THE DISTRIBUTION OF 10 METER SNOW TEMPERATURES ON THE GREENLAND ICE SHEET

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Abstract. All available 10 m. snow temperatures from the Greenland Ice Sheet have been analyzed using multiple regression techniques to develop equations capable of accurately predicting these temperatures. The analysis was carried out for all Greenland and for various sub-areas. The resulting equations show that 10 m. snow temperatures can be accurately predicted from the parameters latitude and elevation. Longitude was found to be a further significant parameter in south Greenland.

Gradients of 10 m. snow temperatures versus elevation for north Greenland are close to the dry adiabatic lapse rate indicating adiabatic warming of katabatic winds as the controlling mechanism in their altitudinal distribution. In both south Greenland and the Thule peninsula, gradients of 10 m. snow temperatures versus elevation are markedly greater than the dry adiabatic lapse rate and are highly dependent upon elevation, indicating downward transfer of latent heat in the snow, largely as a result of percolating melt water.

An isotherm map, showing the distribution of 10 m. snow temperatures on the Greenland Ice Sheet calculated from the prediction equations, was prepared. The map is based upon a revised contour map of the ice sheet made from a compilation of all known elevations.

Résumé. *Distribution des températures à 10 mètres de profondeur de la neige de l'indlandsis du Groenland.* Toutes les températures à 10 m de profondeur disponibles de l'indlandsis du Groenland ont été analysées avec de multiples techniques de regression pour développer des équations capables d'une prédiction précise de ces températures. Cette analyse a été conduite pour tout le Groenland et diverses zones analogues. Les équations obtenues montrent que les températures de la neige à 10 m de profondeur peuvent être prédites avec précision à partir des paramètres latitude et altitude. La longitude est un paramètre supplémentaire valable pour le sud du Groenland. Les gradients des températures de la neige à 10 m de profondeur en fonction de l'altitude pour le nord du Groenland, sont proches des valeurs d'abaissement adiabatique de l'air sec, montrant ainsi que l'échauffement adiabatique des vents catabathiques est le mécanisme qui contrôle leur distribution en fonction de l'altitude. A la fois dans le sud du Groenland et dans la péninsule de Thule, les gradients des températures de neige en fonction de l'altitude sont nettement plus forts que l'abaissement adiabatique de l'air sec et dépendent grandement de l'altitude, montrant ainsi le transfert vers le bas dans la neige de la chaleur latente due en majeure partie à l'infiltration de l'eau de fonte. Une carte des isothermes des températures à 10 m de profondeur de l'indlandsis du Groenland a été établie, températures calculées à partir des équations de prévision. Cette carte se base sur une nouvelle carte altimétrique de l'indlandsis du Groenland établie après compilation de toutes les altitudes connues.

Zusammenfassung. *Die Verteilung der Schneetemperaturen in 10 m Tiefe auf dem grönländischen Inlandeis.* Alle verfügbaren Schneetemperaturen in 10 m Tiefe auf dem grönländischen Inlandeis wurden durch Mehrfach-Regression analysiert, um Gleichungen zur genauen Vorhersage dieser Temperaturen zu gewinnen. Die Analyse erstreckte sich über Gesamt-Grönland und über verschiedene Teilgebiete. Die resultierenden Gleichungen zeigen, dass Schneetemperaturen in 10 m Tiefe in Abhängigkeit von zwei Parametern, nämlich geographischer Breite und Meereshöhe, genau vorhergesagt werden können. In Süd-Grönland erwies sich die geographische Länge als weiterer bedeutsamer Parameter. Die Änderung der Schneetemperatur in 10 m Tiefe mit der Höhe stimmt in Nord-Grönland eng mit dem trocken-adiabatischen Temperaturgradienten überein, womit sich die adiabatische Erwärmung von Fallwinden als der bestimmende Faktor für ihre Höhenverteilung erweist. Sowohl in Süd-Grönland wie auf der Thule-Halbinsel ist die Änderung der Schneetemperatur in 10 m Tiefe mit der Höhe merklich grösser als der trocken-adiabatische Temperaturgradient und hängt stark von der Meereshöhe ab, was auf einen Abwärts-Transport von latenter Wärme im Schnee, weitgehend gebunden an durchsickerndes Schmelzwasser, schliessen lässt. Eine Isothermen-Karte wurde entworfen, aus der die Verteilung der Schneetemperaturen in 10 m Tiefe auf dem grönländischen Inlandeis auf Grund der Vorhersage-Gleichungen hervorgeht. Die Karte beruht auf einer revidierten Höhenlinienkarte des Inlandeises, in der alle bekannten Meereshöhen berücksichtigt sind.

Introduction

In the dry-snow and upper percolation facies (Benson, 1962) of glaciers and ice sheets, the snow temperature measured at 10 m. depth is a close approximation to the mean annual air temperature at the surface. Detailed analyses of the seasonal variation of snow temperatures as a function of depth (Sorge, 1935; Benson, 1962) have shown that, in the dry-snow facies of north Greenland, the 10 m. snow temperature theoretically varies no more than 0·30°C. from the mean annual air temperature. It is therefore of interest to study the spatial distribution

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of 10 m. snow temperatures in order to gain insight into the climatic patterns and meteorological and glaciological processes which control these temperatures. The present paper is an analysis of all available temperature data from Greenland using multiple regression techniques. Brief studies of two areas from the Antarctic are also included. Unless specifically stated otherwise, all temperatures and temperature gradients in this paper refer to 10 m. snow temperatures.

Previous Work

Isotherm maps of mean annual air temperature in Greenland have been prepared by Diamond (1960) and Benson (1962) and of mean annual snow temperatures by Bader (1961) using all data available at those respective times. The gradients of the mean annual air temperature (derived from 10 m. snow temperatures) versus elevation and versus latitude have been determined for north Greenland by Benson (1962). Benson found a latitude gradient of $-1\cdot 08°$C./° lat. existing on the ice sheet at 2,000 to 3,000 m. elevation and from lat. 71° to 77° N. This gradient was calculated using the temperature–elevation gradient found to exist at lat. 77° and $70\cdot 4°$ N. ($-1°$C./100 m.). Similar analyses have been made by several authors treating other specific areas of the Greenland Ice Sheet (Langway, 1961; Ragle and Davis, 1962; Mock, 1965). In all these studies the assumption has been made that the temperature–elevation and temperature–latitude gradients are mutually independent and that simple linear relationships exist. As a first and generally good approximation this appears to be the case.

