Institute of Polar Studies

Report No. 17

Glaciological Studies on the South Pole Traverse, 1962-1963

by

Lawrence D. Taylor

Institute of Polar Studies

August, 1965

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**The Ohio State University
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Columbus, Ohio 43212**

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ABSTRACT

Glaciological observations were made at 25 stations along a 1448 km over-snow traverse between the South Pole and the Queen Maud Mountains during the austral summer, 1962-63.

Annual accumulation was determined from pit studies by the distribution of impermeable crusts, revealed by ink-staining pit walls, and the fluctuations in density of firn to a depth of 2 m. Climatological data indicates that impermeable crusts rarely form during winter. Stratification is complicated by disconformities, a break in the normal sequence of layering due to absence of one or more years of accumulation.

Accumulation varies from 6.8 gm/cm^2 at the Pole to 10.8 gm/cm^2 near the Queen Maud Mountains. Annual accumulation, mean density between 0 and 2 m, integrated ram hardness, and 10 m temperatures in firn all decrease with increasing latitude between $87^{\circ}08'S$ and $90^{\circ}S$ as follows: accumulation, $1.2 \text{ gm/cm}^2/\text{degree latitude}$; density, $0.037 \text{ gm/cm}^3/\text{degree latitude}$; integrated ram hardness, $1160 \text{ joules/degree latitude}$; and 10 m temperature, $1.5^{\circ}\text{C/degree latitude}$. Ten m temperatures in firn near the Pole indicate a lapse rate of $-0.8^{\circ}\text{C per 100 m}$. At depths between 15 and 40 m, negative temperature gradients of 0.56 to $0.64^{\circ}\text{C per 100 m}$ depth were recorded with largest gradients near the Pole. There is some indication that ice flow may be responsible for the gradient change.

Annual accumulation, 10 m temperature, and mean density of firn (0-2 m) plotted against each other have linear correlations of 0.94. The regression lines of these plots may be used to determine any two of the values if the third is known within the region of the traverse.

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INTRODUCTION

Glaciological and geophysical observations were made at 25 stations along a 1448 km over-snow traverse between the South Pole and the Queen Maud and Horlick Mountains from December 2, 1962 to January 22, 1963 (Fig. 1).

The party consisted of two geophysicists and two mechanics from the Geophysical and Polar Research Center, the University of Wisconsin; one geophysicist from the U. S. Coast and Geodetic Survey; and two glaciologists from the Institute of Polar Studies, The Ohio State University. The party leader and chief geophysicist was Dr. Edwin Robinson of the University of Wisconsin.

The glaciological program included pit studies, measurement of firn temperatures, and the collection of firn cores for microparticle studies. In two-meter pits annual snow accumulation was determined from firn stratigraphy, density, hardness, and grain size. Temperature studies included the determination of the temperature gradient in 40, 20, and 10 meter bore holes. Thermistors, a Wheatstone Bridge and null detector were used to measure temperatures to an accuracy of 0.01°C . Analysis of these data will contribute to the knowledge of the thermal diffusivity of firn, the heat budget and annual mean surface temperature of the Antarctic plateau.

Firn cores were obtained at each station to a depth of 10 m. These cores, to be studied for contained particulates, were shipped under refrigeration to the Institute of Polar Studies in Columbus, Ohio and to the Geophysics and Polar Research Center at Madison, Wisconsin. At the Institute of Polar Studies the distribution of particulates in these cores is being determined, in a dust free laboratory, by an electronic method using a Coulter Counter. This study may lead to an independent means of determining age and the amount of firn and ice accumulation.

Table I is a summary of the important data obtained on the traverse. Elevations and ice thickness (measured by seismic methods) were determined by Dr. Edwin Robinson of the Geophysical and Polar Research Center. Mr. David Perkins of the U. S. Coast and Geodetic Survey determined station locations by solar navigation.

The glaciological program was supported by the National Science Foundation under Grant No. G 23005 to the Institute of Polar Studies through the Ohio State University Research Foundation. The author gratefully acknowledges the logistic support offered by the Geophysical and Polar Research Center in charge of traverse operations. The author wishes to thank Mr. Henry Brecher for his excellent assistance both in the field and in the reduction of data and Mr. Tony Gow and Mr. Lyle Hansen of the Cold Regions Research and Engineering Laboratories for advice concerning firn stratigraphy and temperature measurements and for supplying calibrated thermistors. The author appreciates the many suggestions offered by Dr. Colin Bull. The manuscript was reviewed by Dr. Richard L. Cameron, Dr. Colin Bull, and Dr. Arthur Mirsky.

