetto, 1960, p. 2 and pp. 98–102).

Speculation: 1. Present glaciation

Let us now speculate on the agreement between pole temperatures on our "hypothetical North Polar ice sheet"** and the actual South Polar ice sheet. In the Arctic,

* The gradients were determined (Benson, 1962 a, pp. 53–60) as follows: (1) The altitude gradient was based on snow temperatures obtained 10 m below snow surface on traverses at 70 and 77 degrees north latitude. The gradient is essentially the dry adiabatic lapse rate for air and is interpreted as the result of a simple mechanism. Strong outgoing radiation produces strong surface inversions. The cold air within the surface inversion layer flows downhill by gravity to form the frequently observed katabatic winds. The katabatic winds are warmed adiabatically as they descend along the surface of the ice sheet, and this is assumed to be the primary mechanism determining the altitudinal temperature gradient along the snow surface. (2) The latitude gradient cannot be interpreted according to a mechanism or model as can the altitude gradient. Instead, it was based solely on data from two sources: (a) Average air temperatures per decade from 1870 to the present, measured at the Danish meteorological stations spanning 20° of latitude on the west coast. These stations are all within 32 m of sea level and, with exception of Alert, on Ellesmere Island, they all lie on the west coast of Greenland. Therefore, they were considered equal in altitude and in their general climatological setting. (b) Observed temperature measurements 10 m below snow surface, made above 2 000 m altitude on the ice sheet. These data give a latitudinal component of the temperature gradient which agrees well with that obtained from coastal temperature data north of Upernavik.

** I hope not to be accused of implying that there ever *was* such an ice sheet—this is purely a mental model.

the complex distribution of water with alternating flat and mountainous lands break our "hypothetical North Polar ice sheet" into fragments*. The largest "fragment", the Greenland ice sheet, is roughly one-eighth of the area of the Antarctic ice sheet and is not symmetric about the pole; instead, its center is located about 18° from the geographic pole. However, in spite of this it appers to control its climate, at least its temperature, in a manner similar to that actually observed on our only large ice sheet with polar symmetry. In this sense, it behaves as if it were part of a much larger ice sheet.

The dry-snow line may be expressed in terms of altitude and latitude because it is independent of accumulation (Benson, 1962 a, p. 74). On the Greenland ice sheet it descends toward the north at the rate of 1.15 m/km (127 m, per degree latitude) being 3 000 m above sea level at 70°N. Before proceeding, it is emphasized that one cannot expect the dry-snow line as measured in Greenland to be directly applicable to areas of intermittent oceans and mountainous land masses; these effects should displace it vertically upward. However, as a first approximation the linear extrapolation is useful as a rough guide for future work. With these reservations in mind, the altitude-latitude relationship of the dry-snow line observed in Greenland has been generalized in Fig. 1, which includes points on mountain glaciers in Alaska and on Axel Heiberg Island as well as Antarctic and Greenland altitude profiles. The stippled pattern indicates the dry-snow facies, the darker pattern refers to Greenland, the lighter to Antarctica. This diagram suggests that the dry-snow facies exists in the Wrangell Mountains and on Mt. McKinley, but not on the McCall and Upper Seward Glaciers. Let us examine the available evidence.

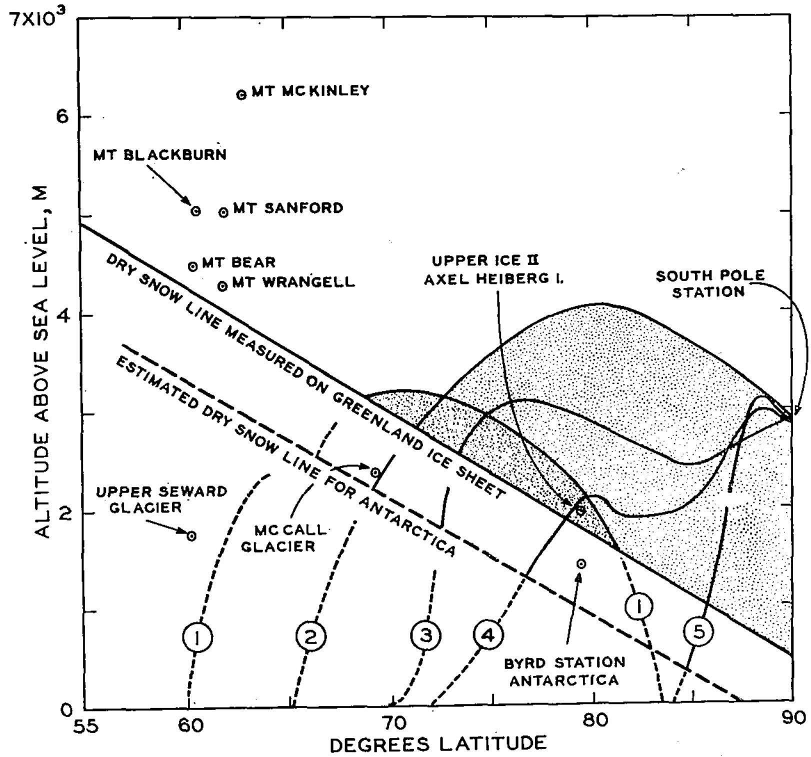
Guided by the above concept, a reconnaissance glaciological study made on top of Mt. Wrangell in 1961 (Benson, 1963), indicated that the dry-snow line is close to, but slightly above, the 4 200 m altitude on Mt. Wrangell. Data from the Upper Seward Glacier indicate that its accumulation zone is mainly in the wetted facies (Sharp, 1951 a, b; Benson, 1962 a, pp. 67–73). On the basis of this information it may be safely stated that the dry-snow line, and even the saturation line, lies above the Upper Seward Glacier which has a maximum altitude of 1 790 m at 62°23′N. On the other hand, existence of the dry-snow facies has been reported (Wallerstein, 1959) above 4 200 m on Alaska's Mt. Bear at 61°21′N. The upper part of the McCall Glacier in the Brooks Range is definitely below the dry-snow line according to mean annual temperature measurements between −1 and −2°C obtained from 91.5 m bore hole (Orvig and Mason, 1963). Müller's (1963) Upper Ice Station at 2 000 m on Axel Heiberg Island shows a mean annual temperature of −22°C which puts it close to the dry-snow line. All of these examples are reasonably compatible with the linear extrapolation shown in Fig. 1.

