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Climate Change over the Polar Ocean

II. A Method for Calculating Synoptic Energy Budgets[1]

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With 3 Figures

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Summary

The objective of this method is to obtain numerical values for all terms in the heat balance equations for the atmosphere and the earth's surface. The method is to calculate complete energy budgets using synoptic data. Once a certain number of circulation types have been investigated, over different regions, it may become possible to discuss the influence of particular synoptic events on the climate and general circulation. The model can be used for grid points, or for single stations, or for geographical areas.

Zusammenfassung

Klimaänderung über dem Polarmeer

II. Eine Methode zur Berechnung synoptischer Energiebilanzen

Es wird eine Methode entwickelt zur Gewinnung numerischer Werte für alle Glieder der Wärmebilanzgleichungen für die Atmosphäre und die Erdoberfläche; die Methode beruht auf der Berechnung vollständiger Energiebilanzen auf Grund synoptischer Daten. Wenn dann einmal eine gewisse Anzahl von Zirkulationstypen für verschiedene Gegenden untersucht sind, kann es möglich werden, den Einfluß spezieller synoptischer Vorgänge auf das Klima und die allgemeine Zirkulation abzuleiten. Dieses Modell ist auf Netzpunkte, geographische Gebiete, aber auch auf einzelne Stationen anwendbar.

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1. Introduction

Most of the published energy budget calculations fall in two main categories. The first includes those studies which consider climatological conditions over large areas, such as whole hemispheres or the globe. Physical processes tend to be obscured in such studies, because the climatological averages conceal the great variation in individual values. Discussions of individual budget terms over short periods have illustrated this (GAGNON [10], VOWINCKEL [36], CHISHOLM [8]). The second category includes those studies which are based on calculations for individual stations. The problem in these investigations is to determine the degree to which the single spot is representative for a large area.

The greatest difficulty in explaining the energy budget on a large scale is the fact that climatological calculations eliminate extreme conditions. Many of the relevant transfer processes operate in "spurts", thereby reducing the significance of mean values. Atmospheric heat advection is an example. It ranges from high positive to high negative values, often with a small net transport. Evaporation and turbulent heat flux are also good examples. A change of an order of magnitude over short periods is not unusual. It therefore seems that it should be desirable to calculate complete energy budgets on a synoptic basis. Such an approach must be more suitable than climatic calculations in attempts at elucidating the interrelation between energy budgets and the general circulation, and in answering questions about climate change. It is quite possible to investigate an individual term in the energy budget and examine how its variation will influence the balance, as was done for radiation by VOWINCKEL and ORVIG [40]. The results must, however, remain tentative until account can be taken of the intricate interactions which automatically follow the alteration of a single budget term.

The most promising method is to examine various real synoptic situations, which furnish examples of highly diversified energy budgets and individual components. The first attempts at such studies were made by KORB et al. [18], although they were generally restricted to the radiation budget. In the present examination all terms are included.

Synoptic calculations can either include data from the network of climatological stations or restrict the data to synoptic observations only. The first alternative will supply more stations and, in certain cases, additional parameters such as ground temperature, pan evaporation, cloud density. On the other hand, the published material differs from one country to another, and only a small part of the wealth of information is readily available.

Synoptic observations have the advantage that uniform information is available over a world wide network. It is quite clear, however, that the synoptic code was not designed for energy budget considerations. Important information is not included, such as ground conditions and certain characteristics of clouds. Nevertheless, the present method is based almost exclusively on synoptic observations. Thereby the method should have wider possibilities of application.

2. Data

All calculations are done for grid points. The numerical grid used by the Canadian Meteorological Service was chosen, the area covered includes the northern hemisphere down to 45^0 N, with an addition down to 30^0 N over the Gulf of Mexico. Upper air values of temperature, dew point and height are obtained from the numerical analysis of the Canadian Meteorological Service for the following four pressure levels: 850 mb, 700 mb, 500 mb and 300 mb. No dew point values are given at 300 mb in the Canadian analysis, in the present method the moisture content at 300 mb is assumed to be 15 % of that at 500 mb. Separate analyses are done for surface and cloud parameters, and grid point values are then extracted from the maps of each element.

The synoptic observations of sea surface temperatures show a considerable scatter. It was therefore decided to find, for each reporting ship, the deviation of its sea surface temperature from the mean temperature as depicted on maps published by the British Meteorological Office [4]. The obtained deviation field is analysed. Grid point values are extracted and used together with the mean temperature, to arrive at the actual sea surface temperature for the particular time.

A number of other variables are required.

2.1. Albedo

Land surface albedo depends on soil type, vegetation and snow cover, particularly snow depth and extent. Both of these are given in the synoptic observations. The surface albedo for a given snow amount depends on the type of vegetation. Kung et al. [20] have shown that albedo increases markedly with greater snow depth up to about 5 inches thickness, whereupon the albedo remains nearly constant. The actual value with this maximum snow depth varies with vegetation, and KUNG et al. presented a map for North America showing albedo values under maximum snow conditions. In the present method it is assumed that these values represent a snow depth greater than 5 inches.

The albedo of snow free ground is shown in KUNG's July map. No seasonal albedo change is included for snow free ground.

Albedo values for maximum snow cover are not available for Eurasia. The January albedo map, published by MUKHENBERG [26], was taken as representative for maximum conditions. The July map of the same author was taken for snow free conditions, and grid point values were obtained for highest and lowest albedo over land. For sea ice areas the values published by LARSSON and ORVIG [21] were taken for January and July.