Recently there has been considerable interest in the determination of trends and residuals in the areal analysis of geological and geophysical data by fitting first or higher order polynomials using least-squares regression techniques. Grant (1957) has defined a trend as a function $\tau(X, Y)$ that describes the behavior of the parameter of interest independently of any experiment. We seek, therefore, to obtain from a given set of physical measurements T_i, the function $t(X, Y)$ that provides the best possible unbiased estimate of τ. It is assumed that at each data point $T_i = t_i + \epsilon_i$, where $t_i = t(X_i, Y_i)$ is the estimate of the trend and ϵ_i is the residual. The residual ϵ_i is assumed to be a random variable with zero mean and a variance σ^2 which is the same for all observations. In a subject-matter context ϵ_i is usually associated with local and random effects. The decision regarding what terms of the polynomial should be utilized in estimating the trend and what percentage of the total variability in the data should be allotted to ϵ depend upon the investigator and the problem. Recent discussions of this general problem may be found in Grant (1957), Krumbein (1959), and Chayes and Suzuki (1963).

Method of Analysis

It is assumed that 10 m. snow temperatures can be predicted by an equation of the form $Y_p = b_0 + b_1 X_1 + b_2 X_2 + \ldots + b_n X_n$ where Y_p is the predicted temperature, $X_1, X_2, \ldots X_n$ are the independent variables, and the b's are the multiple regression coefficients. In normal trend surface analysis X_1 and X_2 are map or grid co-ordinates and the remaining X terms are quadratic or higher order powers of X_1 and X_2. In the present study the functional relationships between temperature, elevation, latitude, and in some areas longitude are investigated. The data consist of some 112 points on the Greenland Ice Sheet for which latitude, longitude, elevation, and 10 m. snow temperatures are available. The data were processed on a Control Data G-15 digital computer using a multiple regression program which calculated the following: the mean of the variables $\bar{X}$, $\bar{Y}$; the standard deviation of the means, $\sigma_{\bar{x}}$; the matrix of simple correlation coefficients, r; the multiple correlation coefficient, R; the standard error of the estimate, S; the F-test; the intercept b_0 and the multiple regression coefficients b_i; and the standard deviation and Student's t-test for each b_i.

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Figure 1 shows the location of the Greenland sample points. The data were processed on an areal basis in the following manner:

(a) All Greenland—includes all data points.

(b) North Greenland—includes all points north of lat. 67° N. minus one-half the points between lats. 76° and 77·5° N. that are west of long. 60° W.

(c) South Greenland—includes all points south of lat. 67° N.

(d) Thule peninsula, Greenland—includes all points between lats. 76° and 78° N. and west of long. 60° W.

It is obvious from Figure 1 that the distribution of points is far from uniform. The only correction for clustering was to eliminate approximately one-half of the points from the Thule peninsula area in the north Greenland analysis.

Each of the six areas was analyzed in the following manner. Two relationships, a linear and a linear plus a quadratic were assumed between the 10 m. snow temperature in degrees centigrade (Y_p) and the independent variables latitude in degrees and hundredths (X_1) and elevation in meters (X_2)

$$Y_p = b_0 + b_1 X_1 + b_2 X_2,$$

and

$$Y_p = b_0 + b_1 X_1 + b_2 X_2 + b_3 X_1^2 + b_4 X_1 X_2 + b_5 X_2^2.$$

In addition longitude in degrees and hundredths (X_3) was added to both models for south Greenland. The previously mentioned statistical parameters were calculated for each model and the values of F and t examined. In all cases the F-test, which is a measure of the significance of the regression equation as a whole, indicated that the null hypothesis, that all the true partial regression coefficients were equal to zero, could be rejected at the 1 per cent level of significance (Table I). Then the t values were used to test at the 1 per cent significance level the null hypothesis that each individual population partial regression coefficient (β_n) was equal to zero. If for a specific b_n, the null hypothesis is accepted (i.e. b_n is not significantly different from zero), a matrix shrinking sub-routine was used to recalculate the aforementioned statistical quantities deleting the particular non-significant X variable. This process was continued until the prediction equation was reduced to a form in which all the X variables were significant.

Meteorological Aspects of the Temperature Distribution

Greenland and Antarctica provide two of the largest areas of essentially uniform surface conditions existing on the Earth today. This is particularly true of Greenland which, although nearly an order of magnitude smaller than Antarctica, has over its greatest extent only minor surface irregularities to impede air movement. Under such conditions, the distribution of 10 m. snow temperatures can be expected to follow some regular pattern governed by elevation, latitude and possibly longitude.

Loewe (1936) noted that the annual air-temperature–elevation gradient measured between the Wegener expedition's western base and "Eismitte" in the interior was close to the dry adiabatic lapse rate (−0·98°C./100 m.). Similar temperature–elevation gradients in the snow were observed by Benson (1962) who expressed the view that the main factor controlling the altitudinal distribution of the mean annual air temperature is the relatively thin (*c.* 1 km.) layer of cold air which is chilled by radiational heat losses from the snow forming a temperature inversion. As this high-density layer flows downhill under the influence of gravity, it warms at the dry adiabatic lapse rate producing the observed gradient.

Essential to the view of the dry adiabatic control of the temperature distribution is the predominance of down-slope gravity (katabatic) winds. Other winds, particularly those flowing up-slope and associated with cyclonic systems will become rapidly saturated and will cool at a wet adiabatic lapse rate which is less than the dry adiabatic rate. This would tend to