Since Alaskan glaciers are mountain glaciers they will show a pronounced effect of north and south slope exposure on the location of facies boundary lines. This effect was not encountered in the Greenland research because of the gently sloping uniform surface of the ice sheet. Also, as mentioned above, the discontinuous, complex distribution of ice masses in the western Arctic should lead us to expect the dry-snow line

* Sharp (1956, Figs. 1 and 2) gives a good picture of these "fragments".

to be at higher altitudes than observed in Greenland, but the slope of the line may well be the same. By the same token, in Antarctica the slope of this line may be the same as in the Arctic but with the line itself displaced to lower altitudes. The dry-snow line as shown in Fig. 1 is consistent with Antarctic field observations in 1961 by the writer at Roosevelt Island (79°16′S 600 m above s.l.) and Byrd Station, (80°S 1 500 m above s.l.), and with field discussions between Mario Giovinetto and the writer (see also Giovinetto, 1964).

In summary, Fig. 1 shows the differences between the North and South polar areas in terms of topography and the dry-snow line. In this frame of reference, the west slope of the Greenland ice sheet is obviously an entity comparable with parts of the Antarctic ice sheet. It is also clear that both polar areas have a multiplicity of climates —indeed, Greenland in itself has at least five climatic zones.



(1) NORTH-SOUTH ALTITUDE PROFILE ACROSS GREENLAND ICE SHEET
(2) ALTITUDE PROFILE ACROSS ANTARCTIC ICE SHEET ALONG 90°E MERIDIAN
(3) " " " " " " " 0° "
(4) " " " " " " " 90°W "
(5) " " " " " " " 180° "

Fig. 1. The dry snow line extrapolated in terms of latitude and altitude for the North and South Polar Regions. Ice sheet regions above the dry-snow line are indicated by the stippled patterns, the darker pattern is for Greenland the lighter one is for Antarctica. Qualifications resulting from obvious climatic differences are discussed in the text

Speculation: 2. Pleistocene glaciation

We all say, "The present is the key to the past". Fortunately, glacial geologists have been able to study evidence from extinct glaciers while glaciologists still work on "the key to the present". Prolonged speculation about Pleistocene glacier environments is not appropriate here. However, it is clear that our improved knowledge of the Greenland and Antarctic ice sheets contributes to understanding the nature of ice sheets in general, including extinct ones of the Pleistocene. At this juncture we shall only briefly consider two points.

First, the present ice sheets are not mere dying relics of Pleistocene glaciation. The Greenland ice sheet is in healthy equilibrium with present day climate (Benson, 1962 a, pp. 85–89); the same is true of the ice caps on Spitsbergen (Schytt, 1960).

Second, based on our present understanding of glacier facies and climates in both hemispheres, we may use Fig. 1 to speculate about the facies and climate which prevailed on the Pleistocene North American ice sheet. Assume that the altitude-latitude relations estimated for the dry-snow line on the present day Antarctic ice sheet may be applied to it. This assumption is consistent with the decrease in altitude of Pleistocene snow-lines compared with existing snow-lines as summarized by Flint (1957, pp. 46–49). Also, mid-continent glacial deposits show that the southern margin of the North American ice sheet extended a bit south of 40°N latitude, and that it barely reached the Cordilleran glaciers in the west and was bounded by oceans to the north and east. From present altitude profiles of the Antarctic ice sheet, and from present topography of North America, including Greenland, it is unlikely that the maximum altitude of the North American ice sheet exceeded 3 000 m south of 60° latitude. From Fig. 1, these conditions imply that the maximum latitude span of the dry-snow facies on the North American ice sheet was within 83°N and 60°N, and that the percolation facies spanned at least 20° of latitude at the southern end alone. Thus, if we compare the two hemispheres, the dry-snow facies was far more widespread in the Southern Hemisphere during the Pleistocene just as it is today, when, aside from the north dome of the Greenland ice sheet there are only isolated patches of the dry-snow facies in the Northern Hemisphere.

