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Quartermaster Research and Development Command [1]

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ABSTRACT

Study of thermal storage in the local environment (landscape and climatologic air) provides an approach to an understanding of surface-created local climates, such as those in snow-covered regions. Such a local climate, dominated by the snow surface, occurs 40 to 45 per cent of the time on the Greenland icecap, in spite of convexity of terrain. In winter, the local climate is distinguished by long-term thermal storage of 100 to 150 calories per square centimeter in a three-layer environment. This heat, conserved from periods of advection of maritime air into interstorm periods, supplies two-thirds of the heat lost from the local climate. In summer, short-term thermal storage conserves approximately 30 cal/cm^2 from the high-sun to the low-sun part of the day. During the daily cycle, the range in radiation balance is 17 cal cm^{-2} hr^{-1}; but because the regional specific heat is small, range in snow and air temperature is relatively large.

In addition to its value in applying meteorology to problems of engineering and military design and operations, the study of thermal characteristics of local climates also has value in research on the relations between surface and atmosphere. The "classic" theory of influence of snow cover on climate does not accurately describe the local climate over a snow surface, even when, as in Greenland, modifying factors are much less important than in other snow-covered regions.

1. Introduction

Observations of weather on the ice plateau of Greenland provide an opportunity to examine the classic theories of influence of snow cover on climate. The local climate formed on this high, convex surface during periods of weak advection has distinctive thermal characteristics, well illustrated by the conservation of heat from one weather type to another. Some thermal features fit into the classic theories, while others diverge from them because only two physical phenomena, low conductivity and high albedo, come into play.

The concept of local, or independent climate, occurring in periods when advection is weak, has been found useful when meteorology is applied to problems in engineering design, testing and operations. Local climates occur more frequently than is generally realized; and because they often bring extreme stress on material and great activity in hydrologic and biologic processes, they form an important part of environmental analysis of engineering problems. Accurate prediction of effects of the environment on such phenomena as melting of snow, deterioration of stored supplies, cooling power of the atmosphere, or functioning of equipment must be based on an examination of local climate. Study of local climate is also basic to the general problem of relations between surface and atmosphere, which is of great importance in meteorology.

[1] Presented at Baltimore meeting of American Meteorological Society, April 1954.

Slope of the surface, its albedo, thermal diffusivity, emissivity, roughness, concavity or convexity, degree of dissection, and cover affect the climate that prevails in the local environment, not only in the thin microclimate layer next to the surface, but often through the entire friction layer. Special types of surface — forest cover, concave topography, and snow cover — produce specific effects. There is a general qualitative knowledge of the direction in which each of these surfaces influences the local climate, but little quantitative knowledge upon which usable relations between free air and surface can be based.

Snow cover has conspicuous effects on local climate, which were worked out qualitatively some 70 years ago by Voeikov [1; 2]. His penetrating and well-illustrated analysis, amplified by other climatologists [3; 4], was immediately accepted in its entirety. It appeared in the second edition of Hann's classic textbook on climatology [5], and has come down through the years with little modification. Now, however, more information is available on the four elements he considered important, and on the ways in which their action may be modified by external factors. Voeikov's four elements are: (*a*) strong radiation from snow in the long wavelengths, (*b*) high albedo toward radiation of short wavelengths, (*c*) low thermal conductivity, and (*d*) heat required for change of physical state. These factors help determine the energy balance of the surface and thus affect the thermal level of the local climate; cool days, very cold nights, and a low heat surplus are expected.

Since the 1880's, it has been found that snow does not radiate as strongly in the long wavelengths as Voeikov had thought. Its emissivity is near that of a black body, but no higher. On the other hand, later measurements of albedo are much higher than his estimate, so that absorption of solar energy is less than he computed. Also, it has been found that the effectiveness of each factor depends on other factors in the local environment. Albedo is low, for example, in environments that favor rapid weathering of the snow and consequent change in its crystalline structure, and its importance depends on the presence or absence of other absorbing surfaces in the landscape.

Several such modifying factors exist in the Sierra Nevada of California, where the observed climate departs greatly from the classic pattern. The high frequency of anticyclonic flow at 700 mbs [6] results in subsidence. When sinking air reaches the surface, there is an appreciable import of heat in the local environment. Even more important to the regional heat economy are the long sequences of days of intense solar radiation and its efficient absorption by the landscape of dark, open pine stands and weathered snow cover of low albedo. Curvature of contours on the 700-mb surface allows a favorable heat economy; the snow cover does not produce the expected cool days or extremely cold nights, nor does it prevent generation of a large surplus of heat. In Greenland, on the other hand, external factors that might modify the influence of snow cover on local climate are different than in the Sierra. Unfortunately, the lack of upper-air observations prevents assessment of the effects of such considerations as curvature of upper-air flow, but the classic theory of influence of snow cover may be modified in the light of an examination of the local climate, with special attention to storage of heat.