The albedo often varies sharply over short distances. Values attributed to particular grid points were therefore obtained by taking the average over a 25-point sub-grid, located with the actual grid point in the centre and reaching a distance in both directions equal to the grid spacing of the main net. The actual grid values of maximum and minimum albedo are used to calculate the albedo for a particular time, considering also the synoptic snow observations, as follows:

$$ALB = AF + [(AMS - AF)\,ETA],$$

where AF = albedo of the snow free surface, AMS = maximum albedo with snow cover, ETA = coefficient, which after KUNG [20], depends on snow depth SN:

SN	0	1	2	3	4	≥ 5	inches
ETA	0.0	0.55	0.80	0.90	0.95	1.00	

2.2. Clouds

The synoptic cloud information is generally insufficient. Optical satellite pictures are used to obtain some indication of cloud density. The cover of "high brightness" (the two classes named 1 and 2 in a study by ABER [1]) is determined from the photographs, and expressed as a fraction of the area centered on the grid point and having sides equal to the normal grid distance. The bright areas are assumed to be covered by thick clouds, and the remaining cloud cover is assumed to be thin.

The grid point values of fraction of thick cloud (FNS) are available only once each day and are arbitrarily ascribed to both synoptic observations for the day.

2.3. Vegetation

Evaporation depends greatly on the vegetation type. A method was used, similar to that employed when obtaining albedo values, to ascertain the percentage frequency of vegetation types for each square, containing 64 points, centered on the grid points. The following vegetation types were used: desert, grassland, mixed shrub-grass, shrub, grass and forest, broadleaf forest, mixed forest, taiga, alpine tundra, arctic tundra.

For each grid point, ten percentage values are given, adding up to 100 %. The data ware compiled from HALLIDAY [13], USSR Academy of Sciences [34], and a map edited by KÜCHLER [19].

2.4. Elevation

The height above sea level for each point was obtained from data supplied from a personal study by Professor B. E. O'REILLY of McGill University.

2.5. Sea Ice

The distribution of sea ice was taken from the U. S. Navy Hydrographic Office [33], corrected in the North American sector by aerial survey data supplied by the Canadian Meteorological Service.

2.6. Ground Heat Flux

As an aid in the determination of heat flow from the ground, a map was prepared of the annual range over land surfaces of average daily temperature. Data from the British Meteorological Office [5] were used, and grid point values were extracted from the map.

2.7. Duration of Daylight

The period of daylight in each 24-hours was determined for each grid point for use in the calculation of solar radiation.

2.8. Fields Available

Grid points values of the following fields are available for the computations:

Constant fields: Maximum and minimum albedo, frequencies of vegetation types, elevation, sea ice distribution, extraterrestrial radiation, duration of daylight.

Variable fields: Temperature, dew point and height of the 850 mb, 700 mb, 500 mb and 300 mb surfaces, temperature and dew point at the surface, amount of low, medium and high cloud, as observed from the ground. (When high and medium clouds are present simultaneously, their amounts are described as 0.5 [N-NL], where NL is amount of low clouds and N is total cover), surface wind speed, sea surface temperature, height of low clouds, snow cover, satellite brightness, precipitation during last 24 hours (for North America only).

3. Method

3.1. Short Wave Radiation

The desired components of solar radiation are: SAA = short wave radiation absorbed in the atmosphere and SGA = short wave radiation absorbed at the surface.

Neither of these terms can be observed directly. The atmospheric absorption is best determined by first calculating the absorption under clear sky and then correcting this value for the cloud effect. A radiation diagram by YAMAMOTO and ONISHI [41] permits the calculation of absorption for various layers in the atmosphere. However, the present method is concerned with the total absorption, and it is more convenient to use the results of HOUGHTON [15]. It shoud be borne in mind that the results obtained by the two methods differ slightly, especially with high water vapour content. The following figures show the percent of incoming solar radiation which is absorbed in the atmosphere according to the two methods:

Precipitable water, cm	0.5	1.0	2.0	3.0	4.0	5.0
Yamamoto	0.08	0.09	0.12	0.12	0.13	0.14
Houghton	0.07	0.09	0.12	0.14	0.16	0.17

For the cloud absorption, use was made of calculations performed by KORB et al. [18], who found the increase in total absorption under overcast sky to be dependent on the cloud thickness and height above ground, as well as on the drop size distribution. Only the height above ground is available synoptically. By using KORB's results it is possible to estimate the following approximate values for ε, the increase in absorption over clear sky conditions, caused by overcast sky:

	ε
Thick, low clouds	1.35
thin, low clouds	1.11
thick, middle clouds	1.26
thin, middle clouds	1.17

In order to obtain the amount of these cloud groups, use is made of synoptic cloud amount observations (from the ground), and the amount observed of bright (i. e., thick) clouds from the satellite

(FNS). These two types of observations are not simultaneous. It is therefore assumed that the ground observations give the correct total cloud amount (N) and, as the amount of thick clouds is given by FNS, the amount of thin cloud will be N-FNS. It is further assumed that, when more than one cloud type is present, the satellite cloud amount observation (NS) refers initially to low cloud (NL). Only if $FNS > NL$ will thick medium cloud be assumed to be present.

In order to calculate SAA, the solar radiation at the top of the layer must be known, i. e., the extraterrestrial radiation minus ozone absorption (the top of the atmospheric thickness considered in the present method is at 300 mb). Synoptic values are not available for ozone distribution, and a constant reduction factor is used for ozone absorption:

$$SE = SE' \, [1 - (0.0214 \cdot AM)],$$

where SE = solar radiation at 300 mb, SE' = extraterrestrial radiation, AM = air mass (measure of path length).

No water vapour absorption is considered above 300 mb (since few observations are available). A compensation for this may be had in using HOUGHTON's absorption values rather than those of YAMAMOTO.