V. Comparative Studies

Prior to his 1929 and 1930–31 Greenland Expedition, Alfred Wegener stated: "The time has come to replace adventurous record journeys by serious scientific investigation" (Wegener, 1933, pp. 1–15). This was not an attempt to belittle the work done on previous expeditions which had been invaluable in determining the general size, shape and environment of Greenland. Instead, as stated by Spender (1934): "Wegener was now to attempt to overstep the limits which the earlier travelers set themselves; the inland-ice was to be his field of work, not a desert to be crossed as quickly as possible".

Wegener had remarkable insight and clearly saw the proper direction for future research. His goal of using the Greenland ice sheet as a laboratory has certainly been achieved. He also pointed out that the Greenland research of the 1930's would guide the Antarctic research of the future. All this has come to pass and we are now in a position to use the entire planet as a laboratory.

As an overall concept it is useful to stress the similarities of the polar regions and to look for specific causes and effects to explain their differences. One consequence of this approach is recognition of the need, and usefulness, of considering the Arctic as a unit. The presence of a continental ice sheet nearly symmetric about the pole tends to provide greater unity of thought and research in the Antarctic than has been enjoyed in the Arctic. The only comparable ice sheet in the Arctic has one-eighth the area, and its center is nearly 20° from the pole. At the North Pole, rather than a high continent, we have the Arctic Ocean which contacts both the Atlantic and Pacific Oceans and has a coastline which varies from plains to rugged mountains. The overall difference in altitude between the two polar regions, (sea level in the north compared with 2 000 to 4 000 m in the south) is obviously an important climatic parameter. However, the specific fact that the sea-level surface in the Arctic is indeed an ocean* means that differences between maritime and continental climates are inextricably involved in glaciological comparisons of the two polar regions. It is fascinating to think along this line since our present scientific maturity and logistic capabilities permit us to treat the polar areas as specialized laboratories in which we make observations and measurements on large scale experiments as they are conducted, by nature, at no expense to us. The experiments now being run were "set up" to show the effects produced by having an ocean nearly symmetrical around one pole and a continental ice sheet nearly symmetric about the other. If we were to design a planet for such research we could hardly improve on the one we have!

Comparative model studies of phenomena seen in both polar regions are useful because the manifestations frequently differ from one site to another, and more detailed knowledge of these differences may assist one in interpreting both areas. An example would be comparison of the "unique" dry valleys of the McMurdo Sound area (77–78°S) in Antarctica, with their nearly identical counterparts in Peary Land (82–84°N) of northeast Greenland. These areas have approximately equal temperatures and rates of precipitation, and both are noted for vertical cliffs on the snouts of their valley glaciers which terminate on deserts.

Another example would be comparison of the glacio-hydrothermal phenomena on Mt. Erebus and Mt. Wrangell. These are the extreme southern and northern active volcanoes associated with the Pacific Rim. A start on both the glaciological and volcanological aspects of this problem has been made on Mt. Wrangell, Alaska, where active hydrothermal alteration is underway on the Andesitic rocks of the summit (Benson and Forbes, 1964).

	Elevation	Latitude	Mean annual temperature in snowfield
Mt. Erebus	3 650 m (12 000 ft)	77 : 30°S	approx. −20°C
Mt. Wrangell	4 300 m (14 000 ft)	62° N	approx. −20°C

* The Arctic Ocean represents a large source of water vapor even though it is covered with ice. This is especially true in summer when the ice surface of the entire Arctic Ocean is wet, because the vapor pressures of water and ice are *equal* at the melting point. Furthermore, there are many open leads together with ponds of water on the sea ice during summer.

Another example would be comparison of Antarctic ice shelves (Ross, Filchner, Eights coast, etc.) with the smaller Arctic ice shelves (Ellesmere Island, Victoria Bay, etc.). Ice shelf research ties intimately with sea ice research, and should also include comparison of Antarctic ice islands with Arctic ice islands. There are obviously many other problems such as comparison of the evolutionary development of polar marine life isolated from temperate zones, by the Antarctic convergence, with polar marine life in the Arctic which is not isolated by such a mechanism. The important thing is to capitalize on our ability to study selected problems in both polar areas concurrently, and to look for symmetry, or reasons for lack of symmetry. Two examples of comparative studies will be briefly considered.

Greenland and Antarctic ice sheet stations

The broad-brush comparison of the Greenland and Antarctic ice sheets indicated in Fig. 1 was actually based on detailed measurements at many stations. A specific comparison of physical properties measured at selected points on the two ice sheets is instructive, and we shall do this by considering Byrd Station on the Antarctic ice sheet and seeking comparable stations on the Greenland ice sheet. The facies parameters which can be used to select points for comparison are mean annual temperature and rate of accumulation. Several stations which may be considered are:

Location and altitude above sea level			Rate of accumulation cm water equiv. per year	Mean annual temperature °C
Byrd Station	80°S; 120°W	1 500 m	17	−28
Station 2–200*	77°10′N; 49°46′W	2 460 m	22	−29
Station 4–0*	76°58′N; 46°59′W	2 616 m	16.5	−31

* Traverse stations (Benson, 1962 a).

These stations, with comparable rates of accumulation and mean annual temperatures, have similar stratigraphic features in the upper 10 m of snow, as should be expected. However, the overall range of density is greater at Byrd Station than at the two Greenland stations (Benson, 1967). This is most easily shown by comparing the top 5 m of snow. This depth includes several (about ten) annual units at each station so it is meaningful in terms of facies.

