Heat flow and ice segregation at an Ottawa site
Gold, L. W.

Publisher's version / Version de l'éditeur:
https://doi.org/10.4224/20338092
Internal Report (National Research Council of Canada. Division of Building Research), 1959-06-01

NATIONAL RESEARCH COUNCIL OF CANADA

DIVISION OF BUILDING RESEARCH

HEAT FLOW AND ICE SEGREGATION AT AN OTTAWA SITE

by

L.W. Gold

Report No. 171

of the

Division of Building Research

Ottawa

June 1959

PREFACE

Measurements of ground thermal conditions have been made by the Snow and Ice Section in studies of the conditions and properties of the snow cover. Other Sections of the Division have interest in the temperatures, heat flow, moisture migration and frost effects in the ground itself. Recently the Snow and Ice Section has been asked to broaden its interests and to assume responsibility for some of the Divisional studies of the ground thermal condition as well.

It has been possible from results of several sets of measurements made at the outdoor experimental area adjacent to the Building Research Centre to come to certain tentative conclusions regarding the changing thermal conductivities and the extent of ice segregation in the ground during freezing. These are now reported. Further confirmation is desirable and may be obtained from subsequent measurements.

The author is a research officer with the Division and as head of the Snow and Ice Section has special interests in the properties of snow and ice and the environmental factors which affect them.

Ottawa
June 1959

N.B. Hutcheson,
Assistant Director.

HEAT FLOW AND ICE SEGREGATION AT AN OTTAWA SITE

by

L. W. Gold

During the winters 1955-56, 1956-57, 1957-58, observations were made on the heat flow from the ground in the Ottawa area (1). The observation site was a grassed field with the snow cover undisturbed. Temperatures in the ground down to the 8-foot level were measured at a site nearby during 1956-57 (2) and in the immediate vicinity down to the 25-foot level during 1957-58. It was desired to correlate the heat-flow measurements with the ground-temperature observations. A series of observations was therefore made on the water content of the frozen soil so that an estimate of the heat associated with the freezing of the soil water could be made if necessary.

The analysis of the heat flow and the ground temperature observations was made (3). For this study, the observations related to the freezing of the soil water were not described. Since these observations are relevant to the frost action problem, and since field observations and their interpretation are always of value in understanding what is actually occurring under natural conditions, they are presented and discussed in this paper.

WATER CONTENT OF THE FROZEN SOIL

The soil in which the observations were made is a highly sensitive plastic clay with 65 per cent by weight having a grain size in the clay range (4). Figure 1 shows that the average water content in per cent of dry weight tends to increase with depth down to about the 20-foot level but that below this it is reasonably constant at 80 per cent. The values for the water content for depths below 15 feet were taken from another paper (4).

In 1958, the moisture contents in the frozen layer were measured at the end of February and beginning of March when the depth of the frozen ground was a maximum, about 30 cm. (12 in.). Two methods were used. Initially, soil chips as obtained with an auger were used to determine the moisture content in per cent of the dry weight of the soil. Unfortunately, this method does not give the water content per unit volume which is the quantity required for the heat-flow determinations. Continuous samples were therefore obtained by driving cylindrical steel thin-walled tubes vertically into the frozen layer and extracting a sample of the frozen soil. These samples were immediately removed to a cold room and cut into sections of known volume. The water content and dry density of the soil in these volumes were then determined. Figure 2 shows the observed change with depth of the water content expressed as per cent of dry weight. In Fig. 3 is shown the observed change with depth of the water content per unit volume and the dry density.

It is well known that all the water in a fine-grain porous material does not freeze when the temperature is lowered below 0°C (5). The ground temperature at 10 cm. (4 in.) below the ground surface at the observation site was never lower than -5°C during February 1958. When the frozen sample was taken it had risen to -3°C. From the results of the work of Lovell, it was estimated that the weight of unfrozen water per unit volume was about 25 per cent of the dry density of the soil for the temperatures encountered. This estimate is probably a little high.

Calorimetric observations by P.J. Williams indicated that the estimate was of the right order. Unfortunately, the potential errors when using crude equipment in the field made the calorimetric observations unreliable. In Table I are shown the calculated amounts of unfrozen and frozen water based on the average of the water content observations shown in Fig. 3. In Fig. 3, the change with depth of the calculated frozen water content per unit volume is given. To convert to heat (cal/cm^3) the frozen water content per unit volume was multiplied by 80. The dashed line in Fig. 3 gives the values for the ice content per unit volume used in the heat calculations.

From the ground temperature observations, it is possible to obtain the time dependence of the depth of the 0°C isotherm. This is plotted in Fig. 4 for 1956-57 and in Fig. 5 for 1957-58. From the observations made by the Soil Mechanics Section of the Division of Building Research on the depth to the bottom of the frozen layer during 1956-57, it appears that this depth need not coincide with the 0°C isotherm. This was borne out by the more careful observations during February 1958. Unfortunately, the depth of the frozen layer was not observed continuously during the 1957-58 winter.