2. Long-period thermal storage

Records of the Wegener expedition [7; 8; 9; 10; 11; 12; 13; 14; 15; 16] and the Victor expedition [17; 18; 19], which overwintered on the plateau, are the best published sources of our knowledge of year-round weather on the inland ice. A study of them indicates that alternation of advective and radiative weather is a major characteristic of the climate. These types, which may also be termed dependent and independent climate, or periods of storm and of local climate, exhibit a marked contrast that makes them valuable for classification. Loewe [14] expresses this contrast in graphic terms when he says that the icecap is "a battlefield where forces based on the form and nature of the surface, on dome shape and snow, fight the influences from outside which tend to trouble these native forces." Storms wipe out the local climate; but after they pass, the "native forces" of shape and particularly of surface material soon reassert their power and reestablish the local climate. The local climate is thus created by snow cover, and

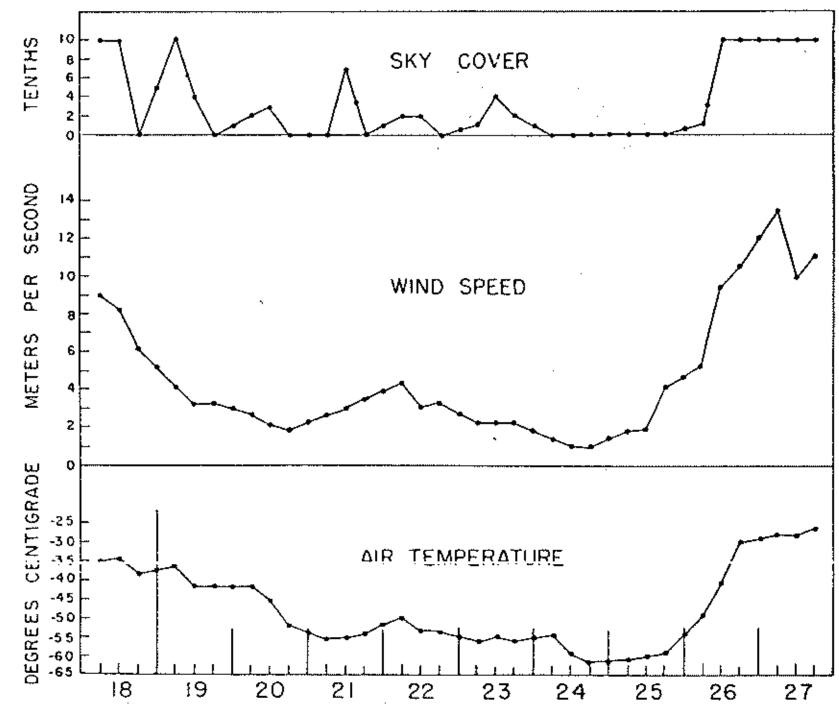

FIG. 1. Weather conditions during typical interstorm period of winter. Fall of temperature takes place irregularly, in lagged response to decrease in wind speed. From observations [10] by Wegener expedition at Eismitte (70°54'N, 40°42'W, 3000-m elevation).

after creation is dominated by it. Disagreement with Hobbs' theories [20], which claimed an almost hemispheric role for the icecap's climate, should not go to the other extreme of totally ignoring the fact of local climate on the plateau.

The frequency of interstorm periods, defined as groups of two or more days without snowfall, is found to be about the same in the observations of the two expeditions. Forty to 45 per cent of the days of winter belong to these periods. About half the days in the periods occur in groups longer than four days. Fig. 1 presents data for a typical interstorm period. These periods average much colder, quieter and clearer than the storms, as is shown by the following tabulation, computed from data in Georgi [10]:

	Interstorm periods	Storm periods
Mean wind speed (m/sec):	4.1	6.3
Mean temperature (deg C):	−48	−34
Mean cloud cover (tenths):	2.3	7.4

Frequencies of completely clear days in the interstorm periods and of overcast days during storms are both 28 per cent. Windchill, 2200 kcal m^{-2} hr^{-1} in the interstorm periods, is slightly higher than during storms, when it is about 2000.

During the periods of local climate, the most important of the influences of snow on climate is low conductivity, and hence low thermal capacity. In winter, albedo is irrelevant; no melting occurs; and longwave radiation, though the principal means of heat loss, does not exceed that emitted by a snow-free surface at the same altitude. Transfer of heat upward from the snow and underlying firn is very small; Albrecht [21] estimates it to be about 200 cal/cm² during a winter month. The low heat capacity of the snow and the slow movement of heat from below permit extreme chilling of the surface, to a temperature at which net exchange of heat by long-wave radiation is low enough to be balanced by withdrawal of heat from the natural landscape.