Location	Load ($g \cdot cm^{-2}$) at 5 m below snow surface	Average density ($g \cdot cm^{-3}$) in top 5 m of snow
Byrd Station	209	0.418
Station 2–200	185	0.370
Station 4–0	175	0.350

The variation between Byrd Station and the two Greenland stations is significant. On the west slope of the Greenland ice sheet the 5 m load value in the dry-snow facies is invariably less than 200 $g \cdot cm^{-2}$. The values exceed 200 $g \cdot cm^{-2}$ a little below the dry-snow line, and increase gradually through the percolation facies to about 225 $g \cdot cm^{-2}$

near the wet-snow line*. Byrd Station clearly does not fit into the percolation facies because of the low mean annual temperature together with the absence of melt evidence, so the differences in density expressed in the above table require explanation. Two factors seem to be involved namely: Wind action and range of temperature. These factors varied only slightly on the west slope of the Greenland ice sheet, where the facies were defined, and consequently they were not treated as variables. However, they vary significantly between Byrd Station and Stations 4–0 and 2–200. Byrd Station is windier than these Greenland Stations and it has a smaller temperature range. In particular, the average temperature of the warmest month based on Weather Bureau records at Byrd exceeds the mean annual temperature by 15°C as compared with 18°C** at Stations 4–0 and 2–200. The lower summer temperatures at Byrd Station cause the Fall temperature gradient to occur in a lower temperature range, and allow relatively less depth hoar development than at the Greenland stations. This is completely analogous to the decreased depth hoar development observed with increased altitude along the 77°N line in Greenland which was referred to above. Also, the relatively greater amount of wind action at Byrd Station would produce more wind packing. A more complete discussion will not be entered here; however, the variations in both of these factors are in the right direction to reinforce one another and contribute to the observed density differences.

Aleutian and Icelandic low pressure centers

Research in the Wrangell and St. Elias Mountains should be done in comparison with studies in the high coastal mountains of eastern Greenland between Angmagssalik and Scoresbysund. These two areas have similar meteorological settings. In both cases there is a marked change from maritime to polar continental climates. They involve the two most significant semi-permanent low pressure centers in the northern hemisphere. Many Pacific cyclones focus in the Aleutian center and hit the Alaskan-Yukon coastal mountains. This is similar to the way cyclonic disturbances in the Western Atlantic Ocean focus in the Icelandic low pressure center and act on the east Greenland Mountains especially south of Scoresbysund. Comparisons of this kind, within the northern hemisphere, and this one in particular, are as important as comparisons between the two polar regions. Detailed work on storm-facing slopes of the eastern Greenland mountains remains to be done, but some results are available from recent studies in the Wrangell Mountains.

The properties of the snow cover on top of Mt. Wrangell, Alaska (62°N; 144°W; 4 200 m above sea level) place it near the dry-snow line with a mean annual temperature of −20°C; but, the overall conditions are diffent from those observed at the Greenland and Antarctic stations mentioned above. Mt. Wrangell is more maritime, it is strongly, and directly affected by storms from the Aleutian low pressure center; also, its rate of snowfall is more variable both seasonally and from year to year, and generally runs 2 to 3 times greater than on the polar ice sheets. Individual storms often bring as much

* See Benson (1962 a, Fig. 46).

** This is estimated from the summertime deviations of 17.5 and 20°C observed at Station 2–100 (77°N; 56°W) and Eismitte (71.3°N; 40.5°W) (Benson, 1962 a, p. 52).

accumulation as the entire annual increment at stations 4–0 and 2–200 in Greenland or at Byrd Station. Also, the average monthly temperature range on top of Mt. Wrangell is about half that observed on the Greenland ice sheet*. The differences in these parameters produce different stratigraphic records; indeed, the record of a single storm on Mt. Wrangell closely resembles the annual stratigraphic record on polar ice sheets.

On Mt. Wrangell a significant variation in the amount of melt action fortuitously exists between the snow covers of two adjacent areas which are at virtually the same altitude. Strong winds pass over the active crater and nearby areas which have exposures of bare sand and ash. Some of the unconsolidated material is blown onto the snow cover of a restricted part of the summit area, accentuating melt action and producing excellent marker horizons; negligible melt occurs in adjacent areas. This situation has been useful in interpreting the snow stratigraphy, because the strata are quite homogeneous and difficult to interpret in places which have very high accumulation rates and no melting at all. In general, the snow stratigraphy on Mt. Wrangell has depended to a large extent on using identifiable marker horizons such as: rime ice layers, volcanic ash layers, and markers (30 cm square plywood boards) placed on the snow surface and located by probe rods after each snow storm.

VI. Seasonal Snow Cover

Although the southern hemisphere has the major part of the earth's perennial snow cover, seasonal snow cover is primarily a northern hemisphere problem. Alaska constitutes a seasonal snow cover laboratory which contains two sharply defined climatic boundaries:

(1) The Alaskan coastal ranges separate the north Pacific maritime climate from a severe continental climate.

(2) The Brooks Range separates the interior continental climate from the Arctic polar basin climate.

These two boundaries give three major climatic types which contain all varieties of snow cover:

(1) The coastal mountains of Alaska, the Yukon Territory and British Columbia are characterized by heavy maritime snowfall.