In order to estimate the rate at which the water was frozen in the soil during 1957-58, it was assumed that the depth to the bottom of the frozen layer coincided with the 0°C isotherm until January 15 and that it decreased linearly thereafter to the maximum depth on February 26.

TABLE I

ESTIMATED ICE CONTENT FOR EACH DEPTH FOR THE WINTER OF 1957-58

Depth (cm.)	(in.)	Average measured dry density (gm/cc)	Calculated unfrozen water (gm/cc)	Average measured water content (gm/cc)	Estimated ice content (gm/cc)
2.5	1	0.81	0.20	0.55	0.35
5	2	0.68	0.17	0.57	0.40
7.5	3	0.82	0.20	0.59	0.39
10	4	0.92	0.23	0.54	0.31
12.5	5	0.83	0.21	0.55	0.34
15	6	0.82	0.20	0.61	0.41
17.5	7	1.10	0.28	0.49	0.21
20	8	1.14	0.28	0.50	0.22
22.5	9	1.16	0.29	0.50	0.21
25	10	1.16	0.29	0.51	0.22
27.5	11	1.32	0.33	0.49	0.16
30	12	1.37	0.34	0.44	0.10

From the information plotted in Figs. 4 and 5 it was possible to estimate the rate of increase of the depth of the frozen layer. With this information, and that contained in Fig. 3, it was possible to calculate the time dependence of the heat associated with the freezing of the water. In Figs. 4 and 5, the heat flow measured at the 5-cm level for 1956-57 and the 10-cm level for 1957-58 and the calculated heat associated with the freezing of the water for these winters are given. The calculations for the winter of 1956-57 were made possible by assuming that the water content per unit volume in the frozen layer was the same as for 1957-58. This assumption was considered reasonable as the moisture contents as a percentage of dry weight determined during the 1956-57 winter by the Soil Mechanics Section agreed with those determined during 1957-58 as shown in Fig. 2.

One of the interesting facts shown in Fig. 3 is the decrease in ice content with depth. In the following discussion, various factors are considered which might be responsible for this observed decrease.

FACTORS RELATED TO ICE SEGREGATION IN THE SOIL

Penner (6), Power (7), and Gold (8) have shown that during ice segregation the temperature of the freezing plane should be depressed below the normal freezing temperature of water (0°C). The observations on the depth to the freezing plane and the depth to the 0°C isotherm support this conclusion. The estimated temperature at the freezing plane at maximum penetration was about -1°C for the 1956-57 winter and about -0.5°C for the 1957-58 winter.

Overburden Pressure

Penner (6) has shown that a clay soil such as that at the observation site, can develop negative pressures in excess of 1000 gm/cm^2 in the pore water, before ice segregation in the soil is halted. He has also shown that if the pore water pressure is zero, an overburden pressure greater than the maximum negative pore water pressure is required to stop the ice segregation (9). For the depths of freezing and soil densities encountered in the field observations, the overburden pressure would hardly exceed 100 gm/cm^2. It is therefore concluded that overburden pressure would have little influence on the observations.

Rate of Flow of Water to Freezing Plane

The water content in the upper 15 cm (6 in.) of the soil in late October and early November before freezing commenced was between 25 to 30 percent of the dry weight. From Fig. 1, which gives also the standard deviation in the water content observations of each level, it is seen that the water content is not much above 30 per cent down to the 45 cm (18 in.) level. Since about 25 per cent by dry weight of the water remains unfrozen the amount of water drawn into any level during freezing would be about equal to the amount of water frozen. In Table II are given the calculated flow rates corresponding to the estimated amount of frozen water for various levels for the two years.

TABLE II

ESTIMATED UPPER LIMIT TO RATE OF FLOW OF WATER TO FREEZING PLANE

Depth		1956-57	1957-58
cm	in.	$gm/cm^2/day$	$gm/cm^2/day$
0-15	0-6	0.21	0.14
15-25	6-10	0.14	0.09
below 25	below 10	0.05	

From Table II it is seen that the maximum rate observed was of the order of 0.2 $gm/cm^2/day$ or 8.3×10^{-3} $gm/cm^2/hr$. The soil has a permeability coefficient of the order of 0.8×10^{-3} (cc/cm^2) / $(cm\ H_2O/cm)$. Penner has shown that this permeability coefficient is almost independent of water content over the soil water pressure range of 0 to -500 gm/cm^2 (10). For pressures less than -500 gm/cm^2, the permeability coefficient decreases slowly with decreasing pressure. To supply 8.3×10^{-3} $gm/cm^2/hr$ would require a pressure gradient in the soil moisture of the order of 10 $(gm/cm^2)/(cm)$. The freezing plane could therefore draw water from a depth below it of about 60 cm (2 ft) before the permeability coefficient would begin to decrease and thus become a limiting factor in the flow rate. As the freezing plane penetrates, the flow rate decreases and water can be drawn from greater depths still without seriously affecting the permeability coefficient of the soil. Since the water table at the observation site was within 3 ft of the surface when the winter began, the observations indicate that availability of water and permeability of the soil should not have interfered seriously with the ice segregation.