PIT STUDIES

General

The annual accumulation of snow in grams per square centimeter has been determined from an analysis of the physical properties of firn to a depth of 2 meters. Pits were dug (2.1 m deep, 2.6 m long, 1.7 m wide) at approximately 40 mile intervals along the traverse. One wall of the pit was used for the analysis, and care was taken not to disturb the surface above this wall. In each pit the density and grain size were measured, and the distribution of crusts, or impermeable layers, was determined using an ink-staining method. After the wall was studied, snow hardness was measured with a Rammsonde to a depth of 3 meters parallel to the wall and a half meter from the edge of the pit.

Interpretation of Firn Stratigraphy

The variation with depth of density, grain size, and hardness follows an irregular cycle. An excellent example of this cycle can be seen in Figure 2 showing the stratigraphy of pit 136. It is assumed that each cycle represents a year's accumulation, and is the result of recrystallization and compaction of firn produced by seasonal variation of: temperature gradient across the air-snow interface; solar radiation; and wind velocity and duration. Also, possible seasonal variation of precipitation, and/or crystal size of new-fallen snow. Other factors which tend to mask or disrupt the seasonal cycle are: compaction due to weight of overlying snow; absence of snow accumulation due to lack of precipitation or removal by wind producing a stratigraphic disconformity; extreme fluctuations in the normal seasonal conditions producing physical characteristics which are not ordinarily representative of that particular season. A disruption of the normal cycle is illustrated in Figure 2. First it must be established that certain seasonal variations in climatic conditions do exist in the area of the traverse. Because the elevations and latitudes of all the traverse stations are nearly the same, within 267 m and 3 degrees respectively, and for most observations not more than 2 degrees, it may be assumed that seasonal variations of climate along the traverse are of the same magnitude as at the Pole; although, gradual changes in absolute values for accumulation and mean annual temperature occur away from the Pole. For example, near the Queen Maud Range, Rammsonde profiles show a distinct increase in hardness and in the thickness of hard layers, indicating that wind velocities are much higher in this area than in other parts of the traverse.

Using the climatic and micrometeorological data obtained at the South Pole (U. S. Weather Bureau, 1963; Dalrymple, 1963) the seasonal variations and their effect on properties of firn at the Pole will now be considered.

Factors Responsible for Stratification

Temperature gradient. It has been established that a temperature gradient in firn produces a water vapor transfer in the direction of heat

flow (Bader, 1939, 1954, Yosida, 1963, Shimizu, 1964) which facilitates recrystallization at depth. The higher the temperature gradient the greater is the vapor transfer. Bader, (1939, 1954) determined that moisture transferred by diffusion alone is too small to account for the formation of depth hoar, and that air flow produced by wind is necessary. The formation of sublimation crystals, which produce a distinct stratigraphic marker, is dependent on the absolute temperature, the temperature gradient, the porosity of the firn, the permeability of the firn, and the wind velocity.

Sublimation crystals are formed by condensation or evaporation in the solid state. The process of condensation produces a general grain enlargement and reduction of porosity in the firn. Condensation occurs preferentially around large rather than small grains; since the former have a larger radius of curvature and thus lower vapor pressures. Evaporation generally destroys the small crystals, since they have a smaller radius of curvature and consequently higher vapor pressures. Thus condensation in the solid state produces a relatively coarse-grained impermeable layer or ice crust, and evaporation a fine-grained permeable layer.

It is unlikely that condensation can produce a thick ice crust, for recrystallization reduces the permeability and impedes vapor transfer. Once an impermeable layer is established, further condensation occurs on one side of the layer and evaporation on the opposite side provided heat flow remains in the same direction. The thickness of the crust is also limited, because the variation of vapor pressure with temperature is small at the low temperatures existing on the Antarctic plateau. For example, the difference in vapor pressure between -30°C and -60°C is 0.19 mm Hg, while between 0°C and -10°C it is 2.63 mm Hg (Dorsey, 1940, p. 600). Most crusts observed in firn on the Pole traverse were less than a centimeter thick.