(2) The snow cover in interior Alaska, between the Brooks Range and Alaska Range, varies markedly from east to west. Toward the Bering sea, temperature and winds are higher than farther east and the climate is decidedly maritime; many of the snowstorms are mixed with rain. Toward the central and eastern parts of the area the climate becomes extremely continental with low winter temperature and low winds. The snow cover near the mouth of the Yukon-Kuskokwim Rivers is characterized by significant amounts of icing; towards the east, especially east of Koyukuk, the snow is dry, and has low density, in some cases the entire snow pack consists primarily of deph hoar with density less than 0.2 g/cm^3 even after it has been on the ground for five months.

* Based on comparing the mean monthly temperatures at the 600 mb levels from radiosonde data.

(3) The Arctic Slope of the Brooks Range has the longest lasting seasonal snow cover in Alaska. For at least eight months the Arctic Slope, from the foothills, across the tundra to the Arctic Ocean, is covered by hard, dry, wind-packed snow. Study of the amount movement, distribution, and physical characteristics of this snow is analogous to research on the physics of wind blown sand and desert dunes, and to studies of the surface layers on the Antarctic and Greenland ice sheets.

Research on the seasonal snow cover of Interior and Arctic Alaska has been underway by members of the University of Alaska since the 1961–62 winter*. Eight field test sites have been studied repeatedly on the Arctic Slope together with an equal number in interior Alaska and several sites along lines from Fairbanks to Valdez on the Pacific Coast, and to Kotzebue on the Bering Coast. The test areas have been established at sites which are topographically representative of large regions. Several glacier sites are included, but most of the test sites are in areas with long-lasting, *i.e.*, 6 to 9 month, seasonal snow cover. These studies require repeated observations over many years, with additional test areas and design modifications on existing ones as experience dictates.

Water equivalent of the snow is calculated from measurements of snow depth and density. Just as on ice sheets, this is the only reliable way of measuring the amount of precipitation which comes in the form of snow in wind swept areas such as the Arctic Slope. It has been shown that the rain gage catch at Pt. Barrow is between 25 and 50% of the actual amount of precipitation as measured in the snow cover (Black, 1945). Similar measurements in northwest Greenland have shown an even greater difference between rain gage catch and water content of snow on the ground (Benson, 1962 a). Also, in some Arctic areas the absence of wind combined with slow rates of precipitation may create errors because the observations, made on a six hourly schedule, sometimes produce a sequence of traces which are not entered quantitatively into the precipitation record (Jackson, 1960).

In addition to measuring the quantity of snowfall, an important objective of these studies is to understand (by measuring temperature, density and hardness profiles) the physical processes active within the snow pack and to relate them to climate. Measurements of stable isotope ratios ($0^{18}/0^{16}$ and H^2/H^1) and electrical conductance of melt water derived from snow strata are included.

Arctic Slope of Alaska

At four of the Arctic sites: Pt. Barrow, Umiat, Lake Noluk and Meade River, detailed plane-table and terrestrial photo surveys have been made on specific features such as river and lake banks which control drift action. The effectiveness of these

* A reconnaissance study of snow cover on the Arctic Slope has been supported by a grant from the National Science Foundation (Grant NSFG 22224), and by a grant from Dr. Terris Moore, former President of the University of Alaska. Logistic support in the field has been provided by ONR Task NR 307–272 through the Arctic Research Laboratory, (ARL), Point Barrow, Alaska.

Similar studies in Interior Alaska have been supported by a grant from CRREL, (DA-ENG-27-021-62-G 4) and by the Geology Department and Geophysical Institute of the University of Alaska with funds provided by the State of Alaska. Some field sites are being studied as a cooperative effort with the U.S. Soil Conservation Service.

"drift traps" depends upon their height and on their orientation with respect to prevailing and storm winds. The length of river and lake banks tabulated according to height, azimuth, and slope direction in the test areas may provide "drift corrections" for measurements made on flat surfaces such as lakes, rivers, and tundra within the test areas. A basic assumption in this work is that the complex drift patterns are reproduced in shape each year with variations restricted to quantity. Two major drift patterns are recognized, one from westerly storm wind and one from the easterly prevailing winds. The basic assumption appears valid, as during 1962, 1963, 1965 and 1966 the volume and vectorial aspects of the snow cover were nearly identical. However, during winter and spring 1964, the generally low precipitation in Alaska resulted in dramatically reduced drifts from storm winds, whereas the drifts from prevailing winds were not affected nearly as much.

Especially hard wind slabs are occasionally found in parts of the northern foothills of the Brooks Range where small twig and leave fragments are blown about and deposited with snow. They have been observed primarily in areas of sparse, dry vegetation in and near the northern foothills, where wind erosion and deposition is extreme and snow cover is discontinuous. The strengthening of the wind slab by bits of incorporated vegetation is analogous to the case of strengthening ice by adding a small percentage of wood pulp to form "pykrete" (Perutz, 1948). However, the exceptionally hard wind slabs and the large drifts on river banks are overshadowed by the dominant feature of the entire Arctic Slope snow cover as seen by air, namely, the wind oriented sastrugi patterns.