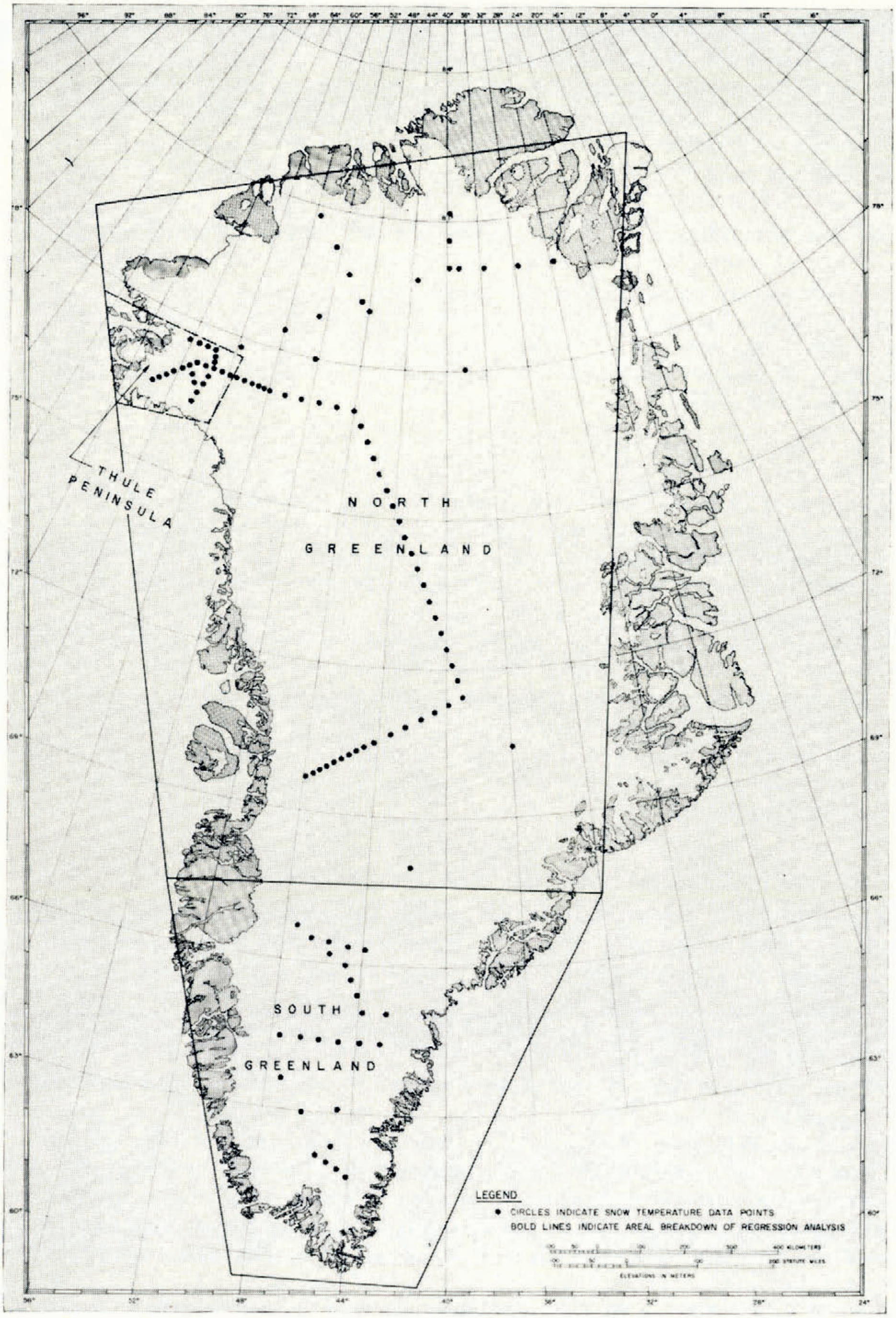

Fig. 1. Map of Greenland showing the location of 10 m. snow temperature data points

decrease the mean annual air-temperature–elevation gradient. It is necessary to know the conditions of the winds aloft in order to determine whether the surface winds are katabatic or simply part of the general circulatory pattern. Unfortunately meteorological records for ice-sheet stations are rare and those with upper-air observations even rarer.

Observations over a 25 month period at "Station Centrale" (near "Eismitte") at 3,000 m. on the ice sheet in interior Greenland lend support to the predominance of katabatic winds. Over the period of observations 81 per cent of all surface winds (during balloon ascents) were from the two eastern quadrants while 3 km. above the surface 59 per cent of the winds were from the two western quadrants (Ratzki, 1960). The observations showed that the wind shift usually started at the top of the temperature inversion. The location of "Station Centrale" was such that katabatic winds would be expected as easterly winds.

TABLE I. MULTIPLE REGRESSION EQUATIONS OF Y_p (10 M. SNOW TEMPERATURE, °C.), VERSUS X_1 (LATITUDE, DEGREES AND HUNDREDTHS), X_2 (ELEVATION, METERS), AND X_3 (LONGITUDE, DEGREES AND HUNDREDTHS) FOR AREAS OF GREENLAND AND ANTARCTICA

Location	*Equation*	*Multiple correlation coefficient, R*	*F*	*Standard error of estimate*
North Greenland	(1) $Y_p = 66.52 - 9.692\times10^{-1} X_1 - 8.504\times10^{-3} X_2$	0·987	1,391·219	0·64
	(2) $Y_p = 21.22 - 4.842\times10^{-3} X_1^2 - 1.145\times10^{-4} X_1 X_2$	0·989	1,570·390	0·60
South Greenland	(3) $Y_p = 92.72 - 1.157 X_1 - 1.508\times10^{-2} X_2$	0·964	143·327	0·81
	(4) $Y_p = 18.24 + 1.736\times10^{-2} X_2 - 5.044\times10^{-4} X_1 X_2$	0·969	169·016	0·75
	(5) $Y_p = 55.99 - 5.322\times10^{-4} X_1 X_2 + 3.8\times10^{-6} X_2^2 - 3.151\times10^{-1} X_3$	0·975	134·250	0·70
Thule peninsula	(6) $Y_p = 125.15 - 1.687 X_1 - 1.034\times10^{-2} X_2$	0·973	186·519	0·56
	(7) $Y_p = 27.16 - 6.470\times10^{-4} X_1 X_2 + 1.20\times10^{-5} X_2^2$	0·984	333·474	0·42
All Greenland	(8) $Y_p = 58.99 - 8.683\times10^{-1} X_1 - 8.574\times10^{-3} X_2$	0·982	1,475·252	0·94
	(9) $Y_p = 130.42 - 1.658 X_1 - 4.472\times10^{-2} X_2 + 3.348\times10^{-4} X_1 X_2 + 2.4\times10^{-6} X_2^2$	0·988	1,124·276	0·77
(dry snow facies)	(10) $Y_p = 24.34 - 5.287\times10^{-3} X_1^2 - 1.176\times10^{-4} X_1 X_2$	0·928	96·851	0·47
Marie Byrd Land	(11) $Y_p = 21.52 - 4.912\times10^{-1} X_1 - 6.596\times10^{-3} X_2$	0·847	94·828	2·16
	(12) $Y_p = 6.930 - 4.711\times10^{-1} X_1 - 5.712\times10^{-3} X_2 + 1.025 X_3$	0·915	126·749	1·65
	(13) $Y_p = 703.57 - 1.797\times10^{1} X_1 + 1.099\times10^{-1} X_1^2 - 7.02\times10^{-5} X_1 X_2 + 8.747\times10^{-1} X_3$	0·929	115·271	1·52
Victoria Land	(14) $Y_p = 32.93 - 5.198\times10^{-2} X_2 + 8.1\times10^{-6} X_2^2$	0·916	99·709	1·84

Other ice-sheet stations do not provide such clear-cut evidence for katabatic winds. "Northice", the base of the British North Greenland Expedition at an elevation of 2,343 m. on the ice sheet, was occupied for a period of 20 months from November 1952 to June 1954. Surface wind observations (Hamilton and Rollitt, 1957[a]) showed the prevailing winds to be from the north-west and west. Observations of surface winds at the time of balloon ascents showed a strong prevailing westerly wind at the surface while at 1·7 km. above the surface westerly winds still prevailed but with higher south-westerly and southerly components (Hamilton and Rollitt, 1957[b]). Since "Northice" was situated to the east of the ice-sheet crest, katabatic winds would be expected to appear as westerly winds. However since winds aloft are also dominantly from the western quadrants, no directional separation of katabatics from the general circulation is possible. Nevertheless the prevailing winds at "Northice", katabatic or not, are in the general down-slope direction and should warm at the dry adiabatic lapse rate. Several other stations have been occupied for varying short periods. Haywood and Holleyman (1961) summarized the available meteorological records for ice-sheet stations. They showed that, while the winds vary considerably in direction, the prevailing wind is generally down-slope.