TWOMEY et al. [32] have shown a relation between cloud thickness and albedo for various drop sizes. Using a liquid water content of 0.5 gm cm^{-3} for Cu clouds and 0.3 gm cm^{-3} for stratiform clouds, the following relation can be obtained:

Cloud thickness (m)	100	200	300	400	500	600	700
Albedo %							
Cumuliform clouds	0.55	0.67	0.72	0.75	0.77	0.79	0.80
Stratiform clouds	0.41	0.60	0.67	0.71	0.73	0.75	0.77

Albedo values may also be obtained from optical satellite pictures, provided that they are of a certain quality, as shown by CONOVER [9] and ABER [1]. Such values are areal averages, however, and corrections must be made over regions with clear sky. If the surface (ground) albedo and the cloud amount are known, the cloud albedo can be obtained by:

$$ASA = N \cdot ACL + (1 - N) \cdot A,$$

$$ACL = \frac{ASA}{N} - \frac{(1-N) \cdot A}{N} = \frac{1}{N} \cdot [ASA - (1-N) \cdot A],$$

where ACL = cloud albedo, ASA = satellite albedo, N = cloud amount, A = surface albedo.

Using the relation between cloud thickness and albedo, the values of ε can be obtained (increase in absorption over that under clear sky conditions), using results from Korb et al. [18]. This work indicates that the total absorption of solar radiation increases by a factor 0.052 for every 100 m cloud thickness. The method causes difficulties because the satellite pictures are not of comparable brightness, and a calibration would have to be performed frequently. At

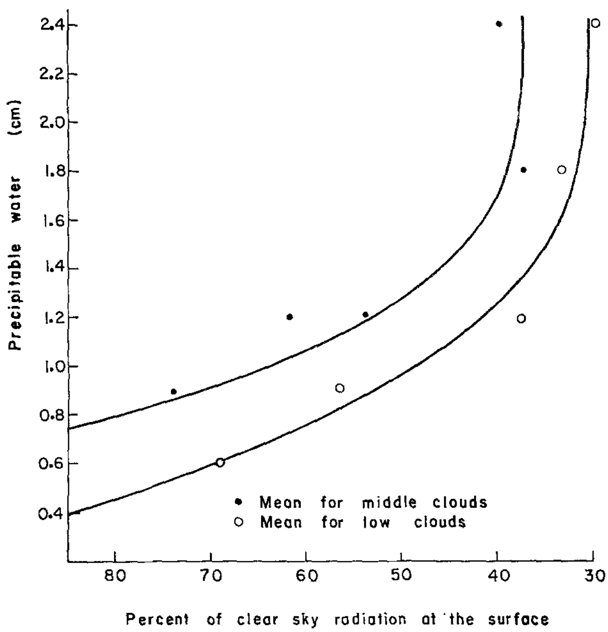

Fig. 1. Depletion of solar radiation by various clouds

present a required number of synoptic (ground) stations is not available in all regions to make this possible. If this lack is overcome, the method is promising.

The calculation of *SGA* (solar radiation absorbed at the surface) may perhaps best proceed along similar lines, i. e., first the determination of clear sky radiation. In this method the scattering coefficients of Houghton [15] were used. The second task is to determine effects of clouds, which are even more important for *SGA* than for *SAA*. This is so because the bulk of solar energy is absorbed at the surface, and even small mistakes in cloud depletion will have large numerical results.

Various examinations (KORB et al. [18], TWOMEY et al. [32]) have shown that, for a given cloud thickness, the depletion is a function of the cloud drop distribution. It is, in turn, probably related to the water vapour content of the air. If so, the geographical variation in depletion, reported by VOWINCKEL and ORVIG [38] can be readily explained. It also becomes evident that the same synoptic cloud type will exercise different influences on the radiation, in different regions. To obtain a measure of this relation, VOWINCKEL's and ORVIG's results were plotted against the mean precipitable water content, as given by the British Meteorological Office [6] in Fig. 1.

The curves for low and middle clouds are essentially parallel. It thus seems justified to use the amount of precipitable water as a first indication of the depletion factor. The data further indicate that, if the precipitable water amount is greater than 2 cm, the depletion remains practically constant. The first correction factor is, therefore, the depletion with 2 cm precipitable water. For this purpose, the averages of results from Dartmouth (Halifax) and Blue Hill were used, as reported by VOWINCKEL and ORVIG [38].

Cloud type	St	Sc	Ac	As	Ci
Solar radiation, per cent of clear sky radiation	24	39	47	38	85

Cloud thickness has not been considered so far. Use was made of KORB et al. [18], who give values for various thickness. It was further assumed that the mean values for Dartmouth and Blue Hill are made up of both thin and thick clouds. The occurrence of precipitation can be taken as indication of thick clouds. Hourly observations are available from Dartmouth for 1959—1961, and the fraction of time with precipitation, of total time with low clouds, can be seen from the following:

Month	J	F	M	A	M	J	J	A	S	O	N	D	Year
Per cent of time with low clouds, which had precipitation	33	35	35	33	19	21	14	15	16	21	30	29	25

It is thus seen that a considerable portion of the low clouds at Dartmouth are thick clouds. Using the mean depletion values for Dartmouth and Blue Hill, a transmission factor of 11 % for rain clouds (after KORB), and the percent occurrence of thick clouds, a transmission factor can be obtained for thin clouds. The following expression was finally used to calculate the transmission coefficients (D):

$$D = C_1 + (2.0 - ppw) \cdot C_2,$$

where ppw (precipitable water content) was set equal to 2 cm if it in reality had a value greater than 2 cm. The constants C_1 and C_2 have the following values:

Clouds	Low, thick	Low, thin	Medium, thick	Medium, thin	High
C_1	0.11	0.38	0.27	0.43	0.85
C_2	0.22	0.19	0.18	0.14	0.04

The final item required for the determination of SGA is the surface albedo. For land areas it was discussed above, for ocean areas it depends on solar altitude, the state of the sea and the ratio of direct to diffuse radiation. For diffuse radiation alone the water surface albedo is independent of solar elevation and state of the sea. Many different values are given for diffuse radiation, generally a water albedo of 6 % to 8 % seems to be well documented (ROBINSON [28]). The same author has shown that the albedo for direct radiation on a smooth water surface increases sharply with decreasing solar altitude. ROBINSON shows a comparison between calculated results for a rough and a smooth surface, which indicates that the albedo of a rough sea does not increase with decreasing solar elevation, but rather grows less. There may, however, be doubts about the assumptions made about the form of the waves.