The wind-blow snow surface of the Arctic Slope is an excellent place to observe the response of sastrugi forms to variations in wind action. Permanent markers erected on the tundra protrude through the shallow snow cover and provide scale and azimuth for sastrugi patterns on aerial photographs. Since the areas involved are easily accessible, repeated flights can be made as conditions change. Improved knowledge of sastrugi response to wind will be useful in interpreting sastrugi patterns as indicators of surface air movement over snow surfaces (Lister and Pratt, 1959; Stuart and Heine, 1961), in connection with research on the evolution of sastrugi such as done by Gow (1965) at the South Pole staion, and in research on the origin and evolution of semipermanent patches of sastrugi on ice sheets (Crary, 1963, pp. 36–37; Benson, 1962 a, p. 32).

Even though annual sequences of snow are not available for stratigraphic analysis we do have two-thirds of an annual unit on the Arctic Slope. Seasonal variations throughout this portion of the annual precipitaion are being observed. Measurements of specific electrical conductance of melt water and variations in isotope ratios are being made in an attempt to detect seasonal variations within a single climatic zone as well as gross variations from one zone to the next.

Interior Alaska

Studies in interior Alaska have been concentrated on the diagenetic processes in the snow. The salient feature of this snow cover is its low density. In many cases almost the entire snow pack, 50 to 70 cm thick, consists of depth hoar with density less than $0.20\,g\cdot cm^{-3}$ just before melting begins. These represent "final" density values for the depth hoar since it has been developing in the snow pack for over 200 days. The final density for depth hoar in the Swiss Alps and in the Mountains of Colorado and Utah averages about 0.28 $g \cdot cm^{-3}$ after 100 to 150 days (Giddings and LaChapelle, 1962, p. 2381). The depth hoar formed at the base of the first annual unit on the Greenland ice sheet can also be represented by the mean value of 0.28 $g \cdot cm^{-3}$; measurents from more than 100 pit studies were nearly always within the range of 0.25 and 0.30 $g \cdot cm^{-3}$ with the lower values being found at higher altitudes (Benson, 1962 a). This value of 0.28 $g \cdot cm^{-3}$, for mountain and ice sheet situations, is significantly higher than the representative value of 0.20 $g \cdot cm^{-3}$ for interior Alaska.

The extreme development of depth hoar in interior Alaska snow cover may be understood by considering the prevailing conditions. The snow pack is shallow, with 50 to 80 cm a representative range of thickness. The bottom temperatures are generally −3 to −5°C and only rarely go as low as −10°C. The temperature on the snow surface is less than −10°C for about 5 months, less than −20°C for about 2 months, and reaches minimum values which go below −50°C. These conditions produce temperature gradients which are both steeper and of longer duration than those in thick mountain or ice sheet snow covers. For example, if the snow surface temperature on the Greenland ice sheet were −45°C, the temperature 50 cm below the surface would be about −40°C (Benson, 1962 a, p. 47), while in the seasonal snow cover of central Alaska it would be about −5°C. Because the Alaskan gradient is an order of magnitude greater and includes higher temperatures, it involves much higher absolute values of vapor pressure. In this example, the top 50 cm of interior Alaska seasonal snow cover has a vapor pressure difference 70 times greater than that in the Greenland case, even though the surface temperatures are equal; specifically the differences are 3.943 and 0.056 mb respectively. Also, on the Greenland ice sheet, strong upward-directed vapor pressure gradients exist only during the short Fall season when they produce the annual reference datum described above. On the other hand, the seasonal snow cover of central Alaska is exposed to such gradients for more than 5 months.

The low density snow described above is obviously not unique to interior Alaska. It is found wherever cold (below −10°C) relatively calm air overlies a shallow seasonal snow cover for several months. A typical example is provided by the following observations made in a wooded area near Kapuskasing, Ontario between 18 and 24 January 1954, (Benson, 1954). The snow cover was shallow, 28–56 cm, with temperatures of 0 to −2°C at the bottom and −15 to −35°C at the top. The temperature gradients averaged 0.4°$C \cdot cm^{-1}$ with extreme ranges of 0.17 to 1.0°$C \cdot cm^{-1}$. With exception of 5 cm new snow on top, the entire snow pack was composed of loosely bonded depth hoar with density nearly constant at 0.20 $g \cdot cm^{-3}$.

VII. Sources of Sedimentation

So far this paper has treated variations in the physical properties of snow layers near the air-snow interface. These properties primarily express variations in the diagenetic environment from one place to another. Variations in the depositional environment also occur and sometimes produce identifiable stratification even during an individual storm. A knowledge of variations occurring within single storms, especially prolonged ones of 4 to 6 days in areas of very high accumulation, may be useful in distinguishing between the layers produced by separate storms and the basic problem of identifying annual units. However, we shall not do this but turn instead to another aspect of the depositional problem and briefly consider the sources of "sedimentation" which, in the case of snow accumulation, are moisture-laden air masses whose motion can be traced on daily meteorological maps.

The most obvious precipitation source areas are the Pacific and Atlantic Oceans, with the Aleutian and Icelandic low-pressure centers constituting special-interest features for the north polar regions. These source areas have been considered in all discussions of large-scale glaciation. However, the Arctic Ocean area has been virtually ignored as a source area for snowfall. There are two main reasons for this: (1) quantitatively, it provides less moisture than the other oceans, (2) conceptually, an important reason for this goes back to 1888 when von Helmholtz advanced the idea of a nearly permanent polar anticyclone*. The smaller "glacial anticyclone" proposed by Hobbs in about 1910 has been dead for over 20 year. Similarly, the von Helmholtz polar anticyclone theory is now dying as a result of the large increase in available data from the Arctic. Indeed, the past two decades have seen marked revision of fundamental meteorological concepts in the Arctic; the large-scale glaciological consequences of all this need reappraisal.