Temperature profiles measured at "Camp Century", on the ice sheet, from 25 cm. below the snow surface to 400 cm. above the surface show the air and snow to be close to thermal equilibrium over a one year period (U.S. Army Signal Corps Meteorological Team Data, 1963). If this condition is general, and we further assume that the prevailing air-flow is down-slope, i.e. katabatic, then the adiabatic warming of this air should provide the major mechanism in the altitudinal distribution of 10 m. snow temperatures in the dry-snow facies.

Benson (1962) has pointed out that several meteorological and glaciological factors can cause the mean annual air-temperature–elevation gradient as estimated from 10 m. snow temperatures to depart from the dry adiabatic gradient. Briefly these are:

(a) *Heat transfer by melt water:* At locations where summer warming is sufficient to cause actual snow melt, water percolates downward into the firn where, upon refreezing, latent heat is liberated. This process has been well documented in areas of substantial melt by Sverdrup (1935) and Hughes and Seligman (1939). Schytt (1955) measured snow temperatures in the soaked facies where subsequently a meteorological station was established. During July and August 1954, temperatures at approximately 9 m. depth ranged from -5°C. to 0°C. Meteorological records (U.S. Army Signal Corps Meteorological Team Data, 1963) for a complete year gave a mean air temperature of -13°C. Where substantial melt occurs (at low elevations and latitudes) the effect is to increase observed 10 m. snow temperatures and hence to increase the observed 10 m. snow-temperature–elevation gradients. Where limited melt occurs, the effect upon 10 m. snow temperatures is not well understood.

(b) *Radiation cooling:* Radiational heat loss from snow surfaces will be more effective at higher elevations due to the decrease in the water-vapor content and the greater clarity of the atmosphere. Radiational cooling can also be expected to increase to the north (in Greenland) as a result of lower solar angles and decreased cyclonic penetration with associated cloudiness. The net effect of this factor would be to steepen temperature–elevation gradients.

(c) *Latent-heat transfer at the surface:* Heat may be subtracted from the snow by latent-heat transfer during evaporation. This in general is a function of wind-speed and humidity. However climatologic summaries for 1962 from "Camp Century" and "TUTO East", located 224 km. and 8 km. respectively from the western edge of the ice sheet, show no significant differences in either mean wind-speed or relative humidity (U.S. Army Signal Corps Meteorological Team Data, 1963). The net effect of this term on the temperature–elevation gradient can undoubtedly be considered small.

Results

North Greenland: Two models were used for north Greenland, a linear and a linear plus a quadratic as shown in Table I. The linear plus quadratic model is only a slight improvement over the simpler model. Figure 2 shows isotherms based on equation (2). (All numbered equations refer to Table I.)

Figure 3 shows the temperature–elevation gradient increasing with increasing latitude. The gradient is not dependent upon elevation. The magnitude of the gradient ($-0{\cdot}85$ to $-0{\cdot}95$°C./100 m.) indicates control by dry-adiabatic processes and suggests that this control increases to the north. That such is the case is not at all unreasonable. Cyclonic activity decreases to the north and with it the transport of air into the interior by up-slope surface winds; this tendency would decrease the observed temperature–elevation gradient. This sector of Greenland can be considered as a type example for adiabatic warming of katabatic winds controlling both the distribution of the mean air temperature and the 10 m. snow temperature.

South Greenland: Figure 4 shows isotherms plotted as a function of latitude and elevation based on equation (4). Figure 5 shows the -20°C. isothermal surface as a function of latitude, longitude, and elevation based on equation (5). The R values of all three models are such that each provides a good fit to the true 10 m. snow temperatures. The main differences in the expressions is in their degree and the functional aspects of the temperature–latitude,

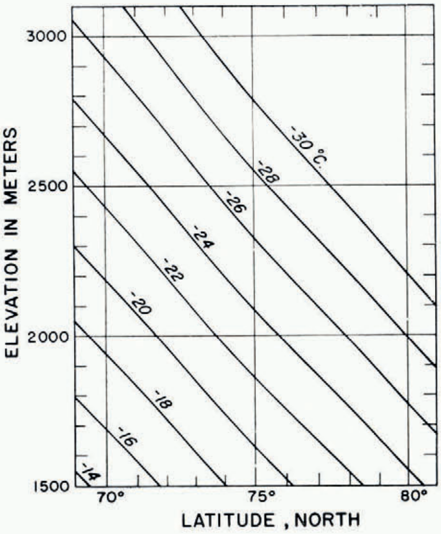

Fig. 2. Isotherms for north Greenland based on equation (2)

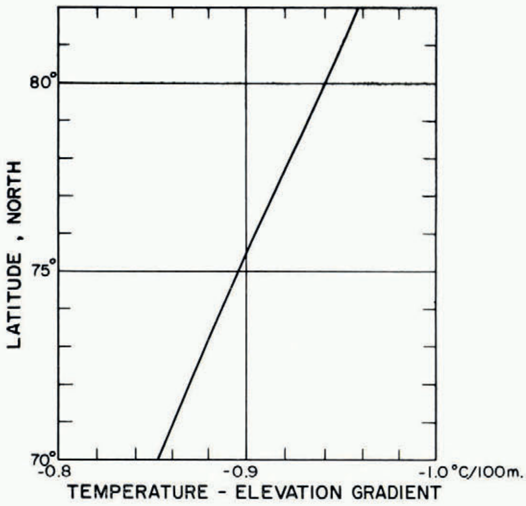

Fig. 3. Temperature–elevation gradient versus latitude for north Greenland based on equation (2)

temperature–longitude and temperature–elevation gradients. Equation (5), the best fitting expression, shows the 10 m. snow temperature to be a function of latitude, longitude and elevation.

Temperature–elevation isograds plotted against elevation and latitude are shown in Figure 6. Within the study area, the temperature–elevation gradient decreases with increasing elevation and decreasing latitude from $-2 \cdot 1$°C./100 m. to approximately $-1 \cdot 2$°C./100 m.

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These steep gradients and their decrease with elevation are best explained by considering briefly certain aspects of the ice sheet in south Greenland.