Ocean albedo has been measured from aircraft with low solar altitudes (BOILEAU and GORDON [3]). Values are indicated of 10 % for a solar elevation of 13°. If it is assumed that about 30 % of the radiation was diffuse, with an albedo of 8 %, the albedo for the direct radiation would be 10.8 %. With a diffuse albedo of 6 %, the direct albedo would amount to no more than 12 %. It would seem unrealistic to use the high albedo values usually given for a smooth water surface at low solar elevations.

Test calculations were done, using smooth water albedos for the direct part of the solar radiation, and 8 % for the diffuse radiation. The values obtained for SGA were very low at latitudes north of 50° N. The annual energy budget of the Norwegian-Barents Sea, for example, would show a sharp decrease. The total energy budget for this area, calculated by VOWINCKEL and ORVIG [39], was characterized by a deficit of heat rather than a surplus, and this deficit would be increased by the acceptance of a higher surface albedo. Based on such considerations it was decided to disregard smooth water albedo values and to use a uniform ocean surface albedo of 8 %.

With these various coefficients and assumptions, the solar radiation absorbed in the atmosphere and at the surface, SAA and SGA, can be calculated, using the following relations:

$$SAA = (SE \cdot ABC) \cdot [1.35\,HNL + 1.11\,LNL + 1.26\,HNM + \\ + 1.17\,LNM + (1.0 - NX)],$$

where SE = solar radiation at 300 mb, ABC = absorption coefficient for clear sky, HNL = amount of thick, low clouds, LNL = amount of thin, low clouds, HNM = amount of thick, medium clouds, LNM = amount of thin, medium clouds, $NX = NL + NM$ (sum of low and medium cloud cover).

$$SGA = \{[SE - (SE \cdot ABC)] \cdot (1.0 - DEC)\} \cdot [D_1 \cdot HNL + D_2 \cdot \\ \cdot LNL + D_3 \cdot HNM + D_4 \cdot LNM + D_5 \cdot NH + (1.0 - N)] \cdot (1.0 - A),$$

where DEC = depletion factor for clear sky scattering, $D_1 \ldots D_5$ = transmission coefficients for various cloud types, NH = amount of high cloud, N = total cloud amount, A = surface albedo.

Since SE depends on solar elevation, which changes during the day and is different at different longitudes at the same synoptic hour, it becomes necessary to use a very dense synoptic observational network in order to obtain accurate values for absorbed solar radiation for particular points and days. Even with such observations, the computations would be unduly time consuming. In the present programme the following procedure was adopted: SAA and SGA are calculated with available observations for a certain synoptic hour at all grid points, for all solar elevations of the particular day at the individual point, i. e., the solar radiation is calculated, which would have been absorbed if the synoptic conditions remain unchanged for 24 hours.

Such calculations are performed for 12-hour intervals. The 24-hour average is obtained by:

$$\overline{SAA} = 0.25\,SAA_1 + 0.5\,SAA_2 + 0.25\,SAA_3,$$

where SAA_1 = the value of SAA at the beginning of the 24-hour period, SAA_2 = the value of SAA for the middle of the period, SAA_3 = the value of SAA at the end of the 24-hour period. — A similar method is used to obtain SGA.

Calculations for a particular hour are unimportant in large scale energy budget considerations, although the hourly synoptic observations might give more accurate numerical values. The shortest practical period is 24 hours, as diurnal variations are then cancelled.

It is possible to check the values of *SGA* against radiation observations in North America. Although these are not available on a synoptic basis, the daily values are published. A comparison for one day showed a reasonably good agreement, with the observed values expressed as percent of the calculated ones ranging from 91 % to 106 %. The average of 14 points was 96 %.

3.2. Long Wave Radiation

For the long wave terms, use is made of the method and values of KONDRATIEV and NIILISK [17]. When the temperatures and water vapour distribution are known, no additional assumptions are needed for the calculations of clear sky values. A comparison was made between calculated values for clear sky and measured atmospheric back radiation at Resolute, North West Territory. The correspondence was within 5 %.

A number of points require consideration in the calculation of the long wave fluxes:

3.2.1. Conditions above 300 mb

Temperature and moisture observations are generally unavailable above 300 mb. It was decided to stipulate a linear change of temperature from that at 300 mb (*T* 300) to the lowest value above, assumed to be -56.6^0C at the top of the atmosphere (*TO*). A linear decrease of moisture is assumed also, from *TD* 300, the dewpoint at 300 mb, to a dewpoint of -90^0C at 0 mb (*TDO*).

3.2.2. Surface Temperature

The terrestrial radiation is determined by the surface temperature. The sea surface temperature, *TS*, is used over ocean areas, but for snow free land areas the air temperature was initially used for the temperature of the radiating surface. The difference between the two remains small as long as the short wave radiation is slight, and there is vegetation cover and sufficient moisture to maintain potential evapotranspiration.

There is, at present, no means of observing the emitting surface temperature directly over a large area, under all conditions. Satellite observations may be used, provided clouds and impurities in the air do not interfere, but the temperature of a certain point would be observed only once per day and would not reflect the daily cycle. An error of 1 degree in the surface temperature is approximately equal to 10 ly day^{-1} in the value of the terrestrial radiation. An indication of likely temperature differences between 160 cm and the surface is given below, after Mather and Thornthwaite [24], for July at Point Barrow, Alaska:

Difference, T deg (0—160 cm)	+11.9 to +9.0	+8.9 to +6.0	+5.9 to +3.0	+2.9 to 0	Negative
per cent frequency, 8.30 (am)	3	19	36	35	7 %
per cent frequency, 13.30 (pm)	11	45	41	3	0 %

It is seen that the air temperature is not a sufficiently accurate indicator of the surface temperature over snow free land. The air temperature is therefore used in a first-guess calculation of the whole energy budget. As it does not result in a balance, the necessary change to obtain a balanced budget is distributed proportionately among all terms depending on the surface temperature. Thus a more accurate surface temperature is obtained and a better value for the terrestrial radiation from land areas.