The chilling produces a layering of the environment into three zones: the top layer of the snow, the ground inversion, and the isothermal layer. Dimensions of the layers differ widely. In periods between winter storms, only a few meters of snow lose much heat. Radiosonde observations of the Victor expedition [17; 18] indicate that the inversion thickness averages 200 m, varying only slightly from one interstorm period to another. The isothermal layer above the inversion is about 600 m thick, and is probably thickest in late winter. In spite of the high elevation of the ground surface, nearly 10,000 ft in the region of Eismitte, cooling does not extend above 4 km above sea level. The top of the isothermal layer in the winter of 1949–1950 never reached above 600 mb; its temperature and height lie in the pattern that Wexler [22] found from radiosondes at Fargo, North Dakota (fig. 6 in his paper).

Computations of heat capacity of each layer, based on observed fluctuations in its temperature and on thermal coefficients for snow of density 0.3 [23; 24], show that thermal storage is about 50 cal/cm² in the top layers of the snow. About the same amount of heat is released as the inversion forms, and 30 to 50 cal in the isothermal layer. The physical environment thus stores from a storm into the succeeding interstorm period, 130 to 150 cal/cm².

During a storm, mixing of maritime air with the cold air of the inversion layer and the consequent warming of the snow represent an addition of heat to the local environment, that is, a filling of the thermal reservoir. After a storm, air temperature falls 15 to 20C, and snow temperature as much or more, representing rapid withdrawal of heat from storage. The fall in temperature is accomplished within the first two days following a storm. Records of both expeditions show that during the first day air temperature falls 9C, during the second day 2C, and thereafter its fluctuations are minor, probably in response to changes in flow aloft.

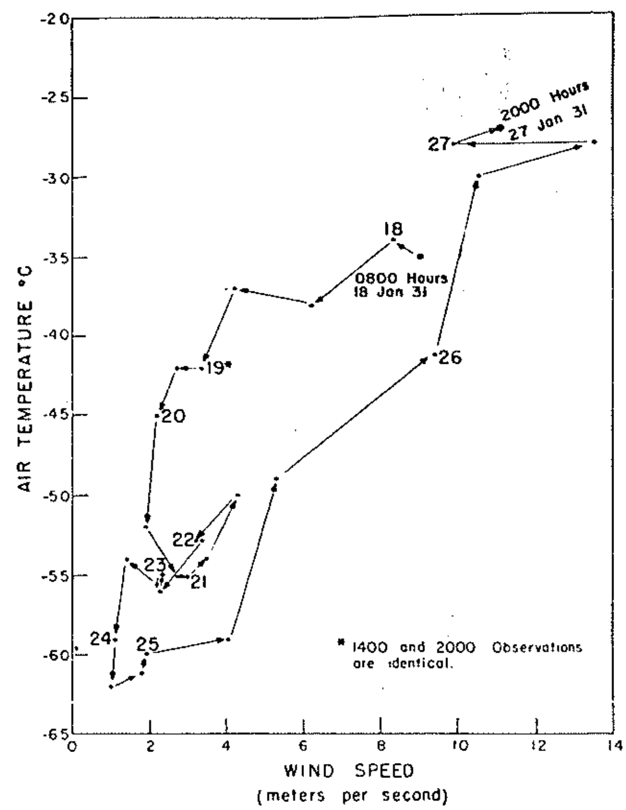

FIG. 2. Plot of air temperature against wind speed during winter interstorm period shown in fig. 1. Observations [10] were made at Eismitte at 0800, 1400 and 2000 local time; dates are shown at 1400 observations. Graph illustrates how relation between temperature and air movement differs in cooling and warming stages of interstorm period.

When temperature is plotted against wind speed, a consistent loop pattern is found (fig. 2). First, the wind drops while temperature falls only slowly; then, after the wind is low and the inversion established, temperature falls more rapidly. In the rising phase, wind picks up sooner than temperature, which is some ten degrees lower at a given wind speed than it is during the earlier phase. Rough equations for the two phases are, respectively, $t = -55 + av$, and $t = -65 + av$, where t is air temperature in degrees Celsius, v is wind speed in meters per second, and a is a coefficient with the dimensions (degrees Celsius) (seconds) $(\text{meters})^{-1}$ and numerical value three. The hysteresis shown in the graph accounts for scattering of points when observations of wind and temperature are plotted without separation of the rising and falling phases of interstorm periods.