The direction of heat flow must be considered in determining where condensation and evaporation occur in firn. If heat transfer is downward, then evaporation occurs in the permeable firn near the surface and vapor is transferred downward condensing in the colder, less permeable firn below, particularly on top of a dense layer deposited the previous winter. This condensation may produce an impermeable crust. Ice crusts already in existence below the surface may act as barriers to the downward transfer of vapor, and condensation may also occur along the top of these crusts. If heat flow is upward, then evaporation occurs in the permeable firn near the surface, as in the previous case, but vapor is transferred upward and condenses at the firn-air interface forming a thin crust. Condensation also occurs along the bottom of pre-existing ice crusts below the surface. Very little transfer of vapor upward or downward takes place at depths greater than two meters, for here the temperature gradient is greatly reduced and permeability of the firn is low.

Dalrymple's analysis of data from the South Pole shows that the largest temperature gradient across the firn-air interface occurs in December and January, with heat transfer downward. These gradients are 0.125°C/cm . The largest gradient with upward heat transfer, 0.075°C/cm ,

occurs during April (Fig. 3). According to these data it is assumed that during December and January ice crusts form below the surface at depths probably not greater than two meters, and usually at shallow depths on top of a dense layer deposited the previous winter. The crusts thicken just enough to prevent vapor transfer. Condensation occurs on top of the crust, and evaporation on the bottom and in the firn just below the crust. The porous firn resulting from evaporation just below an ice crust may remain for some time, because the overlying crust may be strong enough to support the load of firn above so that compaction of the porous layer is extremely slow (Fig. 3). During April when heat transfer is upward, an ice crust develops at the surface. If other crusts are already present below the surface, condensation occurs along the bottom of the crust and evaporation along the top. Porous firn that forms above this crust will soon compact, unless another ice crust is present just above, supporting the load of overlying firn (Fig. 3).

Solar radiation. Observations by Kotlyakov (1961) and Kuznetsov (1960) have indicated that preferential absorption of radiation produces intergranular and intragranular melting on the Antarctic plateau. The mechanism of recrystallization through radiation is not entirely understood, but recrystallization is sufficient along grain boundaries to produce a thin impermeable layer in the firn. Kotlyakov (1961) observed that it forms principally during periods of clear skies with little or no wind. The absence of wind reduces heat loss at the surface. Kotlyakov (1961) describes radiation crusts, in the vicinity of Mirny, as comprising elongated pieces of ice separated by small sections of firn. Almost all crystals in the ice are oriented at an angle to the horizontal plane with "c" axes parallel to crystal elongation and oriented in a northerly direction. Kotlyakov indicates that the structure of these crystals is the result of radiational melting. It seems unlikely that the absorbed energy can be enough to produce melting. Solid diffusion along grain boundaries may be the mechanism, whereby a crystal, whose "c" axis is aligned in the direction of the source of radiation, grows at the expense of an adjacent crystal whose "c" axis is not as close to the principal direction of radiation. In any case, there seems to be agreement that the following conditions are responsible for the formation of these crusts: (1) relatively high influx of infrared radiation which occurs at high elevations during clear weather, (2) presence of foreign particles which have a considerably larger coefficient of absorption of radiant energy than does ice, (3) concentration of radiant energy by the irregularities of the grains which in some cases act as lenses.

In dry snow the absorption of infrared radiation decreases exponentially with depth. Less than 7.5% of the total radiation at the surface reaches a depth of five centimeters (Dorsey, 1940, p. 492); therefore crystal enlargement is limited to the upper centimeter of surface. Ink-staining of the pit walls on the South Pole traverse (Fig. 2) clearly shows that the layers are extremely thin, less than 5 mm. Before staining, nearly all layers contain interlocking grains as large as 3 mm diameter. The discontinuous nature of some of the crusts can be seen in Figure 2. Without a knowledge of crystal orientation, it is difficult to establish whether these crusts are due to vapor transfer or radiation.

Wind velocity. According to observations at Pole Station, 1957-62, winds are generally stronger and more persistent during October (mid-spring), June and July (early winter), and March (early fall) than during other months. Wind variations produce pressure changes and cause air movement in the permeable firn. Air movement, in addition to diffusion, is needed to transport vapor. Winds help in the formation of many ice crusts; but only when the wind occurs at the time a large temperature gradient exists between the air and the firn. At the Pole the strongest and most persistent winds do not occur during the December to January period of largest temperature gradients. In December and January, 1957-62, the wind velocity averaged approximately 5.5 m/sec. It is unlikely, therefore, that ice crusts form during the mid-winter period on the Antarctic plateau where temperature gradients in winter are small, and vapor transport by air movement due to wind is small.