The significance of Arctic meteorological data since World War II has been discussed by several authors. They were included in Klein's (1957) monumental analysis of all existing data on storm tracks for the northern hemisphere which shows many in the Arctic Ocean. Namias (1958 a, b) emphasized that:

> "It is now abundantly clear that cyclones and anticyclones traverse the Arctic in much the same fashion as in temperate latitudes, and upper level troughs, ridges, cyclones, and anticyclones, with dimensions and movements comparable to those of temperate latitudes, are synoptically common and transitory…" (Namias, 1958 a, p. 40).

Keegan(1958) analyzed Arctic meteorlogical data obtained from the winter months of the years 1952–1957. Kunkel (1959) treated the summer months for the same five year span. Both summer and winter data show large scale synoptic activity with cyclones outnumbering anticyclones about 2 to 1, and becoming more intense in the summer. These authors have discussed the new data in terms of general circulation in the Arctic. Perhaps the most important result from a glaciological point of view is contained in the following statement which pertains especially to summer:

> "…The Arctic is characterized by a high frequency of cyclonic activity with the greatest frequency occurring over the Arctic Basin between Alaska and the pole. The high frequency of cyclonic activity in the Arctic may be compared quite closely to the high frequency of cyclones near Iceland and the Aleutians …" (Kunkel, 1959, p. 46).

* A brief history of the influence of von Helmholtz's theory on Arctic meteorology is given by Kunkel (1959).

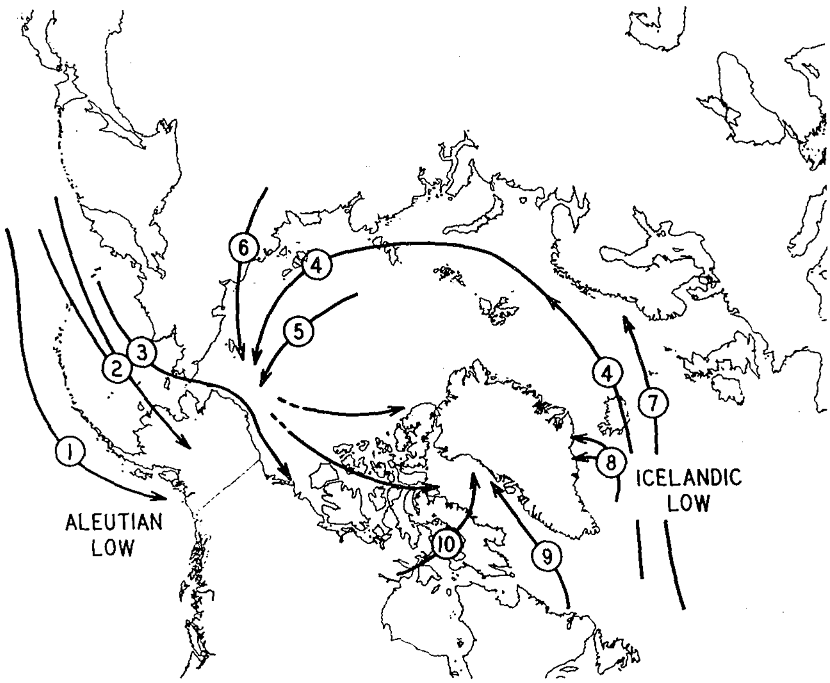


Fig. 2. Generalized Arctic and sub-Arctic storm tracks after Klein (1957), Keegan (1958) and Kunkel (1959)

As an aid to discussion, several generalized storm tracks in the Arctic and sub-Arctic are shown in Fig. 2. They were abstracted from the month-by-month analyses presented by Klein (1957), Keegan (1958), and Kunkel (1959). Storms from the Pacific and Atlantic Oceans often end up in the Aleutian and Icelandic low-pressure centers. These centers are so important and well known that they require no further discussion here. Instead, attention is called to storm paths in the Arctic Ocean. These are more complex because they have several possible origins. As stated by Kunkel (1959):

> "The lows which enter the Arctic Basin usually originate from Siberia (track 6, Fig. 2) or are old Icelandic lows (Track 4, Fig. 2). Most of the Icelandic and Siberian lows travel along the north coast of Siberia and then curve up into the Arctic Basin."

In the light of recent analysis Namias (1958 b) points out that the tendency for the Aleutian center to split into two cells, with one centered off Kamchatka, is greater than had previously been expected—especially in "non-summer months". Storms from this Kamchatka center often move through central Alaska (Track 2, Fig. 2) and in some cases they enter the Arctic Ocean (Track 3, Fig. 2). From the Atlantic side, some storms enter the Arctic Ocean by breaking out of the Icelandic low (Track 4, Fig. 2) as mentioned by Kunkel above. Also, storms moving northward from the Baffin Bay-Davis Strait system sometimes enter the Arctic Ocean (Tracks 9 and 10, Fig. 2). One analytical complexity, in the past and present,

"…is accounted for by difficulty in detecting disturbances breaking away from the Aleutian, Icelandic and Davis Strait centers of action, especially during periods of pronounced blocking." (Namias, 1958 b).