Table 1 and fig. 3 present estimates of heat transfer and changes in storage during a typical storm and interstorm sequence in winter, to facilitate comparisons of the various modes of heat flow. Monthly totals of heat transferred in each way are also shown.

Advection contributes large amounts of heat during storms, largely by mixing of warm maritime air with the cold surface layer. The rapidity of this mixing is shown by temperature rises of the order of 40 to 50C in two or three days as, in Matthes' [15] phrase, the storm wind sweeps away the cold layer near the surface. The association of high winds with high air temperature is seen in the following summary of wind speeds in the winter of 1949–1950. When air temperature was from −61 to −85F, the 90th percentile of wind speed was 14 mi/hr; at −41 to −60F, this speed was 19 mi/hr; at −21 to −40F, it was 26 mi/hr; and at temperature above −20F, when the upper air reached the surface, the speed jumped to 37 mi/hr.

During interstorm periods, Wegener [7] mentions subsidence as replacing air lost by downslope flow. Heat from this source is advective in origin, but arrives in the local environment through a less drastic process than the wholesale mixing of a storm. Through the inversion there is transfer of latent heat also, as vapor diffuses downward to the snow surface. A photograph by Loewe [14] shows a deposit of hoarfrost 2 in deep that had a water equivalent of 0.6 cm. This deposit may be unusually large, since its formation represents an addition of more than 300 cal/cm^2 to the local environment. However, Wegener's estimate [7] that hoarfrost added at least 1 in of water per year to the accumulation on the icecap, and observations that show its frequency in fall and winter to be as high as 33 per cent, indicate that monthly inflow of heat from condensation is significant. Some condensation takes place also during storms, although probably small in amount compared with advective heat gain. The figure in table 1 for total monthly gain of heat by condensation is larger than that presented by Albrecht [21], who gives less weight to observations of hoarfrost.

Net exchange of long-wave radiation is the principal avenue of heat loss in winter, but the figures of table 1 are smaller than those given by Albrecht, who postulated that no inversion existed and hence computed a small rate of downward radiation.

After the initial cooling following a storm, equilibrium is established between loss of radiant heat from the snow surface and downward flow of radiation from the isothermal layer, plus a small amount of heat from the firn and from condensation. From radiosondes of the Victor expedition from December 1949 through February 1950 [17], the temperature of the isothermal layer during each of ten interstorm periods was found to vary from −23 to −40C. Thickness of the layer varied from 42 to 58 mb. Means were −29C and 50 mb. At this temperature and a normal carbon-dioxide content, the layer radiates downward about 11 cal cm^{-2} hr^{-1} according to Wexler [22], and will be in thermal equilibrium with a snow surface receiving no other heat that has a temperature of −57C. The air above the snow

TABLE 1. Thermal balance in winter
(calories per square centimeter)

	Storm period (4½ days)	Interstorm period (3 days)	Monthly total
Heat loss from environment			
By long-wave radiation	−100	−220	−1300
Heat gain by environment			
By condensation of vapor	10	50	250
By advection and subsidence	200	20	900
By upward flow from firn	20	20	150
Cyclical storage of heat			
In inversion and isothermal layers	−80	80	0
In surface layer of snow	−50	50	0

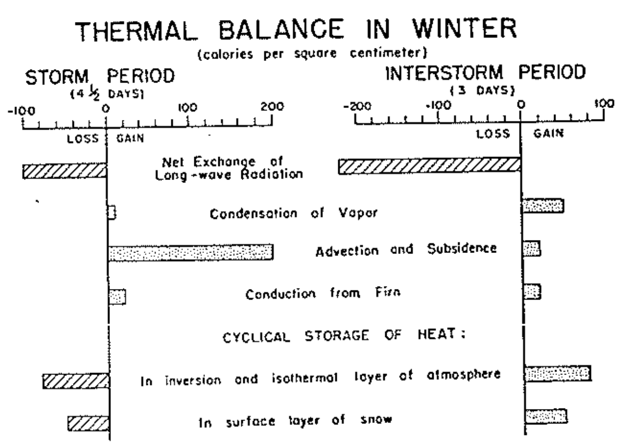

FIG. 3. Thermal balance during storm and interstorm periods at Eismitte in winter. Heat gained from advection of maritime air during storms goes into storage, from which it is withdrawn during subsequent interstorm period to meet demands of long-wave radiation exchange. Storage of heat in landscape and local air amounts to about 130 cal/cm^2 (table 1).

would be a few degrees warmer. Shelter temperature at the time of the 24 ascents in 1949–1950 averaged −44C, some 10C warmer than computed, which indicates a considerable flow of heat into the local environment at these times. The minimum temperature in each period averaged −51C, much closer to the computed temperature at radiative equilibrium. In 1950–1951, the mean temperature of the isothermal layer between storms was −35C. Surface temperature was correspondingly lower than in the preceding winter; the minima in each of the four interstorm periods averaged −58C.