Negative Pore Water Pressures

Penner (6) has shown that the rate of heaving of a frost-susceptible soil, and therefore the amount of ice formed over a given period of time, depends partly on the magnitude of the pore water pressure. No measurement of the pore water pressure was made in the field but the existence of the temperature depression at the freezing plane indicates that a negative pore water pressure did exist. An increasing negative pressure as the frost line penetrates would contribute to the decrease in frozen water content with depth.

Pearce has carried out a detailed analysis of ground temperatures observations made in the Ottawa area (11). He found that below 30 cm (1 ft) the observations satisfied certain requirements for the application of standard heat conduction theory and could be fitted closely by the first harmonic approximation to the annual temperature cycle at any depth. Above 30 cm, deviations occurred which could be attributed to non-homogeneity of the soil, the influence of soil water and to the higher frequency components in the temperature variations. He concluded that the simple thermal model, which was satisfactory for depths below 30 cm, did not accurately describe the thermal behaviour of the upper soil layers. It is necessary to go directly to the observations to obtain an appreciation of the thermal behaviour of these layers.

In Fig. 6 is plotted a series of temperature - depth curves. A sudden change in the temperature gradient at about the 15 cm (6 in.) level is characteristic of these profiles. Above 15 cm, the soil has a significant organic content and therefore a lower thermal conductivity is to be expected.

The thermal properties of this soil are such that daily temperature fluctuations have virtually disappeared at the 15-cm level. Lower frequency temperature cycles due to variations in the weather have virtually disappeared at the 60 cm (2 ft) level. When there is more than 15 cm (6 in.) of snow cover, the daily fluctuations in temperature virtually disappear at the 5-cm level. Generally speaking, the higher the frequency of the temperature disturbance, the larger can be the possible temperature gradients associated with it.

Since the ground temperatures do not change very rapidly when there is a snow cover, it is possible to obtain from the ground temperature and heat flow observations, values for the apparent thermal conductivity of the soil for the upper 15 cm and for the layer below the freezing plane. Average values are given in Table III for the period January 14 to 21 and February 14 to 21, 1958. The heat flow in the unfrozen layer was assumed equal to the heat flow in the frozen layer minus the amount calculated for the freezing of the water.

Over the period used in the calculations, the thermal conductivity is fairly constant, but it changes markedly from January to February. Observations on the thermal characteristics of the snow cover tend to confirm that this change is not due to errors in the measurement of the heat flow and so must be related to changes in the thermal properties of the soil. It is possible that this change with time is related to the movement of water vapour under the combined temperature and vapour pressure gradient which need not be simply related near the ground surface. The figures in Table III indicate the difficulty in calculating the heat flow through soil near the surface from assumed thermal constants even over short periods, the density and the water content of the soil being known.

TABLE III

AVERAGE THERMAL CONDUCTIVITY VALUES CALCULATED FROM THE TEMPERATURE GRADIENT AND HEAT FLOW OBSERVATIONS

Period 1958	Calculated Thermal Conductivity			
	* 0-15 cm (0-6 in.)		40-80 cm (16-32 in.)	
	$cal/cm^2/hr$ / °C/cm	$Btu/ft^2/hr$ / °F/ft	$cal/cm^2/hr$ / °C/cm	$Btu/ft^2/hr$ / °F/ft
Jan 14-21	5.0	0.34	13.9	0.93
Feb 14-21	2.9	0.19	9.1	0.61

* Values calculated for the frozen layer not corrected for heaving of the soil. These values are probably about 30 per cent too low.

Heat Balance at the Freezing Plane

Before discussing the possible relation between the temperature gradient and ice segregation, it will be of value to introduce some simple heat flow conditions and equations which apply. Considering the heat balance at the freezing plane, the heat flowing from the freezing plane to the surface must equal the heat flowing from the unfrozen soil to the freezing plane plus the heat released when the water freezes. If q_f is the heat flow in the frozen layer, q_u is the conducted heat in the unfrozen layer and q_ℓ is the heat released when the water freezes. Then

$$q_f = q_u + q_\ell .$$

If it is assumed that the thermal conductivity of the frozen soil remains constant and that the ratio of the heat flow to the temperature gradient in the unfrozen soil is also constant over a reasonable length of observation time, then

$$q_f = k_f \, (\partial T_f) \, / \, (\partial x)$$

where k_f is the thermal conductivity of the frozen soil and

$(\partial T_f) / (\partial x)$ is the temperature gradient in the frozen soil;

$$q_u = k_u \, (\partial T_u) \, / \, (\partial x)$$

where k_u is the apparent thermal conductivity of the unfrozen soil and $(\partial T_u) / (\partial x)$ is the temperature gradient in the unfrozen soil.

For ice segregation to occur, $k_f \, (\partial T_f) / (\partial x)$ must be greater than $k_u (\partial T_u) / (\partial x)$. In Table IV are given the observed values for q_f and the estimated values for q_u and q_ℓ for various periods. The depth of the freezing plane is also given.