Although ice crusts may not form directly from the wind, high density layers of firn may indeed result. The relatively dense but still permeable firn is formed by packing of the snow; either by the force of the wind directly or later by the load of overlying snow. Wind-borne snow is somewhat rounded and well-sorted (Fig. 4, upper photograph), hence compaction takes place easily; whereas sublimation crystals (Fig. 4, lower photograph), which have irregularly shaped grains of different sizes, compact less readily.

Winds also produce cross-bedded structures as shown on the ink-stained wall of pit 124 at a depth of 160 to 200 cm (Fig. 5). Each layer is not an ice crust; it differs from the layer above and below in grain size, grain shape, and permeability. The character and sorting of the grains which form each layer is controlled by the wind velocity and the distance of transport by the wind.

Precipitation. Direct measurements of precipitation at Pole Station are inaccurate because of the very low accumulation and high wind velocity. If it could be proved that winter produces greater accumulation than summer, then thick layers of high density snow could be used as a winter indicator. Since there is no evidence from direct measurements that precipitation varies seasonally, this cannot be used as a criterion for determining a year's accumulation. Shimizu (1964) has shown that crystal size is generally smaller for snow precipitated in winter at Byrd Station. As far as the author is aware no such observations have been made at the Pole. Although seasonal temperature differences are of the same order at the two stations, the mean temperature at the Pole is so much lower than at Byrd that temperature differences at the Pole may have a negligible effect on the size of precipitated snow. Pit studies show a variation in grain size which to some extent parallels the variation in density. The change in grain size is, however, not a carry-over from the original size of the precipitated snow, but rather the result of compaction and recrystallization controlled by seasonal variations in temperature gradient, wind velocity, and radiation.

Density. Density measurements were made at 5 cm intervals using standard density tubes (500 cm^3) placed along two vertical rows with an overlap of approximately 0.8 cm between succeeding tubes. The tubes were excavated with great care and weighed on a triple beam balance to 0.1 gm. The density profile is the most reliable indicator of seasonal variations in the firn (Fig. 2). In the zone of compaction density generally decreases with increase in grain size in the firn. Since grain size is controlled mainly by temperature gradient, the largest crystals, excluding those comprising ice crusts, are associated with low density hoar formed by recrystallization mainly in December and January.

No relationship is apparent between density and elevation. Mean density plotted against latitude, however, varies according to a curvilinear regression (Fig. 6).

Grain size. Grain size measurements were determined every 5 cm by scraping the wall of the pit and catching the grains on a millimeter grid card (Fig. 4) and visually estimating the mean diameter of the aggregate. This is by no means as accurate as the technique of magnification recommended by Koerner (1964), but time did not permit a careful examination of grain size. In most cases the size varies inversely with density, but measurements were not accurate enough in many pits to be helpful in delineating seasons. Grain size varies from less than 0.25 mm to 5.0 mm as shown in Figure 4. The largest grains are found just below and sometimes above ice crusts. No relationship exists between mean grain size and change in latitude or elevation.

Hardness. Measurements of firn hardness using a SIPRE Rammsonde were made at approximately 5 cm intervals to a depth of 2.5 m, following the technique described by Bull (1956) and Benson (1962). The results are generally unsatisfactory for determining seasonal layers and can not be correlated from one pit to another. Thin ice crusts are difficult to identify as shown in Figure 2; cyclic variations which are clearly shown in the density profile are not obvious in the hardness profile. The ram measurements were useful in revealing a region of extremely hard packed firn along the $87^{\circ}45'$ latitude about 40 miles south of the Queen Maud Range, where values of 800 kg-force cm were measured (Fig. 5) compared to values of 400 kg-force cm elsewhere. Integrated values for ram hardness to a depth of 2.5 m were plotted against latitude (Fig. 6), and vary according to a curvilinear regression similar to density. If the scattered extremely high values obtained near the Queen Maud Range are ignored, the profile of integrated ram hardness versus latitude could be used to estimate density values following a procedure outlined by Bull (1956).