In individual interstorm periods, agreement between computed and observed minimum temperature was good when wind speed at time of minimum temperature was less than 10 kn. At higher wind speeds, which ranged up to 25 kn at time of minimum temperature, observed temperatures were as much as 15C higher than theoretical snow-surface temperatures at radiative equilibrium. It is interesting that the windy interstorm periods came in mid-winter; those with lighter winds and lower temperatures came either earlier or later.

Extrapolated downward to 700 mb, the temperature of the free air in interstorm periods in 1949–1950 was −24C, compared with the January mean at this level of −25C [25].

It may be noted that net loss by long-wave radiation would be smaller than is shown if the emitting surface were only a film of snow instead of sometimes a zone of blowing snow a foot to several feet thick. When snow is blowing, a large mass is cooled, its temperature drops little, and its radiation loss is large.

Figures in table 1 differ somewhat from those given by Albrecht [21], who bases his estimates on an assumption that weather remains uniform at the monthly average value, and so does not discuss conditions during dominance of either major weather types or evaluate conservation of heat from one type to another. This storage factor, given in the last two lines of the table, represents heat stored in the three layers of the environment in each alternation from storm to interstorm period. Warmth stored in air and snow becomes the major source that supplies long-wave radiation during the following period of local climate; it makes up about 60 per cent of the supply of heat.

This thermal storage is less than that occurring where ground is bare and distinct weather types occur. However, it is larger than that found in regions having a smaller contrast between storm and interstorm periods. In the Sierra Nevada, in fact, interstorm periods in winter are somewhat warmer than storms; there is no conservation of heat from storms to periods of local climate.

If storage of heat in the local environment of Greenland were larger, as it might be in a denser substrate, such as rock, a greater loss of heat by long-wave radiation could be supported; equilibrium would be reached at a temperature higher than −60C. However, wind packs the surface snow to a density of 0.3, and its conductivity is double that of unpacked snow; thermal capacity is correspondingly increased. The continual movement of air under the inversion and the strong winds during storms thus help to produce a moderate amount of thermal storage in the environment.[2]

We may visualize the quotient, thermal storage divided by change in surface temperature, as a kind of specific heat of the environment, bearing in mind that it does not have the dimensions of a true specific heat. In the winter situation described, this quotient is 130/25, or 5 cal cm^{-2} deg^{-1}. In regions with greater thermal capacity and smaller difference between weather types, the "environmental specific heat" is at least twice the value cited.

3. Short-period thermal storage

Storage of heat in the landscape over short periods is illustrated by diurnal storage of heat in summer, between low-sun and high-sun parts of a day between storms.[3] Long-period storage of the type described for winter is negligible or lacking, because in both storm and interstorm periods the landscape disburses about as much heat as it currently receives from solar radiation. Air temperature in storms is slightly lower than in intervening periods, and there is no significant carryover of heat from one type to the other. The mean temperature of interstorm periods rises in a regular manner to a peak of −13C in late July and declines in as regular a manner, a regime that suggests that advection during storms has little effect on air temperature in interstorm periods.

Snow cover influences climate differently in summer than in winter, when low conductivity is the only important property. In summer, when insolation on days between storms is very high, albedo of the snow becomes an important factor. On these days, insolation on a horizontal surface at Thule, 5 deg lat poleward of Eismitte, totals as much as 700 to 800 cal cm^{-2} min^{-1}.[4] The clearest days are tropical in their total receipts of solar energy. Frequency distributions of daily insolation in June and July at Thule and at Blue Hill, Massachusetts, are nearly co-incident, a

[2] The records give two values for density of the surface: 0.3 for wind-packed snow, and 0.1 or less for hoar frost. Although Georgi [49] points out that a layer of frost crystals tends to insulate the air from the denser snow beneath, considerable heat may be transferred through such a layer by convection. In any case, the reduction in upward heat flow is far exceeded by the heat of sublimation released as the layer forms.

[3] The high-sun period is taken as the 14 hr from 0500 to 1900 local time, the low-sun period as the rest of the day.

[4] Personal communication from Dr. S. Fritz, U. S. Weather Bureau.

comparison that bears out Matthes' view [15] that the icecap climate is basically that of New England transposed downward on the temperature scale.

This high inflow of heat produces a high temperature in objects that have low albedo, as equipment or clothing. Many explorers, such as Rasmussen [26], have commented on the heat stress felt during exercise, although the air temperature is below freezing. The present writer has computed the surface temperature of a dark, erect object, exposed to direct radiation and radiation reflected from the snow, to be as high as 70F, even when heat is removed moderately rapidly from it by convection and by radiation exchange with the cold terrain.