3.2.3. Cloud Height and Amount

Calculation of atmospheric back radiation with clouds requires knowledge of the cloud height and amount. The synoptic reports of height of low clouds are used, although this will tend to overestimate the downward long wave radiation as the height of the lowest clouds is given in cases where several layers of low cloud are present. Generally, the error will remain small, however, as the difference in cloud temperature will be slight.

The height of middle clouds is assumed to be at 700 mb, or 150 mb above the ground, whichever is greater. For high clouds a uniform height of 500 mb is assumed. Low and middle clouds are assumed to radiate as black bodies, and Ci as 0.5 of the black body value.

For cloud amounts, the surface observations are regarded as representative. With one or two cloud layers, the information given by N (total cloud amount) and N_h (amount of cloud whose base height is reported) is sufficient for a definite determination of each layer. Most of the cloud observations show this group. If there are three cloud layers, the low cloud amount is still unambiguous. Equal weight is given to the two upper layers:

$$NCH = NCM = 0.5 \cdot (N - N_h)$$

3.2.4. Radiation Loss upward at 300 mb

For the determination of $L\,300\uparrow$, under cloudy or overcast conditions, information is necessary about height and amount of clouds as seen from above. Optical satellite observations are the only additional information available. The following assumptions are made:

a) Clouds appearing in ABER's [1] classification as 1 or 2 are regarded as thick clouds.

b) Low, thin clouds are taken to be 30 mb thick.

c) All thick clouds are assumed to have a top at 500 mb.

d) The satellite pictures do not coincide with the time of surface observations. Therefore, those cloud amounts are disregarded by which *FNS* (fraction of thick cloud) is greater than the combined low and medium cloud amount as observed from the surface.

It is apparent that the determination of $L\,300\uparrow$ is the most uncertain part of the long wave calculations. Rather than attempt improvements to the indirect method, it is probably best to await satellite long wave observations which should soon become routinely available. The first observations of extended coverage have been published by RASCHKE et al. [27]. With such observations, $L\uparrow$ at the top of the atmosphere is known and calculations become redundant. In any case, the cloud top temperature, and hence height, would be known and calculations therefore more solidly based.

3.3. Advection

Calculations of advection of sensible and latent heat are carried out for 300, 500, 700, 850 mb and the surface layer. Only horizontal air motion is considered, and the formulae used are (per unit mass):

$$\text{Sensible heat} = c_p \cdot u \cdot dT/dx + c_p \cdot v \cdot dT/dy$$
$$\text{Latent heat} = L \cdot u \cdot dq/dx + L \cdot v \cdot dq/dy,$$

where c_p = specific heat at constant pressure, u, v = horizontal wind speed in the x and y directions, L = latent heat of condensation, q = specific humidity.

u and v are calculated from the height field of the isobaric surfaces. The values for u, v, ΔT and Δq are obtained over two grid lengths, and the resulting advection value is ascribed to the centre point.

The surface friction layer is assumed to be 80 mb thick. The wind direction at the surface is taken to be that of the gradient wind, backed by 10° over the ocean and 30° over land. The surface wind speed is obtained from an isotach analysis based on synoptic observations of wind speed. Above the friction layer, gradient wind velocity is assumed. The advection in the surface layer is obtained by the average wind velocity and surface values of temperature and moisture.

The total advection is obtained by weighting the individual layer's advection according to mass, considering the surface pressure, respectively the height of the ground above sea level:

$$AD = 0.5\,(A\,300 + A\,500) \cdot 200 + 0.5\,(A\,500 + A\,700) \cdot 200 + \\ + 0.5\,(A\,700 + A\,850) \cdot 150 + 0.5\,(A\,850 + ADS) \cdot \\ \cdot [(PS - 80) - 850] + ADS \cdot 80,$$

where AD = total advection, ADS = advection in the surface layer, per unit mass, A 850, A 700 etc. = advection at each level, per unit mass, PS = surface pressure.

This method disregards the vertical component of advection. No special calculations are made for other than the horizontal part of advection. If the values of all the other energy budget terms are accepted, the vertical advection component of sensible heat can be obtained from the energy balance equation for the atmosphere:

$$AV = SAA + L\uparrow + QS + AS + DSTAS - L\downarrow - L\,300\uparrow + PR,$$

where AV = vertical component of sensible heat advection, $L\uparrow$ = terrestrial radiation, QS = turbulent transport of sensible heat, AS = horizontal sensible heat advection, $DSTAS$ = change in sensible heat storage in the atmosphere, $L\downarrow$ = atmospheric back radiation, PR = latent heat equivalent of precipitation.

3.4. Storage Change

The heat storage changes in the system take three forms: change in sensible and latent heat storage in the atmsophere, and change in sensible heat storage in the ground. The first two can be readily obtained from observations of temperature (T) and dewpoint (TD) at the surface and at 850, 700, 500 and 300 mb, using similar weighting factors as for advection, except that no special surface friction layer is considered:

$$STAS = T\,300 \cdot 100 + T\,500 \cdot 200 + T\,700 \cdot 175 + T\,850 \cdot \\ \cdot [(PS - 850) \cdot 0.5 + 75] + TA \cdot (PS - 850) \cdot 0.5,$$

where $STAS$ = sensible heat storage in the atmosphere, TA = surface air temperature.

For the heat storage change in the ground, the heat capacity is required, as well as the temperature change.

Soil temperature profiles are observed at a number of agricultural stations, but an areal analysis is not possible. Also, soil properties vary markedly over short distances. It therefore becomes necessary to introduce an approximation, based on the physical characteristics of soil and vegetation and the available synoptic information. Inaccuracies will certainly be introduced, but the magnitude of ground heat storage change (*DSTG*) is generally small, and the error should remain unimportant except during periods of thawing of the ground in areas where the soil freezes during winter. This has been shown by VOWINCKEL [37] for Resolute, N. W. T., and Ottawa.