It has been noted previously that cyclonic activity increases to the south and this also seems to be true in this case, although there may be some localization of activity near the saddle at lat. 66° N. which separates the high domes of north and south Greenland. As has

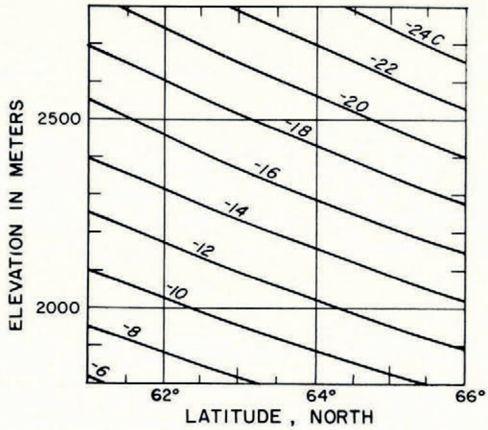

Fig. 4. Isotherms for south Greenland based on equation (4)

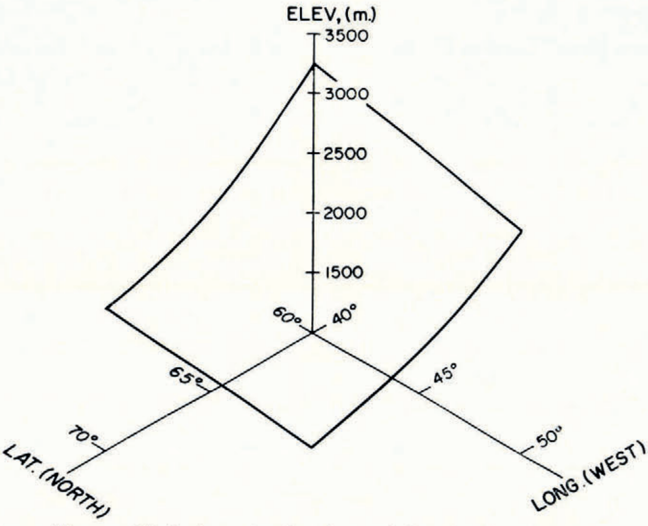

Fig. 5. The −20° C. isothermal surface for south Greenland based on equation (5)

been noted the effect of cyclonic activity should be to decrease the temperature–elevation gradient. However, in this case the gradients are much higher than expected. It seems likely that this is caused by percolating melt water which produces mass transfer of heat into the firn. Ragle (personal communication) and Davis (1964) have noted that the entire south Greenland area treated by equations (3), (4) and (5) lies in the percolation facies and in part approaches soaked conditions. At lower elevations there is proportionately more melt and greater downward heat transfer causing the 10 m. snow temperatures to be appreciably warmer than the mean annual air temperature. The data collected from points showing extensive melt-water migration when combined with data collected at higher elevations where the 10 m. snow temperature is a more representative measure of the mean annual air temperature, then give a fictitiously high air-temperature–elevation gradient.

As expected a temperature–latitude gradient exists, but its magnitude depends upon elevation (Fig. 7). It is also interesting to note that a temperature–longitude gradient of

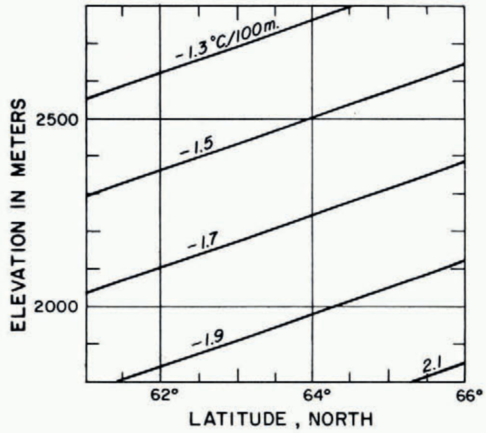

Fig. 6. Temperature–elevation isograds versus elevation and latitude for south Greenland based on equation (5)

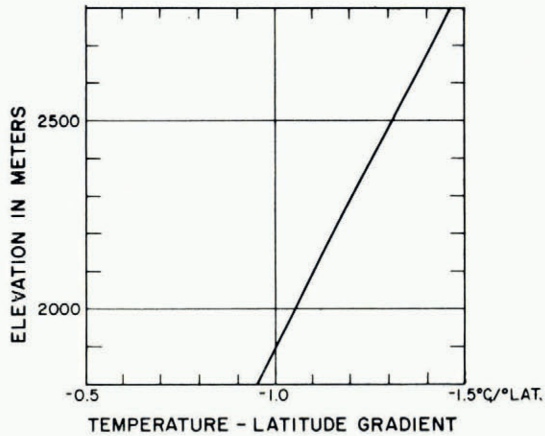

Fig. 7. Temperature–latitude gradient versus elevation from south Greenland based on equation (5)

approximately $-0\cdot3$°C./° long. exists (equation (5)). This may be a reflection of cooler water temperatures and greater ice cover in Davis Strait to the west than in Denmark Strait to the east.

Thule peninsula: The Thule peninsula has been the site of several extensive studies (Benson, 1962; Mock, 1965) and, therefore, has a greater density of temperature determinations than any other corresponding area in Greenland. Furthermore, it has several characteristics which make it glaciologically interesting: a distinct windward and leeward slope with a corresponding precipitation shadow, a distinct ice divide, and three years of continuous weather records from "Camp Century". Linear and linear-plus-quadratic models were used, with the more

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complex model providing a more significant equation. Isotherms as a function of latitude and elevation are shown in Figure 8. Figure 8 also illustrates the danger of extrapolation far beyond the elevation and latitude of the data, in this case above 2,000 m.

Climatic records for the year 1962 are available for Thule Air Base, (U.S. Weather Bureau, 1962), Camp TUTO, "TUTO East" and "Camp Century" (U.S. Army Signal Corps Meteorological Team Data, 1963) all on the Thule peninsula. Thule Air Base is located on the coast 13 km. from the ice sheet at 11 m. elevation, Camp TUTO at the edge of the ice sheet at 489 m. elevation, "TUTO East" on the ice sheet 8 km. from the western edge at 801 m. elevation and "Camp Century" on the ice sheet 224 km. from the western edge at 1,885 m. elevation. All stations are within 1° of latitude and thus provide a profile from the sea coast to the interior. Figure 9 shows the mean surface air temperatures at the four stations