Permeability. The distribution of impermeable layers on the wall of the pits is determined by using an ink-staining method proposed by Shimizu (1964). One wall of the pit is smoothed by scraping and then brushing to make the crusts stand out. The smoothed surface is prepared along side of the excavation left from density measurements. The surface is sprayed evenly with warm dilute ink from a spray gun and allowed to freeze. A blowtorch flame is passed lightly across the inked surface and the melted ink is absorbed differentially by the firn before freezing again (Figs. 2, 5 and 7). Recrystallization occurs on the surface of the wall, but the

ink penetrates beyond the thin recrystallized surface and reveals gross changes in permeability of the firn. A layer of black ink accumulates just above each impermeable zone. If the heat is used sparingly and for the proper length of time, cross-bedded textures are revealed as shown in Figure 5. A centimeter tape is placed down the wall prior to all measurements and left in the same position until staining is completed to assure that a standard reference of depth is used for all observations. The stained wall is photographed in natural light using a 4×5 Speed Graphic, and 35 mm Contaflex camera with Panatomic-X film. The Speed Graphic is mounted on a sliding sleeve attached to a vertical tubular metal pipe pushed into the bottom of the pit and held in place equidistant from the pit wall by a board extending across the top of the pit.

ANNUAL ACCUMULATION

The annual accumulation varies from 6.8 gm/cm^2 at the Pole to 10.8 gm/cm^2 near the Queen Maud Mountains (Table II). Accumulation values near the Pole agree with those obtained by Giovinetto (1963) and Gow (personal communication). Values are based on the average total number of cycles identified for each of the properties of firn measured in the pits, where each cycle represents a year's accumulation. An interval on the wall of the pit is chosen commencing with the first good cycle in the density profile just below the surface, and ending usually at the bottom of the two-meter pit.

Variations in density, grain size and hardness, and the frequency of single crusts or sets of closely spaced crusts are counted over the same interval. The number of years from all four firn properties are combined and averaged (Figs. 2 and 5). The property which shows the clearest cycle in each pit is given preference in many cases. It is assumed from the previous discussion that ice crusts, whether from radiation or vapor transfer, are produced during December, January and April. In most cases below each crust a porous low density layer produced in association with the crust is present, and below this a thick layer of cross-bedded high density firn occurs representing the previous winter period, as illustrated in Figure 5 between 160 and 200 cm. In some cases the high density layer of firn is missing, as at 65 to 75 cm in Pit 136 (Fig. 2); or a series of ice crusts is present separated by only porous layers of firn, as shown at 132 cm depth in Pit 137 (Fig. 7). It is believed that this represents a missing winter layer. After the number of years is assigned to the section, the average density of the entire interval is multiplied by the thickness of the interval to obtain the water equivalent, which is then divided by the number of years or cycles to obtain the average accumulation per year.

All properties of the firn are used in the interpretation of a year's increment. For example in Pit 136 (Fig. 2), a series of impermeable crusts occur between 100 cm and 140 cm in the photograph. This section may represent three or more years with very low winter accumulation between each of the summer crusts. However, an examination of both density and

grain size profiles reveals only two cycles or two years, suggesting that the crusts were probably distributed over two summers.

It must be emphasized that the date of a given horizon or annual layer cannot be determined, because of the extremely low accumulation and the numerous disconformities (missing year or years of accumulation). Based upon the number of cycles of density, grain size, ram hardness, and the distribution of crusts one can only estimate the date, taking into account what appear to be disconformities. Disconformities are believed to exist at horizons indicated in Figures 2, 5 and 7. But do these disconformities represent one, two, three or more missing years? Are they truly disconformities? There is obviously a very definite need for a more precise means of dating firn in low accumulation areas that does not rely upon stratification.

An attempt was made to correlate the hard and impermeable layers using photographs of stained walls, hardness profiles, and cycles in density values. In a few cases the distribution and spacing in layers shows resemblance, particularly between pit 127 and 137 (Fig. 7). These pits are 11.2 km apart, have a difference in elevation of 102 m and a difference in latitude of 11.3'. Although the similarity is remarkable, considering the great variety of layers, it is indeed possible that this correlation is a coincidence. Many impermeable layers and layers representing a winter period are discontinuous and grade into disconformities. Accurate correlation on the basis of firn stratigraphy is impossible. Koerner (1964) has shown from trench studies at Byrd Station the great problem of correlation even over extremely short distances.

TEMPERATURE MEASUREMENTS

Firn temperatures ($\pm 0.01^\circ\text{C}$) were determined with thermistors in 10 m and 40 m bore holes using a Leeds and Northrup 4735 guarded Wheatstone bridge and a 9834 null detector. Calibrated thermistors were supplied by the Cold Regions Research and Engineering Laboratories.