It was judged most practical to begin the calculation on the assumption that the climatological value of *DSTG* is given for every day, and subsequently introducing corrections for the special conditions experienced on a particular day. Average monthly values for *DSTG* are available for a few locations (SCHREIBER [29], ALBRECHT [2], CARSON [7], VOWINCKEL [37]). The annual average daily value depends on the ground temperature amplitude, thermal conductivity (k) and heat capacity ($\varrho \cdot c$) of the ground. The amplitude of the temperature wave at the surface can be defined clearly and is a fixed value for each locality. However, the thermal diffusivity of the soil ($K = k/\varrho \cdot c$) varies with location as well as in time, because of the influence of soil water content. Also, K often varies with depth.

Generally, values for K are not available. Those that are reported, from agricultural stations, can be expected to refer to a grass covered soil column, and conditions such as rock, swamp or muskeg are not represented. Such extreme values for K are unlikely to be observed and reported.

Significant variations in soil water content may, however, occur in the reported values of *DSTG*. Therefore, if the average value of *DSTG* is plotted against the amplitude of the air temperature (which is taken to be representative for the average soil temperature), a certain scatter should be expected. However, Fig. 2 shows that it is possible to establish a relation between the temperature amplitude AT and $\overline{DSTG}$:

$$\overline{DSTG} = AT \cdot 0.584,$$

where AT = amplitude of air temperature in deg C, $\overline{DSTG}$ = average daily value of ground heat storage change in ly day^{-1}.

The variation of ground temperature, and hence of *DSTG*, depends on the vegetation cover. Values for different cover types were reported by GREENE ([12], after JEN-HU-CHANG [16]):

	Ground temperature amplitude (average 0—180 cm depth)
bare field	17.1 deg C
broomsedge field	14.5 deg C
loblolly pine	14.0 deg C
short leaf pine	12.5 deg C

Since the vegetation cover affects the air temperature similarly, the influence is only important where the air temperature observations are taken over a plant cover which is not representative for the wider surroundings. It is difficult to make assumptions in this respect, and the influence is not considered in the present method.

So far, the average *DSTG* has been discussed. Values for the individual months must be determined. As the temperature wave

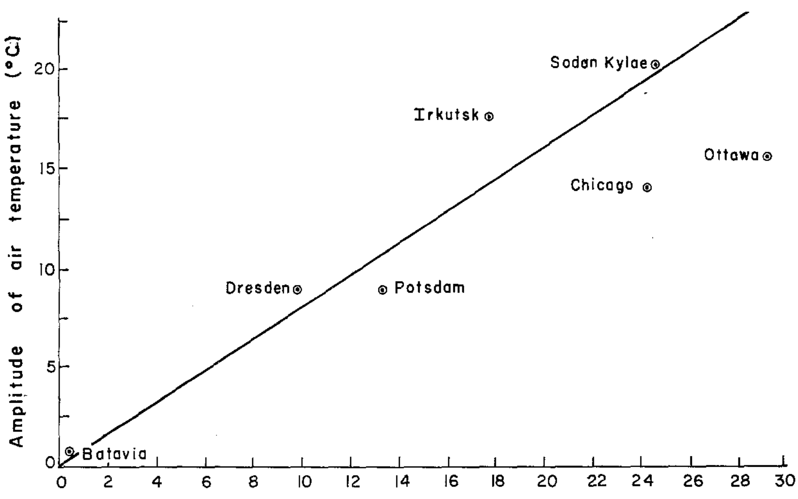

Fig. 2. Amplitude of air temperature versus daily mean ground heat storage change

depends on solar radiation, it can be assumed to have similar shape everywhere. A conversion factor *C* was determined, which gives the individual month's average daily value by multiplying the annual average daily value of *DSTG*. Three years of data from Chicago (JEN-HU-CHANG [16]) were used to determine the factor *C*:

	J	F	M	A	M	J	J	A	S	O	N	D
C	−0.92	−0.46	−0.01	+1.25	+1.42	+1.56	+1.18	+0.51	−0.35	−1.13	−1.53	−1.48

The monthly average daily value is then obtained from:

$$\overline{DSTG} = (AT.\,b) \cdot 0.584 \cdot C,$$

where b is a vegetation factor (omitted in the present method).

This average change must next be corrected for the particular synoptic situation. It is necessary to determine the depth to which short-term changes in heat storage penetrate. The Chicago data for snow free conditions were used. The interdiurnal change in temperature at 1 cm depth was calculated, and the changes at depths of 10, 20, 50 and 100 cm were expressed as a percentage of the 1 cm value. The average values obtained were:

Depth	10	20	50	100 cm
ΔT in per cent of ΔT at 1 cm	79.2	37.5	3.6	0.5 %

It is apparent that the synoptic fluctuations do not penetrate much below the 1 m level. The average temperature change of the affected layer amounted to 0.2 of the temperature change at 1 cm depth. The depth of penetration will vary with the soil properties, which should be considered. Regional information on soil parameters is generally unvailable, however, and the values given above are used for all land surfaces.