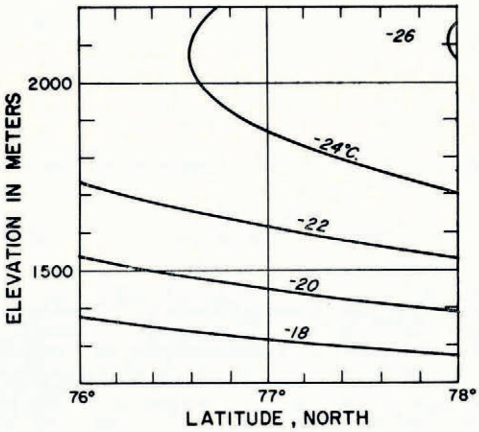

Fig. 8. Isotherms for the Thule peninsula, Greenland based on equation (7)

for January 1962, July 1962, and for the entire year 1962 plotted as a function of their elevation. Also shown are the upper-air mean temperatures for the same periods from radiosonde ascents at Thule Air Base. It is interesting to note the close correspondence of the free-air lapse rate with the temperature–elevation gradient existing along the land surface (i.e. from Thule Air Base to Camp TUTO). Of more interest to this study, however, is the close correspondence of the mean annual air temperature–elevation gradient over the ice sheet (i.e. from "TUTO East" to "Camp Century") with the dry adiabatic rate and the departure from the dry adiabatic rate off the ice sheet. Figure 10 shows the 10 m. snow-temperature–elevation gradients as functions of latitude and elevation. The effect of latitude upon the gradient is small but the elevation effect is even more pronounced than in south Greenland. This is a result of generally heavy summer melt conditions, as in south Greenland, over the relatively low elevations on the Thule peninsula, apparently causing 10 m. snow temperatures to be systematically higher than mean annual air temperatures resulting in temperature–elevation gradients greater than the dry adiabatic rate. That the adiabatic warming of katabatic winds is still the controlling mechanism in the mean annual air temperature distribution is shown by the meteorological records.

All Greenland: Linear and linear-plus-quadratic models were used for analysis of all the Greenland data. Considering the difference in conditions, particularly as reflected in the temperature–elevation gradients, between north and south Greenland, the resulting equations for all Greenland show surprisingly good fits. Figure 11 shows isotherms based on Equation (9).

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Figure 12 shows the temperature–elevation gradient as a function of latitude and elevation. This gradient shows a general decrease with latitude, reflecting the overall decrease in the magnitude of the gradient from south to north Greenland despite the fact that within north Greenland there is a slight increase with latitude.

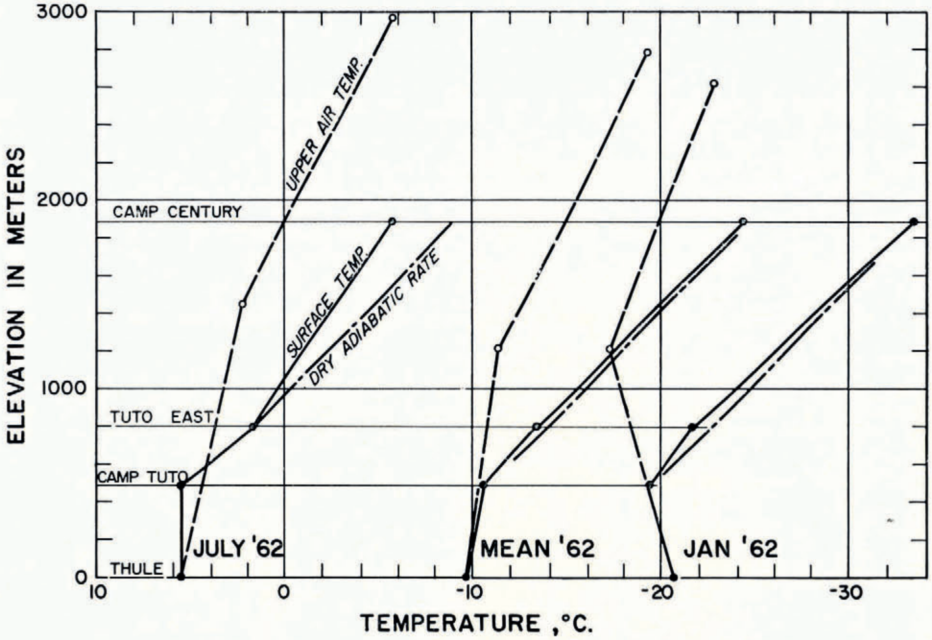

Fig. 9. Comparison of mean surface temperatures at Thule Air Base, Camp TUTO, "TUTO East", and "Camp Century" with mean upper air temperatures at Thule Air Base

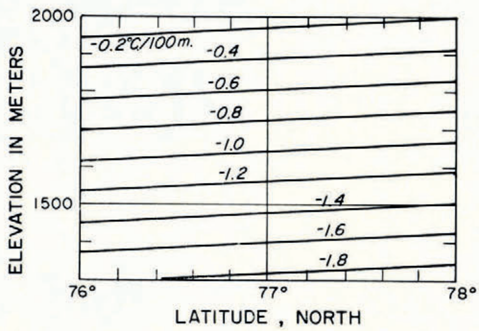

Fig. 10. Temperature–elevation isograds versus latitude and elevation for the Thule peninsula, Greenland based on equation (7)

Figure 13 is an isotherm map of Greenland showing the distribution of 10 m. snow temperatures based on Equations (2) and (5). Temperature measurements from east of the ice-sheet crest and from the area from lat. 66° N. to 70° N. are generally lacking. In particular the area from lat. 66° N. to 70° N. seems to represent a major discontinuity both meteorologically and glaciologically. For this reason the isotherms are dashed through this zone and are not as valid predictions as to the north and south. Inasmuch as the accuracy of such an isotherm map is no better than the ice surface contour map upon which it is based, first a new base map was constructed from a compilation of all known sources of elevations on the Greenland Ice Sheet. These elevations, 1,589 in total, were plotted on the current 1 : 1,000,000

World Aeronautical Charts of Greenland. A few of the points which appeared contradictory were rejected and the final ice surface contour map was prepared by inspection (Fig. 14). The compilation of these data which also include mean annual accumulation, 10 m. snow temperatures and ice thickness where any or all have been determined is included as an

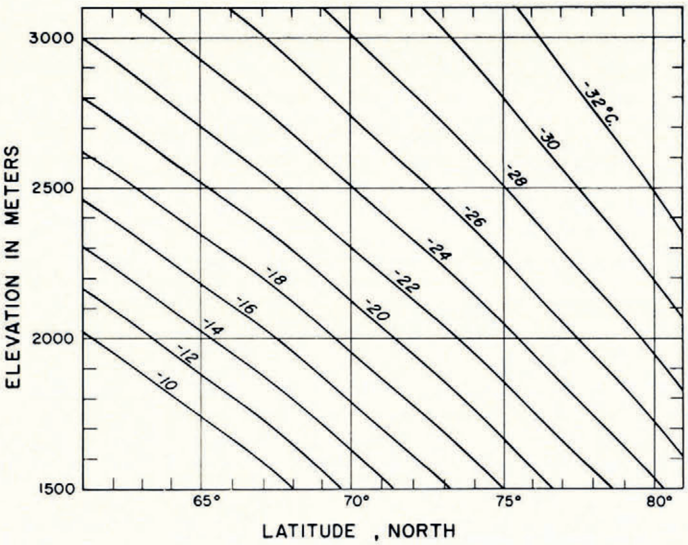

Fig. 11. All Greenland isotherms versus latitude and elevation based on equation (9)