A plot of 10 meter temperature with latitude is shown in Figure 8A. The 10 m temperature values (Table I) are equivalent to the mean annual air temperature to within 0.1°C . A change of -2.0 to $-5.8^\circ\text{C}/1^\circ$ latitude is considerably higher than values obtained by Langway (1961) for dry snow facies in Greenland, $-1.0^\circ\text{C}/1^\circ$ latitude. The regression line for 10 m temperatures versus latitude is not linear. The temperature may be influenced by the Queen Maud and Horlick Mountains, particularly in the vicinity of stations 119 to 125 (Fig. 1).

A further plot of 10 m temperatures at the same latitude, but for different elevations (Fig. 8B) indicates that elevation does influence the temperature in the vicinity of the Pole. The lapse rate at latitude $88^\circ57' - 89^\circ07'$ is $0.8^\circ\text{C}/100$ m. This value is slightly lower for the mean value calculated for eastern Antarctica of $1.1^\circ\text{C}/100$ m, and nearly twice the

value for western Antarctica, $0.35^{\circ}\text{C}/100\text{ m}$ (Bentley, et.al., 1964).

In Figure 8B, the 10 m temperatures at latitude $87^{\circ}55'$ - $88^{\circ}04'$ increase with increase in elevation. This anomaly is probably due to the effect of katabatic winds. Stations 119-121, which have higher temperatures than at other stations at the same or even lower elevations, are located near the crest of a ridge whose axis follows the 88° parallel. Cold air flowing down the flanks of this ridge is replaced at the crest by warmer air above. Without this influence the temperature at these stations would probably be nearly equal to Station 118. Winds in the vicinity of Stations 119-121 are stronger than elsewhere on the traverse as indicated by the high ram hardness and well-developed cross-bedding at these stations (Fig. 5).

At depths between 15 and 40 m negative temperature gradients of 0.56 to $0.64^{\circ}\text{C}/100\text{ m}$ were recorded (Figure 9). Similar negative gradients have been reported by Bogoslovski (1958), Wexler (1959) and Mellor (1960). There is some indication, although statistically not conclusive, that the negative gradient increases with increase in latitude (Fig. 9). Calculations by Mellor (1960) indicate that climatic warming best explains the negative gradients, rather than surface ice flow as proposed by Robin (1955). The fact that the gradients decrease with a decrease in latitude, in the vicinity of the Pole, is an indication that ice flow is responsible at least for the gradient change. Assuming that ice between the Pole and the Queen Maud and Horlick Mountains is flowing northward from the Pole toward these mountains, then it is possible that areas near the Pole are not being replaced by colder ice from higher elevations as rapidly as those areas bordering the mountains at lower elevation and latitude. It is realized that a local ridge divides the area under consideration, but this local topography, a relief of less than 150 m, should have little effect on the overall direction of flow controlled by the regional high of east Antarctica.

FIRM PROPERTIES AND ACCUMULATION AS A FUNCTION OF LATITUDE AND ELEVATION

Accumulation, mean density, integrated ram hardness, grain size, and 10 m temperatures are plotted against latitude and elevation. All but grain size vary as a function of latitude over the traverse range of $2^{\circ}52'$. The regression of each property on latitude is shown in Figures 6 and 8. There is little correlation between accumulation and firm properties plotted against elevation except in the case of 10 m temperatures near the Pole (Fig. 8B), mainly because elevation changes are small compared to changes in latitude.

Density and integrated ram hardness change exponentially with distance from the Pole in the direction of the Queen Maud and Horlick Mountains. Upon approaching these mountains and lower latitude the increase in density and hardness is caused by: a higher mean annual air temperature which facilitates recrystallization; a stronger and more persistent wind which transports, rounds, and sorts the snow; and a greater ice flow particularly at the heads of principal glaciers issuing through the mountains to the Ross Shelf.

The decrease in annual accumulation with increase in latitude, as shown in Figure 8C, is probably the result of cooling of the principal air masses transporting moisture from Marie Byrd Land to the South Pole Region. The drop in mean annual air temperature (determined from 10 m bore hole measurements) between 87°S and 90°S, the limits of the traverse, is 9°C which is sufficient to remove a significant amount of moisture from the air mass.

Annual accumulation, 10 m temperatures, and mean density of firn (0-2m) are plotted against each other (Fig. 10), with a linear correlation of 0.94. The regression lines of these plots determined from a least squares reduction may be used to determine any two values if the third is known within the general area of the traverse.