From Table IV it is seen that from the surface down to the 25 cm (10 in.) level, q_f on the average is greater than twice q_u. The temperature profiles plotted in Fig. 6 show no marked discontinuity in the temperature gradient at the freezing plane, within the accuracy of the experiments, which did not exist while the soil was in the unfrozen state. It would therefore appear that the increase in heat flow on the frozen side of the freezing plane is associated primarily with an increase in the thermal conductivity.

Kersten (12) has shown from laboratory studies that when soil freezes there can be a marked increase in the thermal conductivity. The amount of increase depends on the moisture content; the higher the initial moisture content, the greater the increase. For a fine-grain soil such as clay, he shows an increase of about 50 per cent for an initial water content of 30 per cent of the dry weight. Kersten reported that no ice segregation occurred during freezing. If ice segregation were to occur in the soil during freezing, it is to be expected that the increase in the thermal conductivity would exceed the values reported by Kersten.

Heat Flow and Ice Segregation

Let it be assumed, as indicated by the observations, that the temperature gradient is continuous across the freezing plane. Assume also that the thermal conductivity of the soil increases by a factor α on freezing; $k_f = \alpha k_u$.

Then $q_f = (\alpha)\, k_u\, (\partial T_f) / (\partial x)$.

$$(\partial T_f) / (\partial x) = (\partial T_u) / (\partial x) \text{ at the freezing plane.}$$

$$q_\ell = k_u\, (\alpha - 1)\, (\partial T_f) / (\partial x)$$

$$(q_\ell) / (q_f) = (\alpha - 1) / (\alpha)$$

For $(q_\ell) / (q_f)$ to be of the order of 0.70 the conductivity in the frozen state must be about 3 times the conductivity in the unfrozen. Information is not available to show if this increase can occur with the soil in the upper layers at the observation site, but the temperature profiles in Fig. 6 indicate that it is possible.

TABLE IV

OBSERVED VALUE FOR q_f AND ESTIMATED VALUES FOR q_u AND q_ℓ .

Period	Depth of freezing plane cm	in.	q_f (cal/cm²/day)	q_u (cal/cm²/day)	q_ℓ (cal/cm²/day)	$(q_\ell)/(q_f)$
1956-57						
Nov 15 - Dec 5	0-15	0- 6	24	7	17.0	0.71
Dec 5 - Dec 25	15-25	6-10	21	10	11.0	0.52
Dec 25 - Feb 15	25-55	10-22	22	17.5	4.5	0.20
1957-58						
Dec 1 - Dec 10	0-15	0- 6	16	4.5	11.5	0.72
Jan 10 - Feb 25	15-25	6-10	14.5	8.0	6.5	0.45

If $(\partial T) / (\partial x)$ does not increase by very much at the freezing plane, then q_ℓ should depend primarily on k_u, α and the temperature gradient at the freezing plane. Generally speaking, when the freezing plane is advancing into the ground, the temperature gradient decreases with depth and at any depth, decreases with time. Moreover, the larger q_ℓ or rate of ice segregation, the larger should be the value for α . These factors working together tend to make q_ℓ , the rate of ice segregation, largest near the surface and to decrease as the freezing plane penetrates. The amount of water frozen at any level would depend on the rate of ice segregation and the rate of penetration of the freezing plane. Figure 4 shows that for 1956-57, when no extended warm periods occurred such as during December of 1957, the frost line on the average advanced downward at an almost constant rate. It is of interest to note that the rate of advance of the 0°C isotherm below the 30 cm (12 in.) level agreed reasonably well with that which would be predicted from the thermal model developed by Pearce taking into account the snow cover.

The one factor which has been neglected in this discussion is the heaving of the ground during ice segregation. The decrease in dry density in the upper level shown in Fig. 3 is probably due to this heaving. The estimated 0.36 gm/cm^3 of frozen water at the surface is consistent with an initial dry density of about 1.2 gm/cc assuming this amount of water caused a heave of about 0.36 cm. This value for the initial dry density is not unreasonable. If the swelling of the soil is about 30 per cent, then for the upper 15 cm (6 in.) the temperature gradients given in Fig. 6 are too steep by about 30 per cent and the thermal conductivities in Table III too low by about the same amount. These errors decrease with depth.

The original purpose of the heat flow observations made at the Montreal Road Laboratory site was to determine the effect of snow cover on heat flow from the ground. It was shown (1) that during periods of continuous below freezing temperature, the relationship between the integrated heat flow from the ground and the air temperature and snow depth is

$$\int q_f \, dt = k_s \int (T_a - T_g) \, / \, (h_s) \, dt$$

where the integration is over the time period of interest and

k_s is the thermal conductivity of the snow corresponding to the average density of the snow cover

T_a is the air temperature measured at the 120 cm (4 ft) level in a Stevenson screen

T_g is the ground temperature at the 10 cm level

h_s is the depth of the snow cover plus 5 cm

It was assumed that the insulating effect of the ground was about half that of the snow. The observations on the effect of the snow cover showed that the rate of heat loss from the ground varies directly with the difference in temperature between the air and the temperature at the ground-snow interface and inversely as the depth of the snow for snow of given thermal properties. With these facts, it is possible to account for all the significant fluctuations in the heat flow and shallow ground temperature observations during the winter period.