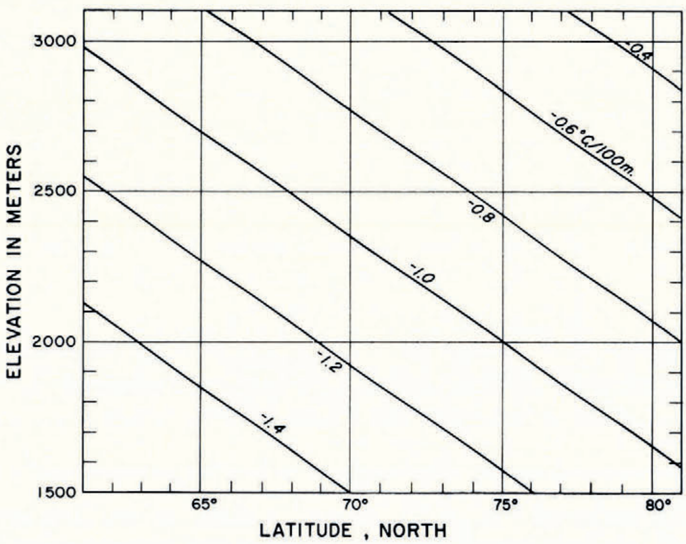

Fig. 12. All Greenland temperature–elevation isograds versus latitude and elevation based on equation (9)

appendix to *U.S. Cold Regions Research and Engineering Laboratory, Research Report* 170 and will be available on file at U.S. Cold Regions Research and Engineering Laboratory.

Dry-snow facies: In an attempt to define the distribution of mean annual air temperatures, as opposed to 10 m. snow temperatures, a multiple regression analysis was made using only 10 m. snow temperatures from the dry-snow facies. Figure 15 shows isotherms for all Greenland calculated from the resulting equation (equation (10)). The fact that the data from which

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Fig. 13. Isotherm map of Greenland based on equations (2) and (5)

Fig. 14. Contour map of the Greenland Ice Sheet

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this equation was calculated all lie within the encircled area indicates the degree of extrapolation. It is believed that this extrapolation is not unwarranted because of the nearly linear character of the isotherms. Mean annual air temperatures from stations having meteorological data for one year or more are shown as triangles. The open circles indicate measured 10 m. snow temperatures in south Greenland. Figure 16 shows contours of the deviations of measured 10 m. snow temperatures from the mean annual air temperature predicted by equation (10). Several interesting features are shown in Figures 15 and 16:

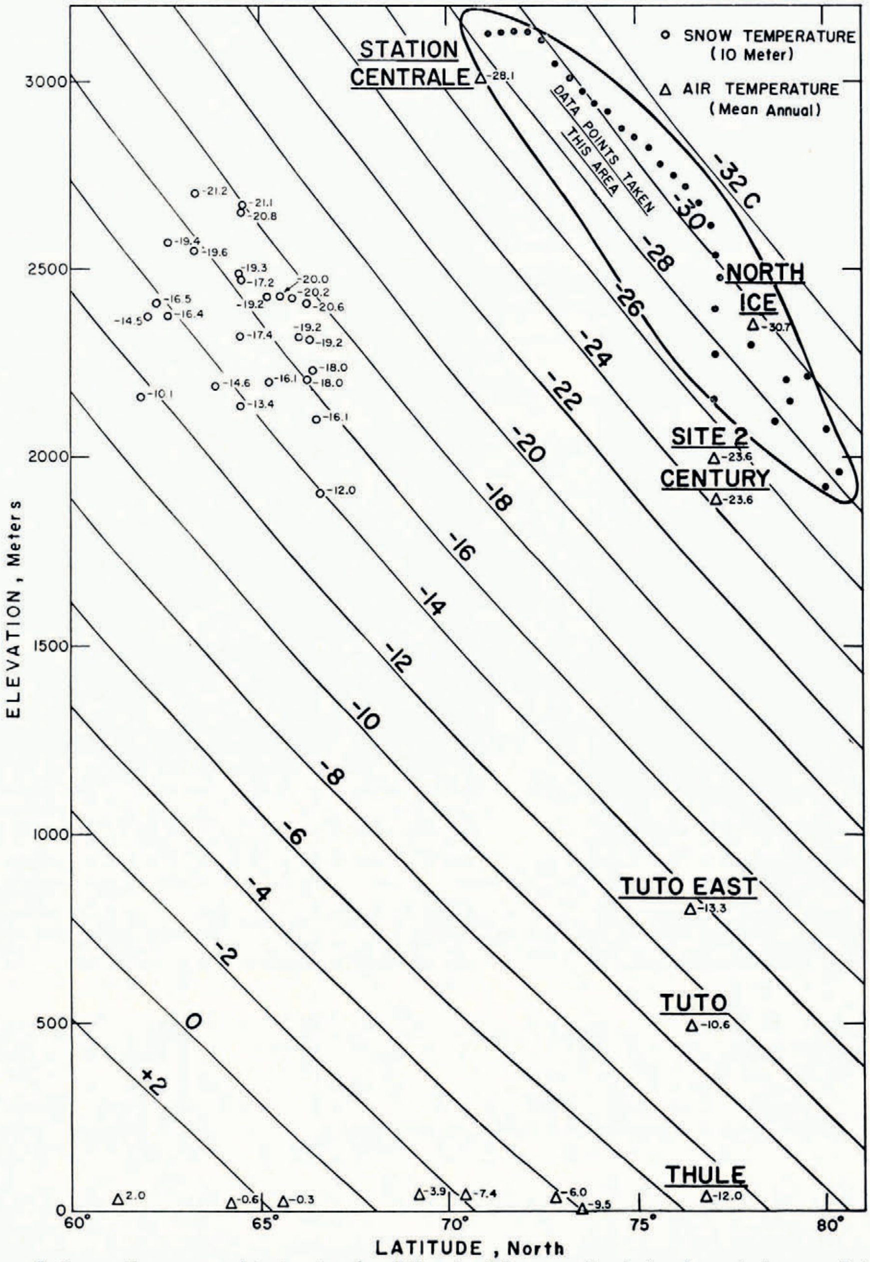

Fig. 15. Isotherms of mean annual temperature for all Greenland from equation (10) using only dry-snow facies data

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(a) Mean annual air temperatures at all coastal stations are lower than predicted by equation (10). This is not unexpected since dry adiabatic temperature control is lost and other processes become established when katabatic winds diminish on leaving the ice sheet.

(b) Mean annual air temperatures of ice-sheet stations in north Greenland and of the single station (TUTO) immediately adjacent to the ice sheet, agree quite well ($\pm 2S$) with the predicted temperatures.