SUMMARY

The annual accumulation for 21 stations on the 1962-63 South Pole Traverse has been determined from an analysis of stratigraphy in two-meter pits. The entire traverse is between 2517 and 3137 m on the Antarctic plateau. The following assumptions are used as a basis for determining a year's accumulation.

(1) Ice crusts and permeable, low density layers of firn are formed by vapor transfer in summer (December and January) and late fall (April) during periods of large temperature gradients on the Antarctic plateau. Ice crusts, possibly formed by radiation, may also form during the summer period.

(2) High density layers consisting of fine-grained somewhat permeable firn, and in some cases containing cross-bedded structures but no ice crusts, are formed during mid-winter. Strong winds are responsible for the formation of many high density layers, for they break apart snow crystals, produce rounded grains, and depending upon the wind velocity and the nature of the undulating surface, deposit grains of approximately the same size. The resulting wind-blown deposits pack efficiently and, through compaction due to burial, form high density but still permeable layers. Strong persistent winds cannot form ice crusts because of the low air vapor pressure unless accompanied by a high temperature gradient, in which case the vapor is derived from the firn. Since high temperature gradients generally do not occur in mid-winter, the high winds during this time cannot produce crusts.

(3) A disconformity, or break in the normal stratigraphic sequence representing one or more years of no accumulation, is represented by either an abrupt change in the normal cyclic variation in density or by the presence of many closely spaced ice crusts separated by coarse-grained permeable hoar layers.

Values for annual accumulation, mean density between 0 and 2 m, integrated ram hardness, and 10 meter temperatures all vary as a function of

latitude between 90°S and 87°08'S with a change of -1.2 g cm^{-2} , -0.037 g cm^{-3} , -1160 joules , and -1.5°C per degree latitude respectively. Most values change exponentially with decreasing latitude. Lapse rates were not apparent except to a limited extent in the 10 m temperatures near the Pole, where a value of -0.8°C per 100 m was calculated. Elevation changes over the traverse were not more than 267 m with respect to the Pole; other factors such as latitude, ice topography and proximity to mountain ranges had a dominating influence. At depths between 15 m and 40 m, negative temperature gradients of 0.56 to 0.64°C per 100 m depth were recorded with largest gradients near the Pole. There is some indication that ice flow may be responsible for the gradient change.

Annual accumulation, 10 m temperatures, and mean density of firn (0-2 m) plotted against each other have linear correlations of 0.94. The regression lines of these plots may be used to determine any two values if the third is known within the region of the traverse.

CONCLUSIONS

Future studies should be directed toward a means of determining absolute dates of firn independent of accumulation rates or stratigraphy. Stratigraphic methods are limited to shallow depths, 10 m or less, and depend upon the observer's interpretation of disconformities, ice crusts, and hoar layer distribution. It is hoped that improvement in techniques of oxygen-isotope, deuterium-hydrogen, and lead²¹⁰ analyses and particulate analyses will supercede the stratigraphic approach of determining age and accumulation. Laboratory experiments which reproduce the seasonal conditions of the Antarctic plateau are needed to further clarify the processes of firnification particularly in the production of radiation crusts.

Future pit studies should include ink-staining techniques for revealing the distribution of impermeable crusts. Accurate density profiles are most helpful in determining the annual layer and should be continued, perhaps using narrower density tubes and sampling at closer intervals in areas of low accumulation.