A survey [6] of observations of albedo of new snow in the Arctic and in mid-latitude mountains indicates a high degree of agreement among investigators. On a day following a storm, albedo averages between 0.83 and 0.85, varying slightly in response to type of crystal and temperature at which flakes were formed.

Between storms, albedo declines as the crystals of the snow surface grow and change shape. This decline is hastened when melting occurs in the daily cycle. But melting is not a necessary condition; according to de Quervain [27] and others, re-orientation and growth of surface crystals are caused by cyclic changes in gradient of vapor pressure within the top layer of the snow. These changes, in turn, are caused by changes in surface temperature which, as will be seen later, are quite large on the icecap.

From a study of observations of albedo made by the Swiss Institute for Snow and Avalanche Research [28; 29; 30] and by the Cooperative Snow Investigations in the United States [31; 32; 33; 34; 35], the writer estimates that, by the fifth day after a snow storm, albedo of a new snow surface may decline to nearly 0.70 [36; 37]. If storms are spaced widely, albedo declines more, and the amount of heat absorbed within the local environment is significantly increased. During summer, the spacing of storms in Greenland is such that, over the interstorm periods, albedo probably averages about 0.75.

Interstorm periods in summer average a degree or two colder than storms; but most of the cold weather occurs the first night, and the days warm steadily until the 24-hr mean is higher than the storm temperature. If interstorm periods were longer, albedo could decline further than it does and a higher level of temperature in the local climate could be reached, subject, of course, to truncation if the snow-surface temperature rose to the melting point. At elevations below 6000 ft, melting occurs frequently [38] and becomes effective in restraining rise of temperature of snow and air, but at Eismitte it is rare.

The daily cycle of weather on days between storms in summer is well marked, although the sun may be above the horizon every hour in the day. U. S. Weather Bureau records from Thule indicate that the cycle of insolation is from 5 to 10 cal/cm² during an hour of low sun, to 60 or more near noon. Since a quarter of the incident radiation is absorbed by the snow, the cycle of energy input is from 2 to 15 cal. This range may be compared with Sverdrup's values at 870-m elevation in Spitzbergen [40] which, on clear days, ranged from 4 to 12 cal between low- and high-sun hours. The higher latitude of Spitzbergen (79°N) probably accounts for reduced daily amplitude.

Deducting net exchange by long-wave radiation, one finds the snow at noon to have a net gain of 10 to 12 cal cm^{-2} hr^{-1}. This estimate is confirmed by a few observations [39] made with an Albrecht radiation balance at a lower elevation (1400 m), somewhat farther south (66°N), but later in the year (6 September). Daytime observations in the sun ranged up to 8.5 cal. At night, long-wave radiation is dominant over the low sun.

The variation in heat supply causes a variation in snow-surface temperature of 15 to 20C, as Brooks [41] noted from the records of the de Quervain expedition. Wegener [7] found that the snow is 1C colder than the air at 2100 local time, 2C warmer at 1400. Its daily range is probably about 5C more than that of the air. Contact with the warming and cooling surface of the snow produces a range in air temperature of 10 to 15C. Daily maxima and minima lag the

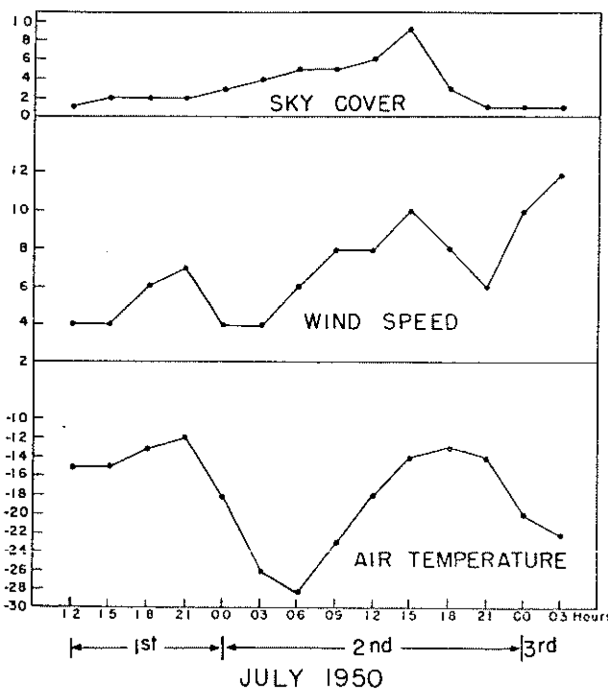

FIG. 4. Typical summer day on icecap, between storms. Delay in maximum and minimum air temperature is noteworthy. From observations [17] by Victor expedition at Central Station, near site of Eismitte. Sky cover is in tenths, wind speed in knots, temperature in degrees Celsius.

times of high and low sun by about 7 hr; the periods of most rapid rise and fall of air temperature are centered at the hours of high and low sun. A typical summer day is illustrated in fig. 4.