The soil heat storage changes induced by synoptic events must be treated as a correction to the climatological $DSTG$. It is obtained as follows: knowing the air temperature 24 hours previously, the change of temperature in the top 1 m of the ground can be approximated. From this ground temperature change, and the heat capacity, the synoptically induced change of soil heat storage can be evaluated. The surface temperature change must first be related to the air temperature change. This relationship was evaluated, using the Chicago data. The results are shown in Fig. 3. The following relationships are adopted between DTA (24-hour change in air temperature) and DTG (24-hour change in temperature at the surface):

if DTA positive: $DTG = DTA \cdot 0.34$
if DTA negative ($\leqq 6.1$ deg C): $DTG = DTA \cdot 0.35$
if DTA negative (> 6.1 deg C): $DTG = 2.07 + [(DTA - 6.1) \cdot 0.74]$

The Chicago observations refer to a grass covered experimental plot. It is known that DTG varies with the density of the vegetation cover. According to GEIGER [11], the temperature amplitude in the ground is reduced from 4.2 to 0.6 deg from outside to inside a forest. Using these values, and assuming that the temperature values from Chicago are representative for grass vegetation outside a forest and further assuming that, for bare ground: $DTG = DTA$,

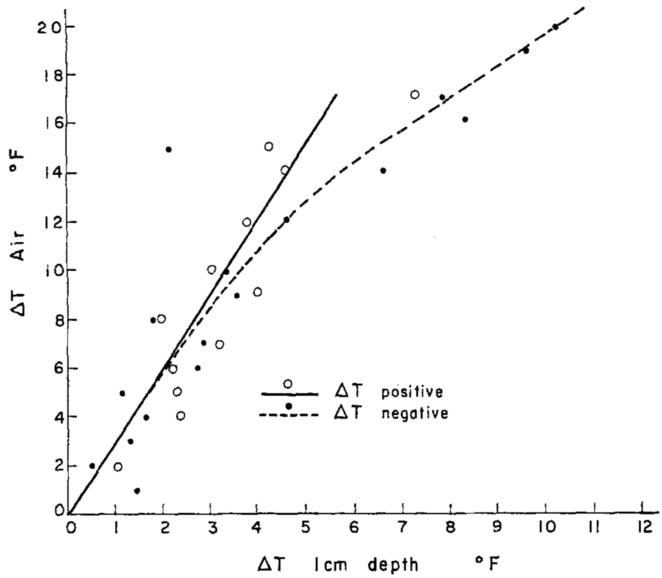

Fig. 3. Air and surface temperature change

the following relationship can be used to determine the mean value of DTG for an area:

$$\overline{DTG} = DTG \cdot d + DTA \cdot e,$$

where d = weighting factor for vegetation covered area in relation to a grass covered (experimental) site and e = weighting factor for bare ground.

General lack of sufficient observations make it necessary to assume the following weighting factors for the various vegetation types:

	d	e		d	e
mixed forest	0.25	0	grass	1.0	0
mixed shrub-grass	0.8	0.2	grass and forest	0.56	0
arctic tundra	1.0	0	shrub	0.5	0.5
broadleaf forest	0.36	0	alpine tundra	1.0	0
desert	0	1.0	taiga	0.14	0

For the further calculations it was assumed that the mean ground temperature change decreases linearly from its surface value to zero at 1 m depth.

In order to assess the heat capacity, the amount of water in the ground must be known. This is obtained from evaporation and precipitation budget calculations, as proposed by Thornthwaite and Mather [31]. A uniform water holding capacity is assumed. For day number one, full soil saturation is ascribed, and the budget is subsequently calculated for each day, using evaporation as expenditure and precipitation as income. The soil heat capacity is obtained as the sum of that of dry soil and of the water held in the soil.

With the known temperature change, that change in ground heat storage can now be calculated which is caused by synoptic events. The total change in ground storage is then obtained by adding the climatological change. Similar considerations are used if the ground is snow covered. The snow surface is then assumed to have the same temperature change as the air. The specific heat for snow is used, and the assumption that the synoptic temperature changes do not penetrate deeper than 20 cm (8″) into the snow, i. e., each inch of snow reduces the temperature amplitude at the bottom of the snow by a factor of 0.125. A similar calculation is done for ice surfaces, using the specific heat of ice. For the climatological storage change over the polar ice, those values of heat release through the ice are used, given by Vowinckel [35].

3.5. Turbulent Fluxes

Over snow free land surfaces, the amount of heat involved in evaporation (QE) is, apart from other requirements, determined by the availability of water. If sufficient water is assumed, the evaporation may be obtained after Hofmann [14]:

$$QE = K\,(R + DSTG) + V\mu.\,(qTSAT - qT)\cdot 43.2,$$

where R = surface radiation balance, $qTSAT$ and qT = specific humidity at temperature T, for saturated and actual conditions.
K and V depend on the air temperature:

°C	−40	−30	−20	−10	0	+5	+10	+15	+20	+25	+30
K	0.10	0.17	0.24	0.31	0.49	0.51	0.58	0.65	0.70	0.75	0.80
V	3.45	3.00	2.55	2.10	1.55	1.38	1.17	0.97	0.80	0.65	0.50

μ, the heat transfer coefficient, depends on the wind speed:

Knots	0	1	2	3	4	5	6	7	8	9	10	11	12
μ	0	8.0	10.6	12.7	14.3	15.5	16.9	18.1	19.2	20.3	21.3	22.5	23.5

Knots	13	14	15	16	17	18	19	20	21	22	23	24	25
μ	24.6	25.5	26.4	27.4	28.1	29.0	29.7	30.5	31.3	32.2	33.0	33.8	34.6

μ depends also on vegetation type. The above values refer to a meadow. When saturated conditions prevail, the differences are rather small for various vegetation covers. Experiments have been performed in a moist climate in Germany to find the water loss in a forested area and later in the same area after being deforested. The water loss difference was reported to be 5 percent. It has been accepted in the present method that the evaporation from a natural surface, under saturated conditions, is given by the formula and values above.

The next step requires the determination of available soil moisture. Plant evaporation continues at a fairly steady rate between saturation and the permanent wilting point. From that point to air dry soil, the rate of water loss is retarded significantly. The amount of water available during this period depends on soil characteristics but is, under all circumstances, at the most 10 percent of the total soil moisture content. It has been assumed in the present that there is a uniform availability of water as long as the soil is not completely dry.