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TABLE I

Data Summary of South Pole Traverse, 1962-63

Station	Location	Surface Elevation	Ice Thickness	Annual Accumulation	Mean Density 0-2 m	Integrated Ram Hardness	10-Meter Temperature
114	90°S	2800 m	2820 m	6.8 gm/cm ²	0.351 gm/cm ³	1079 joules	-50.7°C
115	89°38.1'S 141.4°W	2812	2922	7.5	0.360	788	-52.2
116	88°58.0' 140.0°	2912	2562	6.8	0.360	1049	-51.4
117	88°26.9' 140.4°	2955	2150	7.2	0.387	1454	-48.6
118	88°04.2' 163.5°	3136	2251	8.2	0.383	1522	-48.1
119	87°57.6' 153.0°	3002	1382	9.7	0.417	4144	-44.9
120	87°55.6' 139.1°	2835	2456	10.4	0.420	3874	-45.4
121	87°55.0' 126.2°	2840	2280	8.8	0.406	1437	-46.2
122	87°33.0' 122.0°	2749	2019	10.8	0.428	3742	-44.1
123	87°15.1' 121.5°	2645	1415	10.2	0.446	----	-41.4
124	87°08.7' 112.1°	2619	1779	10.0	0.400	2510	-43.0
125	87°45.7' 113.4°	2790	1780	9.5	0.399	1554	-45.9
126	88°22.1' 114.0°	2841	2191	8.6	0.385	1137	-49.0
127	89°06.7' 114.9°	2815	3185	7.2	0.365	659	-51.3
128	89°31.6' 112.5°	2772	2952	---	-----	649	----
129	89°31.8' 94.0°	2764	2849	---	-----	659	-51.1
130	89°02.2' 87.1°	2730	2710	7.6	0.363	584	-50.7
131	89°31.6' 84.7°	2696	2531	---	-----	693	-49.1
132	88°00.9' 83.2°	2586	2306	8.5	0.384	1157	-46.3
133	88°06.9' 69.7°	2578	2458	9.5	0.398	1507	-47.1
134	87°58.8' 56.2°	2517	2387	8.1	0.393	1103	-47.1
135	88°32.2' 48.9°	2611	2791	7.5	0.373	497	-49.1
136	88°57.0' 38.5°	2651	3041	7.3	0.382	681	-49.3
137	89°18.0' 14.8°	2713	2858	6.8	0.377	601	-50.5
138	89°42.1' 11.0°	2769	2799	---	-----	490	-51.3

TABLE II

Annual Accumulation, South Pole Traverse, 1962-63

<u>Station</u>	<u>Grams per cm²</u>	<u>Number of Years Represented</u>	<u>Total Depth Interval Cm.</u>
114	6.8	9	20 - 195
115	7.5	7	55 - 195
116	6.8	9	35 - 203
117	7.2	10	60 - 205
118	8.2	8	25 - 190
119	9.7	8	20 - 200
120	10.4	7	35 - 205
121	8.8	7	40 - 190
122	10.8	7	20 - 190
123	10.2	5	40 - 160
124	10.0	7	30 - 200
125	9.5	7	40 - 200
126	8.6	8	30 - 200
127	7.2	9	30 - 200
130	7.6	9	20 - 200
132	8.5	7	50 - 200
133	9.5	7	30 - 190
134	8.1	8	40 - 200
135	7.5	9	25 - 200
136	7.3	9	35 - 200
137	6.8	10	25 - 200

Figure 1

STRATIGRAPHIC COLUMN OF PIT 136 ILLUSTRATING CYCLIC VARIATIONS IN DENSITY, GRAIN SIZE AND RAM HARDNESS

Figure 2

TEMPERATURE GRADIENTS IN FIRN AND THEIR EFFECT ON VAPOR TRANSFER
AND THE FORMATION OF DEPTH AND SURFACE HOAR.

(TEMPERATURE CURVES ADAPTED FROM DATA IN DALRYMPLE, 1963 P.53)

Figure 4 - Grains Scraped from Wall of Pit (each division equals 1 mm)
 Grains at Top Represent a Winter Layer; at Bottom, a Hoar Layer

STRATIGRAPHIC COLUMN OF PIT 124 ILLUSTRATING CROSS-BEDDING AND UNUSUALLY HIGH RAM HARDNESS

Figure 5

VARIATION OF INTEGRATED RAM HARDNESS (0-2.5M DEPTH)
AND MEAN DENSITY (0-2 M DEPTH) WITH LATITUDE

Figure 6

Figure 7

A VARIATION OF TEN - METER TEMPERATURE WITH LATITUDE

B TEN-METER TEMPERATURE VERSUS ELEVATION FOR STATIONS WITH THE SAME LATITUDE

C VARIATION OF ANNUAL ACCUMULATION WITH LATITUDE

Figure 8

Figure 9

A MEAN DENSITY (0-2M DEPTH) VERSUS ANNUAL ACCUMULATION
 $R = 0.94 \quad 0.7 \text{ GM/CM}^2 / 0.01 \text{ GM/CM}^3$

B TEN-METER TEMPERATURE VERSUS ANNUAL ACCUMULATION
 $R = 0.90 \quad 0.55 \text{ GM/CM}^2 / ^\circ\text{C}$

C TEN-METER TEMPERATURE VERSUS MEAN DENSITY (0-2 M DEPTH)
 $R = 0.96 \quad 0.009 \text{ GM/CM}^3 / ^\circ\text{C}$

Figure 10