Change in temperature of air and snow components of the environment indicates a flow of heat into storage during the day and out at night, when absorption of radiation from the low sun does not meet the demands of long-wave radiation. From observations of change in temperature and estimates of mass of air and snow affected, the writer computed this thermal storage to be about 30 cal cm^{-2}, 10 in the snow and 20 in a shallow ground inversion. Table 2 and fig. 5 show this storage in relation to gains and losses of heat in the local climate during the high- and low-sun parts of the day.

Loss of heat by conduction downward into the firn is taken from Albrecht [21]. Absorption of short-wave radiation is calculated from observations at Thule, under the assumption of an albedo of 0.75 as discussed earlier. Net loss of heat by long-wave radiation is computed by procedures from Möller [42] and Bolz and Falckenberg [43] which, over snow cover in the Sierra, give values in good agreement with observations by Gier-Dunkle radiometers [44].

Loss of sensible and latent heat to the free air is small; the local climate is reasonably independent of the surrounding oceans, in that it exchanges little sensible or latent heat with maritime air. From Wegener's observations of evaporation during a few days and condensation during a few nights in summer [7], transfer of heat from the snow is estimated about 15 cal during the high-sun half of the day and to the snow 5 cal during the low-sun half. These days probably had unusually high rates of moisture transfer, so the values above were reduced and included in the figures for cyclic storage of heat in the surface layer of air. Gains of latent heat are also small during storm periods, in contrast to the situation in winter.

Daily storage of heat, 30 cal/cm^2, is small in comparison with other regions. Conservation of heat from day to night in summer in Potsdam, for example, is given by Albrecht [21] as about 150 cal/cm^2; and in Finland, Franssila [45] found about 100 cal/cm^2.

TABLE 2. Thermal balance during summer day between storms (calories per square centimeter).

	High-sun period (14 hr)	Low-sun period (10 hr)	Total day
Heat losses			
By long-wave radiation (net loss)	−80	−40	−120
By downward conduction to firn	−10	0	−10
By loss to the free air	−20	−10	−30
Heat gain			
By absorption of short-wave radiation	140	20	160
Cyclic storage of heat			
In the air	−20	20	0
In the surface layer of snow	−10	10	0

In spring, such conservation of heat in the snow-covered Sierra Nevada is about 65 cal/cm^2, a large fraction of which is caused by refreezing of meltwater in the top layers of the snow pack [6].

Daily range of temperature of the snow surface in the Sierra is only a third of the 15 to 20C range in Greenland. Larger thermal storage in the Sierra curtails the range of temperature; at the same time, change of state allows development of a large amount of storage with comparatively little change in temperature. Range in hourly net radiation in the Sierra is from −5 cal/cm^2 at night to about +50 by day, or 55 cal/cm^2; in Greenland, the range is from −2 to +15, or 17 cal/cm^2, only a third of the Sierra value. The combination of small range in net radiation with large range in surface temperature reflects the small thermal capacity of the environment.

Storage of heat in the environment for each degree Celsius of range in surface temperature is about 1.7 cal/cm^2, in contrast with the figure given earlier for winter, about 5 cal cm^{-2} deg^{-1}. The longer cycle of heating and cooling in winter permits a deeper penetration of the waves of heat and cold. The value of "specific heat" of the environment during the daily cycle in summer is extremely small in comparison with the figure for the Sierra, about 13 cal cm^{-2} deg^{-1}.

Snow cover in summer in Greenland has an entirely different influence on local climate than it does in the Sierra, because rise of temperature in the daily cycle is not curtailed by reaching the melting point, nor is its fall restrained by release of heat of fusion of liquid water in the snow. Change of state, as one of the means by which snow cover influences climate according to Voeikov, is inoperative in Greenland.