For bare soil the availability of capillary water is reduced sharply. When a thin top layer of dry soil is formed, the water transport must be via vapour. This is a much slower process. Results reported by Bartels (Geiger [11] p. 278), indicate that evaporation during a dry spell over bare ground was 23 percent of that from short grass and 7 percent of that from a water surface. For bright days the figures were 22 and 13 percent, respectively. The values for a water surface were obtained by pan evaporation and are probably not very representative. The grass value, 22 percent, was accepted in the present. Evaporation from bare ground is thus obtained as 0.22 of the calculated value, using the formula above, if no rain fell on the particular day or the day previous.

For the individual vegetation types, the following weights are used, from the first day after the last precipitation:

Weight for bare ground (vegetation cover is then 1 minus this value): Tundra 0.7, mixed shrub-grass 0.2, grass 0, Taiga 0, broadleaf forest 0, grass and forest 0, Alpine tundra 0, mixed forest 0, desert 1.0, shrub 0.5.

The calculations of QE are valid only when sufficient water is available. The results are therefore always tested against water in the ground. If the available water is not sufficient to cover QE, the latter is set equal to the residual ground water. The value for the turbulent flux of sensible heat, QS, over vegetated surfaces, is obtained as a residual in the surface energy budget.

Over snow or ice covered land surfaces, the turbulent transport will be similar to that over ocean areas. The formulae given by MALKUS [23] are used:

$$QE = 8.7018 \cdot (qTS - qTD) \cdot F,$$
$$QS = 3.594 \cdot (TS - TA) \cdot F,$$

where qTS and qTD are specific humidity at ocean or snow surface temperature, and at the dew point temperature, F = wind speed, TA = air temperature.

These formulae are valid only for near neutral stability. Since inversions are very frequent over snow and ice surfaces, the constants in the formulae are modified for stable conditions, using results of SHULEIKIN [30]:

$$QE = 1.038 \cdot (qTS - qTD) \cdot F,$$
$$QS = 0.432 \cdot (TS - TA) \cdot F.$$

The computations are first done for $TS = TA$, and then TS is modified to fit the surface energy budget requirements.

In North America and Eurasia, large areas in the north consist of swamp and open water. For an investigation of winter-time conditions this is irrelevant, as these areas are then frozen and snow covered. In other seasons, however, it must be considered. A special grid has been prepared, giving the percentage of swamp and water for each grid area. Little is known about evaporation rates from swamps. MEINZER [25] concluded from hydrological considerations that swamp evaporation is higher than that of open water. This is quite conceivable as the relatively shallow water may be heated considerably. Also, over fresh water the evaporation will be retarded in spring, until the ice cover disappears. In the autumn the evaporation will be enhanced, before the ice forms. A factor must therefore be determined which will permit the defining of the time lag of water temperature versus air temperature. These problems were not attacked in the present method.

To calculate the turbulent fluxes over the oceans, the formulae of MALKUS are used. The change in heat storage plus ocean advection is obtained as a residual in the surface energy budget.

4. Method Finally Adopted

Calculations of snow and ice surface budgets have shown that a balance is not generally obtained, mainly due to the rather uncertain values used for ground temperature, and probably also because the turbulent transfer coefficients are inaccurate. There are also other assumptions which will add to the imbalance. It was therefore decided in this case to use the described method to calculate $L\uparrow$, QE, QS and $DSTG$ as first approximations only. The final correction necessary is then found as the result of the adjustment needed to balance the surface budget:

$$SGA + L\downarrow + L\uparrow + QE + QS + DSTG + \text{Residual} = 0.$$

The value of $L\uparrow$, QE, QS and $DSTG$ all depend on the magnitude of the surface temperature. It is thus necessary to determine the proportion of the budget imbalance which must be allocated to each of the four terms. Approximately, the change in $L\uparrow$ for a change in surface temperature of 1 degree C is 10 ly day^{-1}, within the temperature range experienced. This change is considered to be constant (a). The other three terms are allocated their shares (per degree) — (b, c, d) — of the residual, according to their relative effectiveness in heat transport at the actual time of consideration. With these values the approximate required change in surface temperature can be obtained by:

$$DTG = \text{Residual} / (a + b + c + d).$$

Using the corrected surface temperature, the terms are recalculated and produce another, smaller, imbalance. Thus, after successive calculations, the original residual in the surface balance equation will be allocated to those individual terms which depend on surface temperature.

The energy budget changes markedly from day to night, and the calculations with each set of data are performed separately as for day and night, using as radiational income at the surface ($SGA + L\downarrow$) in the first case and ($L\downarrow$) alone for night. The 24-hour values, derived from one set of observations of the various parameters are obtained by weighting the "daytime" values by the period of sun above the horizon for various latitudes, and the "nighttime" values by the remainder of the 24 hours.

The method described permits the calculation of all energy budget terms on a synoptic basis, essentially with the observations generally available in ordinary synoptic reports. The calculations are done each 12 hours, and the average 24-hour values are thus obtained by using three sets of observations. A calculation of the energy budget for different synoptic hours would be possible, taking into account the changing solar elevation with longitude, but it would contribute little as it would mainly reveal the diurnal variations, unless 3- or 6-hourly observations were used. These results would then have to be added over a 24-hour period to have meaning.

There are a number of other parameters which might be considered. The main improvement, however, would be the possibility of having available actual temperature observations from the surface and near-surface active layer of the ground. The programme has been made independent of grid points and can be used to calculate the budget for a single station or a number of chosen locations. The use of synoptic station observations rather than grid point values of the various fields is preferable over land, with real cloud observations rather than the described cloud assumptions. Over the ocean, grid point values are preferable, due to the small number of ship observations.

Adjustments and improvements have, naturally, gradually changed the method in some details as it has been used. Short wave radiation, for example, depends on a number of variables, one of which, "dust" depletion, must also be included. Further, the ozone absorption will have an upper limit and not increase indefinitely with air mass. The calculation of ground heat storage change under land surfaces will be improved by the use of results recently published by Lettau [22]. Even with the present limitations in the method it is possible to obtain synoptic energy budgets and, implicitly, water budgets, which should prove useful for the understanding of the behaviour of the earth-atmosphere system.

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