SUMMARY

At an observation site at Ottawa, ice segregation was observed in a grass covered clay soil with undisturbed snow cover. The amount of ice segregation decreased with depth. The observations, interpreted in the light of published laboratory investigations, indicate that the ice segregation was not seriously limited by overburden pressure or by availability of water.

Two factors were considered which would account for the decrease in frozen water content with depth. The first factor was the development of negative pore water pressures which affects the rate of ice segregation. The second was the thermal balance at the freezing plane. Since the observed temperature gradient showed no discontinuity at the freezing plane, it was concluded that the thermal conductivity of the frozen soil must be greater than the unfrozen in order for the heat flow to balance at the freezing plane during ice segregation. The rate of ice segregation would thus depend on the increase in the thermal conductivity of the soil on freezing, the magnitude of the temperature gradient and the magnitude of the pore water pressure. The thermal conditions and the hydraulic conditions must always adjust themselves so that equilibrium is maintained. The decrease in the temperature gradient and the likely increase in the negative pore water pressure with depth and with time during the freezing period combined with an almost constant rate of penetration of the freezing plane would account for the decrease in ice content with depth.

References

1. Gold, L.W. Influence of snow cover on heat flow from the ground. Some observations made in the Ottawa area. From Association Internationale d'Hydrologie Scientifique, Union Geodesique et Geophysique Internationale, Assemblée Générale de Toronto 1957, Gentbrugge 1958, Tome IV, pp 13-21. NRC 4827.

2. Legget, R.F. and C.B. Crawford. Soil temperatures in water works practice. Jour. Amer. Water Works Assoc. vol. 44, no. 10, October 1952, pp 923-939. Division of Building Research, National Research Council, Ottawa, NRC 2910.

3. Pearce, D.C. and L.W. Gold. Observations on the ground temperature and heat flow at Ottawa. Submitted for publication.

4. Eden, W.J. and C.B. Crawford. Geotechnical properties of Leda clay in the Ottawa area. Division of Building Research, National Research Council, Ottawa, 1957, NRC 4383.

5. Lovell, C.W. Temperature effects on phase composition and strength of partially-frozen soil. Highway Research Board, Bull. 168, 1957. pp 74-95.

6. Penner, E. Soil moisture tension and ice segregation. Highway Research Board Bull. 168, 1957. pp 50-64. Division of Building Research, National Research Council, Ottawa, NRC 4738.

7. Powers, T.C. Resistance of concrete to frost at early ages. RILEM Symposium on Winter Concreting, Copenhagen, 1956, Proceedings, Session C, General Report, 50 p.

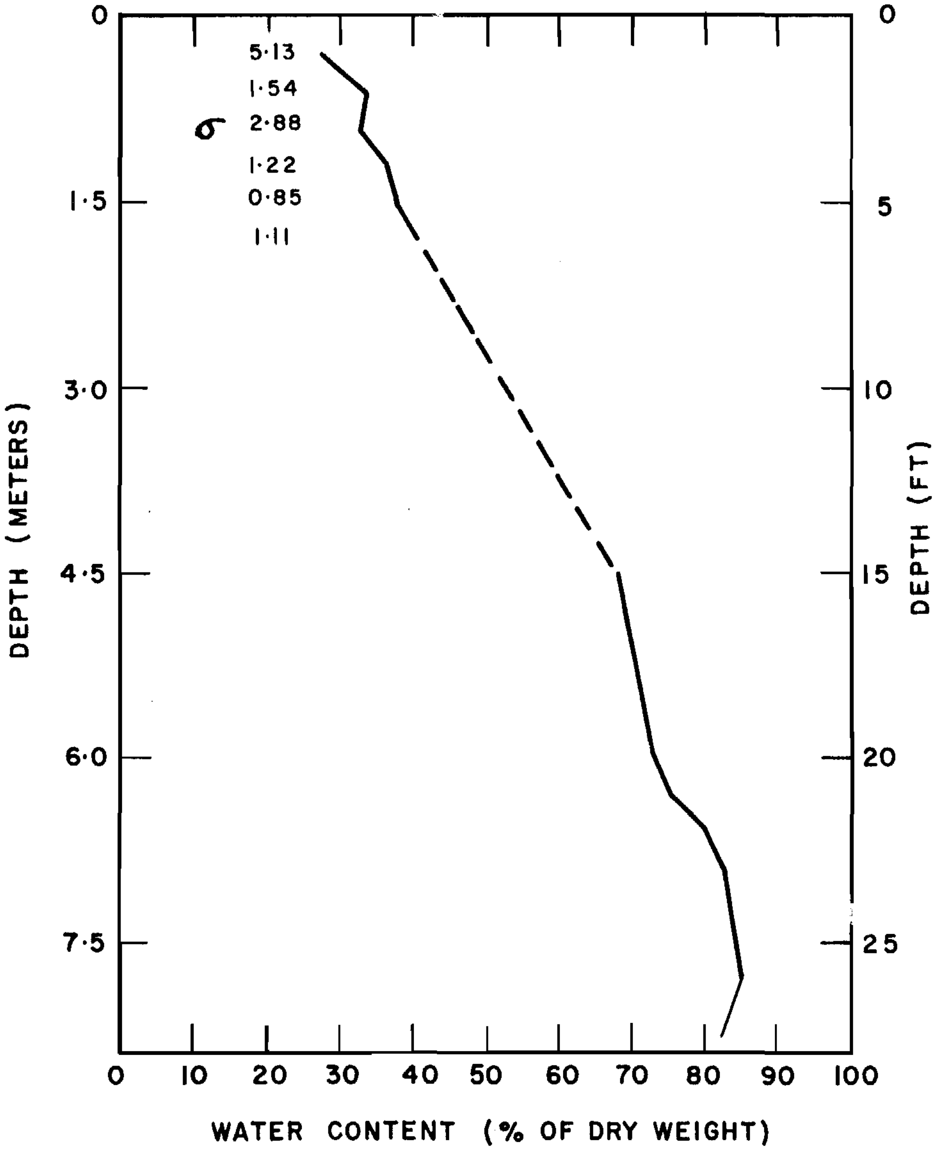
8. Gold, L.W. A possible force mechanism associated with the freezing of water in porous materials. Highway Research Board Bull. 168, 1957. pp 65-73. Division of Building Research, National Research Council, Ottawa, NRC 4737.