In contrast with winter, there is little transfer of heat in summer from the surrounding oceans to the local climate of the plateau, either as sensible heat or

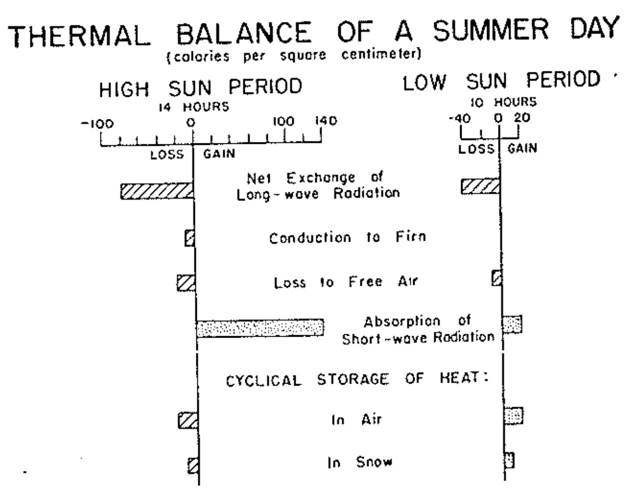

FIG. 5. Thermal balance during high- and low-sun parts of summer day between storms, at Central Station (71°N, 41°W). Excess of incoming over outgoing radiation during high-sun hours permits cyclic storage of about 30 cal/cm^2 in landscape and local air, which at night makes good the deficit of radiation (table 2).

as heat of condensation. Between storms, there is a small surplus of heat; more energy is gained from absorption of insolation by the snow than is lost by net exchange of long-wave radiation. This surplus heat is comprised of two parts: (*a*) approximately 10 cal/cm^2 in excess of the amount that flows back out of the firn at night penetrates deeper into the ice, and (*b*) approximately 30 cal/cm^2 escapes aloft, beyond nocturnal return flow to the surface, and are thus lost to the local, or climatologic, air as defined in Leighly's [46] study of heat turnover in California.

Surplus heat is thus about 40 cal/cm^2 during a day between storms, of which a typical summer month has a dozen.

It is likely that during storms in summer enough short-wave radiation reaches the lower atmosphere and the surface to meet current loss by net exchange of long-wave radiation. Possibly a small surplus of heat exists and goes to warm the firn. Even so, the total monthly heat surplus in storms and interstorm periods probably does not exceed 800 cal/cm^2, a figure that is small compared with that produced by the Sierra Nevada in spring [47]. Although insolation is about the same in the two regions, the heat surplus on the icecap is about one-sixth of that in the Sierra. It is about the same fraction of the surplus in May from bare ground in Sweden, according to Ångström [48]. The small heat surplus, which, moreover, occurs in only a brief period of mid-summer, indicates the role of snow cover in the local climate when modifying external factors are negligible in importance.

In the segment of the hydrologic cycle that passes through this environment, water substance is exported at the lowest level of energy and carries no latent heat with it. It is transferred from the region not by expenditure of thermal energy but by use of the potential energy that results from interception of falling snow by a high-altitude deposit of the precipitation of past centuries.

4. Conclusion

This article has described the local climate that develops on an elevated, convex, snow-covered surface in high latitudes. The situation is unsheltered by terrain but nevertheless harbors a marked independent climate nearly half the time, winter and summer. This climate is created and dominated by a surface that has low density, low conductivity and high reflectance, and that provides a small amount of thermal storage. In winter, heat fluxes to and from the local environment are so small that even this minor amount of storage is important; it makes up nearly half the total heat turnover between storm and interstorm periods; in summer it provides a fifth of the total heat turnover in a 24-hr day. At the low temperature of the landscape in summer or winter, radiative loss of heat is slow, so that the small mass and specific heat of the environment, when coupled with large variations in temperature, become important in the climate.

The snow cover produces a characteristic independent local climate by the chilling caused by its low conductivity, low density, and high albedo. Its influence is not significantly modified by external factors other than wind, which packs the surface to a density greater than it would otherwise have, and at times blows the snow into a zone several feet thick that emits long-wave radiation.

Surface dominance produces, in winter periods when the circulation aloft is weak, a total environment pervaded by cold — cold earth, cold air, cold sky. Although sandwiched between warmer firn and warmer upper air, the vertical dimension of this environment is still large with respect to foreign objects, which are likewise cold. From this zone, heat is withdrawn during periods when the marked independence of the local climate forbids its import from other regions. Slowness of withdrawal implies a low rate of drawdown, which in turn implies a low level of temperature.

In summer, solar radiation is intense but makes little impression on the local environment. Snow and air remain cool, although foreign objects may reach a high temperature. Heat surplus for export outside the environment is small. Daily rise and fall of the sun sets up a contrast within the diurnal cycle; but although the range of thermal elements is large, carryover of heat is small, compared to regions without snow cover or in which snow cover is modified by external factors.

The ultra-cold winter on the icecap corresponds to the cold nights specified by Voeikov, the anomalously cool summer to his cool days. Both effects arise from the low specific heat of the landscape and climatologic air. In summer, high albedo enters as a second process of importance. Change of state is of no importance as a climate-forming influence; and since its effect is to increase the specific heat of the environment, the climate conforms more closely to the classic mold than if melting and refreezing were common occurrences.

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