Hopefully, such disturbances may be easier to detect and follow now that polar-orbiting satellites are operational.

At this point it is useful to refer to specific interrelations between topography and moisture-bearing air masses revealed by the distribution of snowfall on the Greenland ice sheet (Benson, 1962 a, pp. 35–41). In two places the zone of maximum accumulation lies close to the coast. The most pronounced example, is where storms from the Icelandic low meet the high east coast mountains between Angmagssalik and Scoresbysund (Track 8, Fig. 2). This was mentioned above in comparison with the action of Aleutian storms on the Alaska-Yukon coastal mountains (Track 1, Fig. 2).

The other case involves the Davis Strait center of action. Here, the place where the zone of maximum accumulation approaches the coat lies on the south slope of Thule Peninsula. This is where the trend of the coastline changes from north-south to east-west as it forms the north shore of Baffin Bay. Cyclones moving northward along, and nearly parallel to, the west slope of the ice sheet (Tracks 9 and 10, Fig. 2) are forced to ascend Thule Peninsula directly here or to deflect westward to Devon and Ellesmere Islands. Accordingly, the ice sheet near the coast between 74°30′ and 76°30′N and the east-facing mountain slopes of the Canadian Islands between 75° and 81°N receive relatively heavy accumulation. Immediately to the south of this region, accumulation on the ice sheet gradually decreases with increasing latitude. A precipitation shadow exists on the lee side of Thule Peninsula which shows up on the ice sheet itself by a marked decrease in accumulation on the lee side of the crest line (Benson, 1962 a, p. 35 and Fig. 28). It is also apparent on the 1 : 5 000 000 map of Greenland since the coast of Melville Bay is characterized by glaciers which flow to the sea, while the north side of Thule Peninsula is comparatively free of glaciers. The Inglefield Land Peninsula also has glaciers on its south side but not its north coast.

The Davis Strait center of action is not as active as the Aleutian and Icelandic centers, but it is an interesting analogue to the Arctic Ocean center in the Beaufort Sea area. The analogy between storms moving northward along the west coast of Greenland and those moving eastward along the Arctic coast of Siberia and Alaska is not as clear-cut as the analogy between the Aleutian and Icelandic lows. Nevertheless there is an analogous situation. Studies underway on the Arctic Slope of Alaska show storm winds from the west and prevailing winds from the east. The same is true for Severnaya Zemlya and the Siberian coast in general (Kimble and Good, 1954). Such storms, moving along the Arctic Ocean coast, are not forced to ascend mountain slopes until they encounter the Romanzof Mountains of the Brooks Range near the Alaska-Yukon border. Similarly, they are not forced to move over-land to any major extent until they encounter the Canadian Islands and mainland near the MacKenzie River. These storms deposit relatively little snow as they move along the Siberian-Alaskan coast. However, they are induced to precipitate as they encounter the Romanzof Mountains which lie in their path much the same way as Thule Peninsula induces precipitation from the Davis

Strait storms. The effect produced by this air being forced to move onto the mainland or to deflect northward into the Canadian Arctic Islands varies with season. In the Fall, when the continent is cool relative to the ocean there is greater inducement to precipitation than during winter when the surface of the ocean and adjacent land areas are more similar.

Glaciologically, the Alaskan and Canadian Arctic Coast is an exciting area for speculation. The center of cyclonic activity in the Arctic Ocean is adjacent to it, and during Pleistocene time Cordilleran Glaciers joined the North American ice sheet in the lowlands east of the Romanzof Mountains, while the Arctic Slope immediately to the west was unglaciated. The total precipitation was probably less during Pleistocene than at present because most Pacific cyclone paths would be displaced southward. However, the unglaciated Siberian and Alaskan areas in contact with the Arctic Ocean probably maintained baroclinicity along the coast, and the resulting cyclones originating in, or at the edge of, the Arctic Ocean could be expected to provide precipitation like they do today, as demonstrated by Jackson (1961). It is likely that lower temperatures accompanying the southward displacement of Pacific cyclones would more than offset the decreased precipitation by converting some rainfall to snow*, and by reducing the current high ablation rates such as those measured on the McCall Glacier (Keeler, 1958). Basically, a lowering of temperature by one means or another seems to be the essential prerequisite for glaciation in this area. In addition to reducing ablation rates on glaciers it would also increase the chances for large seasonal snow drifts to survive the summer as occasionally happens in Arctic Slope river-banks today. Such persistent snow patches would reduce albedo and contribute to self-enhancement of glaciation, especially during cold cloudy summers. The next step in speculation involves variations in the sun's activity, *i.e.*, total radiation, the quantity of charged particles and the spectral distribution of the radiant energy combined with variations in the composition and albedo of the earth's atmosphere as discussed by Wexler (1953).

Speculation on causes of glaciation seems to be approaching a firmer foundation as knowledge of the polar regions increases**, and it is suggested that the Canadian Arctic Islands and the Arctic Coast in the Alaskan-Yukon border region constitute a critical area of unstable equilibrium.

Acknowledgments

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* Flint (1957, pp. 487-495) summarizes much of the literature on the interrelations between temperature, precipitation and ablation as related to causes of glaciation.

** A recent study by Fletcher (1965) is an important step in this direction.

at the University beginning with the Mt. Wrangell research in 1961.

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