(c) Measured 10 m. snow temperatures outside the dry-snow facies show systematic deviations from predicted temperatures. Figure 16 suggests that in areas of limited melt,

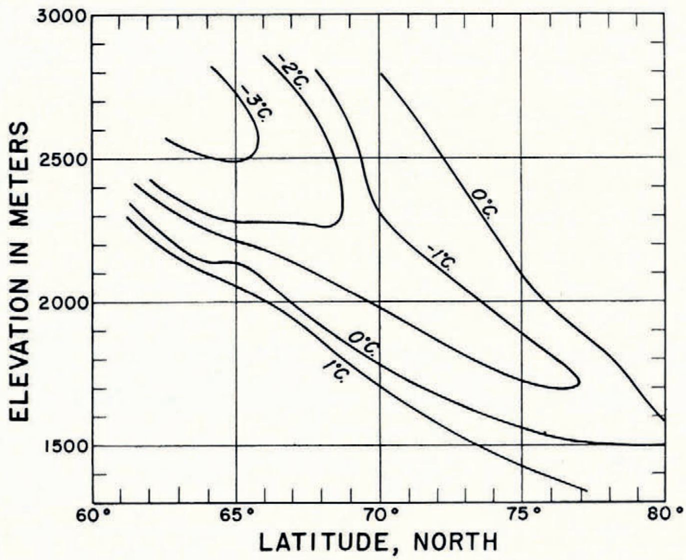

Fig. 16. Contours showing deviation of measured 10 m. snow temperature from mean annual air temperature predicted by equation (10)

10 m. snow temperatures are actually lower than the mean annual air temperature, but with increased melt, temperatures approach and then become higher than the mean annual air temperature. In south Greenland the deviations shown in Figure 16 are of such a magnitude that a climatic difference between north and south Greenland is suggested.

Antarctica: Because of the high degree of success in fitting the Greenland data with polynomial regression equations, a similar analysis was carried out with temperature data obtained from traverses in Marie Byrd Land (Anderson, 1958; Long, 1961; Pirrit and Doumani, 1961; 78 data stations) and Victoria Land (Crary, 1963; 41 data stations). The resulting equations (Table I) provide a somewhat poorer fit to the data than the Greenland data (higher S values and lower R values). Figure 17 is a plot of the isothermal surfaces determined from equation (12), indicating the control of latitude, longitude, and elevation upon temperatures in Marie Byrd Land.

Figure 18 shows the best-fitting curve for Victoria Land. It is of interest that in this case only elevation was found to be a significant factor in controlling the temperature. It will be interesting to compare these results with results obtained from the central part of East Antarctica.

ERRORS

The data used in this analysis were collected by a number of investigators over a period of some 15 yr. The following is a brief description of the possible errors and their effect:

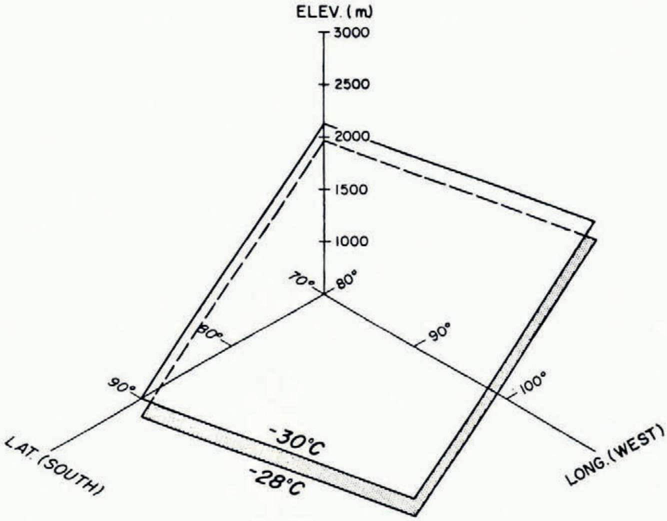

Fig. 17. The −30°C. and −28°C. isothermal surfaces for Marie Byrd Land, Antarctica, based on equation (12)

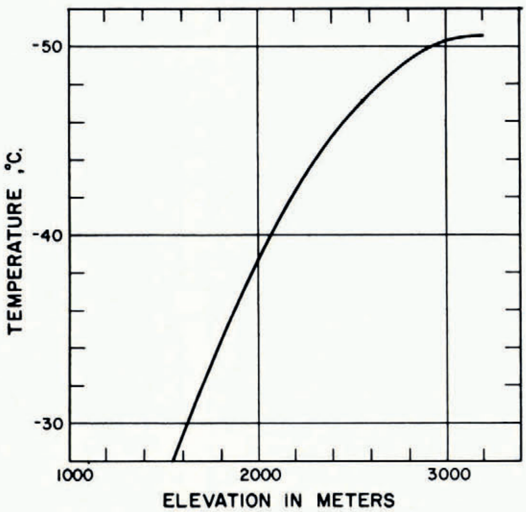

Fig. 18. 10 m. snow temperatures versus elevation for Victoria Land, Antarctica, based on equation (14)

(a) *Temperature:* The majority of the temperatures were measured with thermohms and a Wheatstone bridge in 10 m. bore holes. Some of the earlier temperature measurements were made with standard mercury or alcohol thermometers. Many of the 10 m. snow temperatures reported by Benson (1962) are calculated values from direct measurements at shallower depths. The temperatures used can be considered accurate to $\pm 0 \cdot 5$°C.

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(b) *Latitude and longitude:* Positions reported are from solar observations and dead reckoning. In general latitude can be considered accurate to $\pm 0\cdot 03^\circ$ and longitude to $\pm 0\cdot 08^\circ$.

(c) *Elevation:* Elevation determinations are almost entirely the result of barometric altimetry and as such are subject to considerable error. In general it is believed that the elevations can be considered accurate to ± 30 m.

Since errors of these types can be considered to be randomly distributed around the true values, on the average the Y_p values predicted by the regression equations should give quite satisfactory estimates of the true 10 m. temperatures.

Conclusions

Where sufficient control is available, multiple regression analysis of 10 m. snow temperatures provides a valid and highly useful method for predicting their areal distribution. The following specific conclusions seem justified:

(a) The altitudinal distribution of mean annual air temperatures over the Greenland Ice Sheet is controlled primarily by adiabatic warming of air moving down-slope under katabatic conditions.

(b) Temperatures at 10 m. in the dry-snow facies closely approximate the mean annual air temperature and temperature–elevation gradients within this zone are close to the dry adiabatic lapse rate.

(c) Temperature–elevation gradients increase to levels greater than the dry adiabatic lapse rate in the percolation facies indicating 10 m. snow temperatures warmer than the mean annual air temperature, while predicted temperatures based on the dry-snow facies equation are warmer than measured 10 m. snow temperatures, indicating that the 10 m. snow temperatures are colder than the mean annual air temperature. A need for research into the problem of heat transfer in snow undergoing limited melt is indicated.

(d) 10 m. snow temperatures in south Greenland appear to be lower with respect to their latitude and elevation than those of north Greenland.

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