9. Penner, E. Pressures developed in a porous granular system as a result of ice segregation. Presented at the 37th Annual Meeting of the Highway Research Board, Washington, D.C. January 1958.

10. Penner, E. On the mechanism of frost heaving in soils. Presented at the 38th Annual Meeting of the Highway Research Board, Washington, D.C. January 1959.

11. Pearce, D.C. Ground temperature studies at Saskatoon and Ottawa. From Association Internationale d'Hydrologie Scientifique, Union Geodesique et Geophysique Internationale, Assemblée Générale de Toronto 1957, Gentbrugge 1958, Tome IV, pp 279-290. Division of Building Research, National Research Council, Ottawa, NRC 4796.

12. Kersten, M.S. Thermal properties of soils. University of Minnesota, Eng. Exp. Station, Bull. No. 28, Minneapolis, June 1949. 225 p.



VALUES FOR 1 TO 6 FT. AVERAGE OF MONTHLY READINGS TAKEN OVER PERIOD OCT./56 TO OCT./57.
σ IS STANDARD DEVIATION OF THE 13 OBSERVATIONS AT EACH DEPTH.
BELOW 15 FT, VALUES TAKEN FROM REFERENCE 4.

FIGURE 1
WATER CONTENT OF SOIL AT OBSERVATION SITE

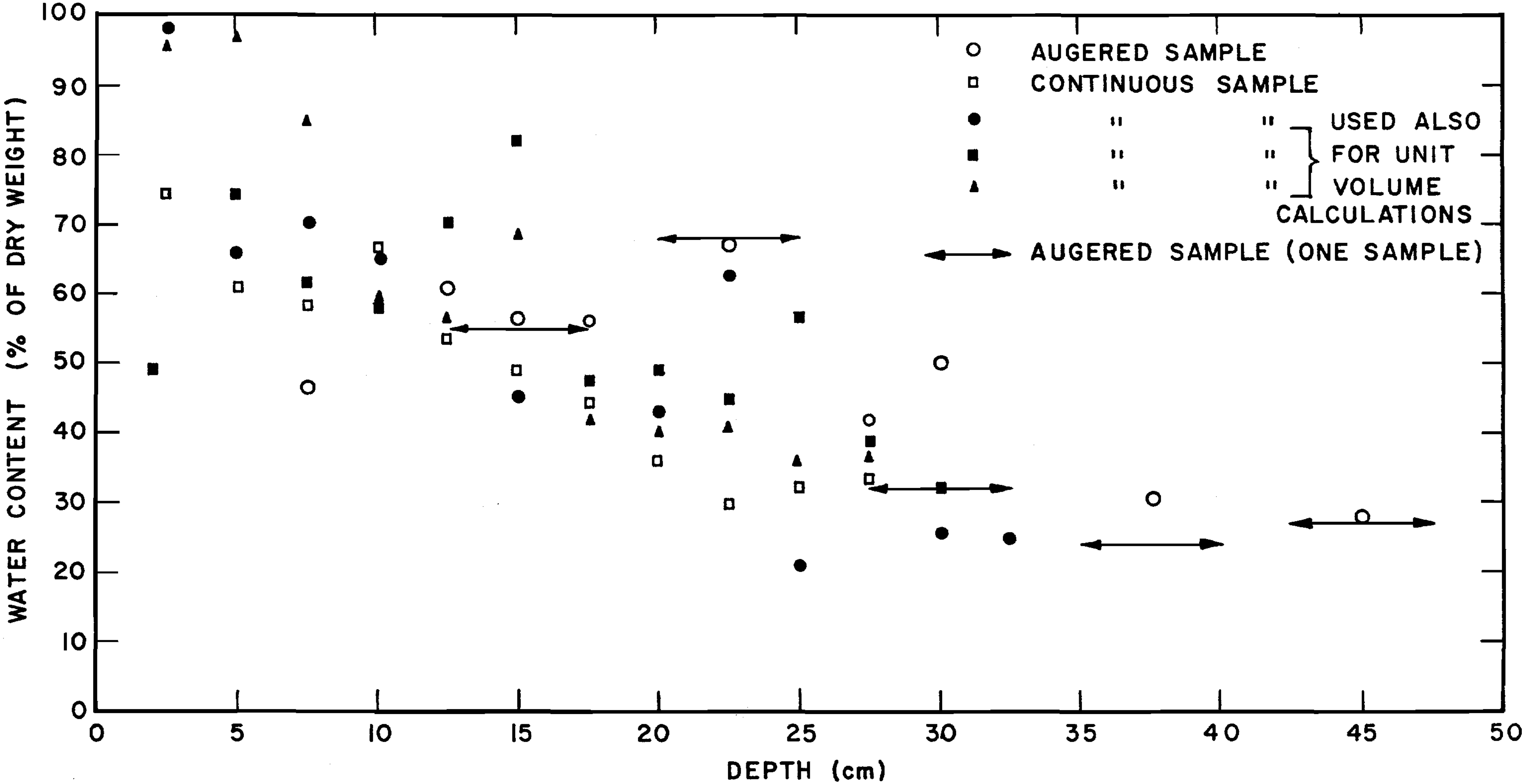


FIGURE 2

DEPENDENCE OF MOISTURE CONTENT OF FROZEN SOIL ON DEPTH

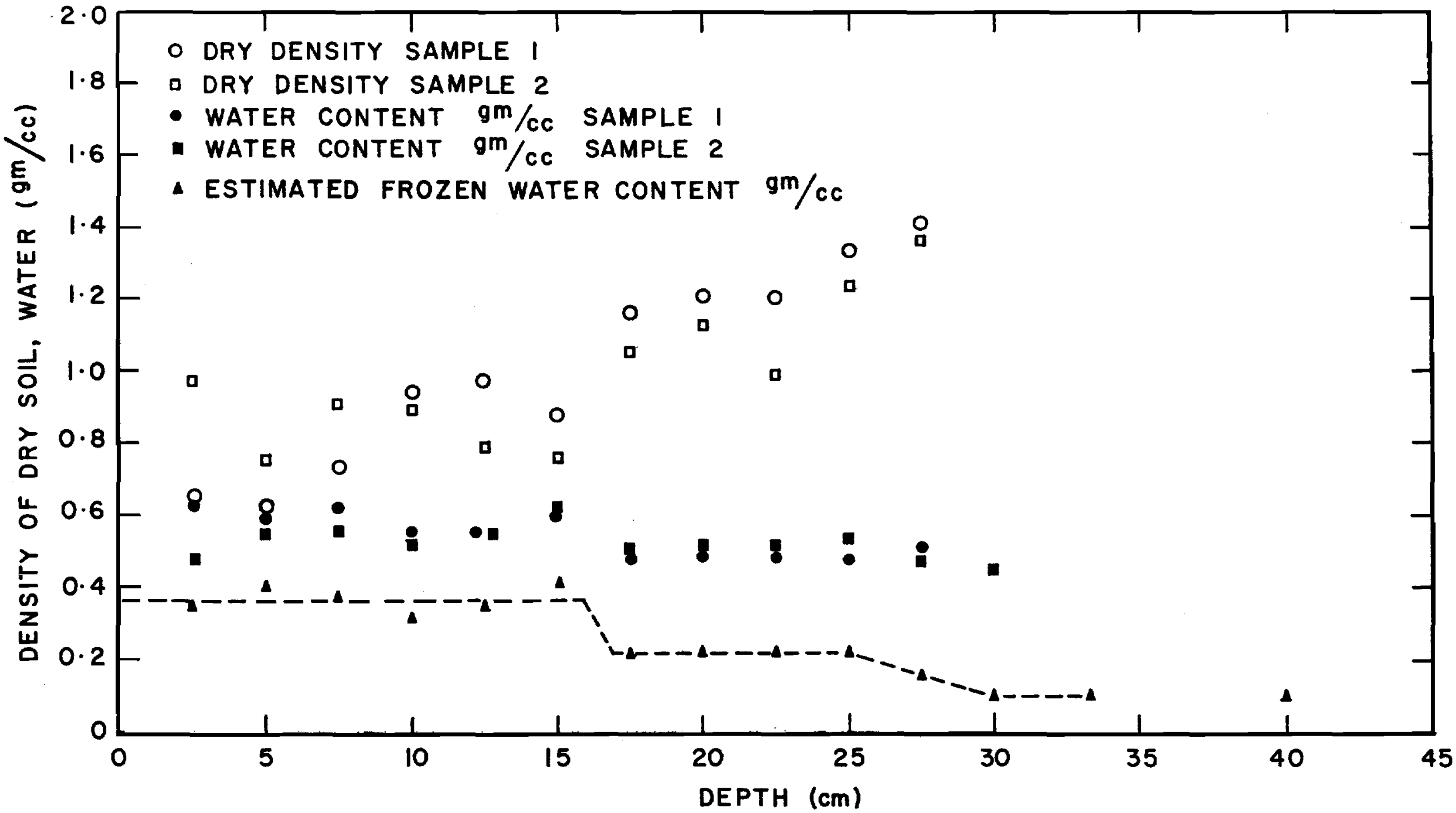


FIGURE 3

DEPENDENCE OF THE DRY DENSITY OF THE SOIL AND THE WATER CONTENT ON DEPTH

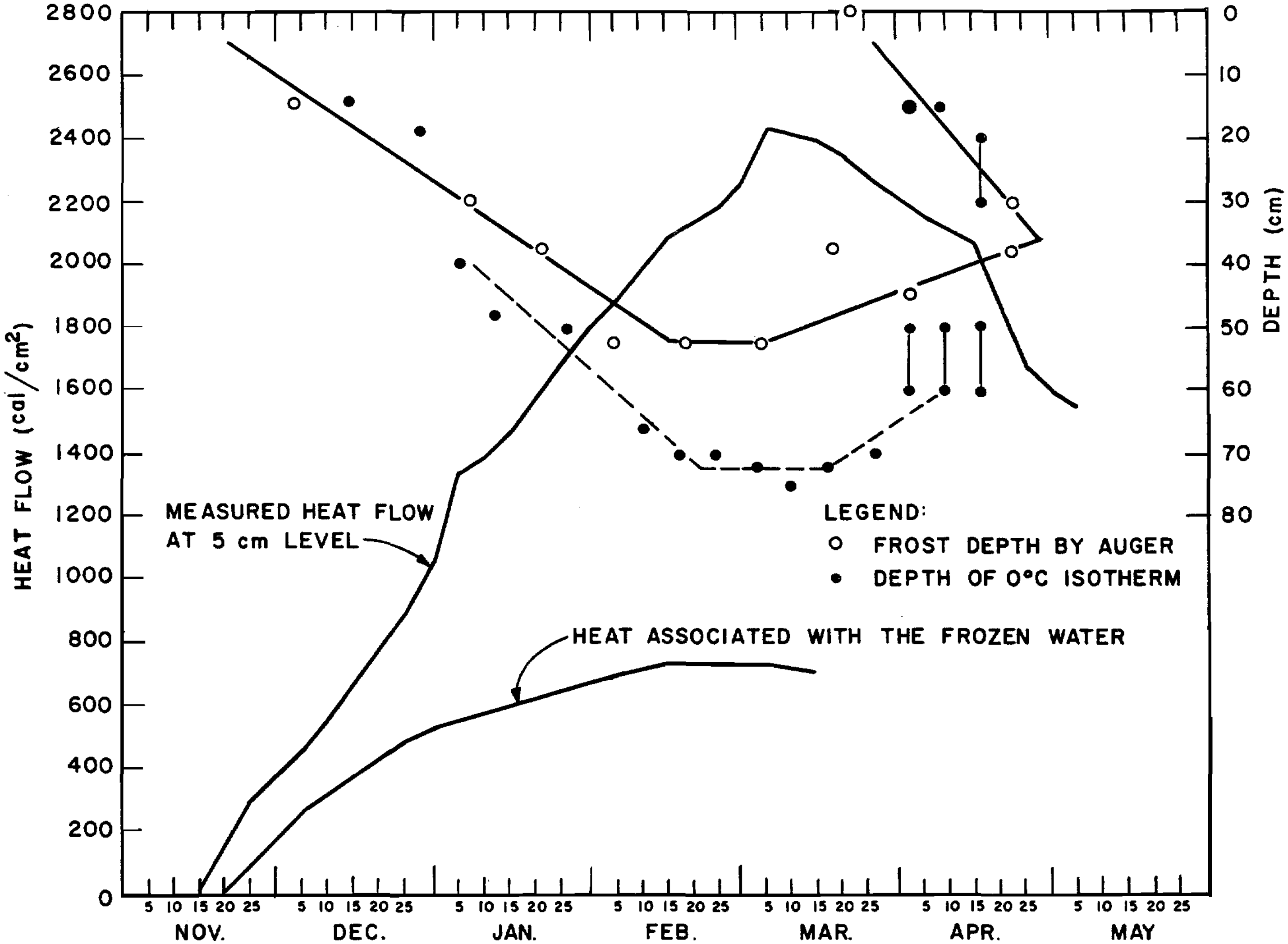


FIGURE 4

HEAT FLOW AND DEPTH OF FROST OBSERVATIONS FOR 1956-57

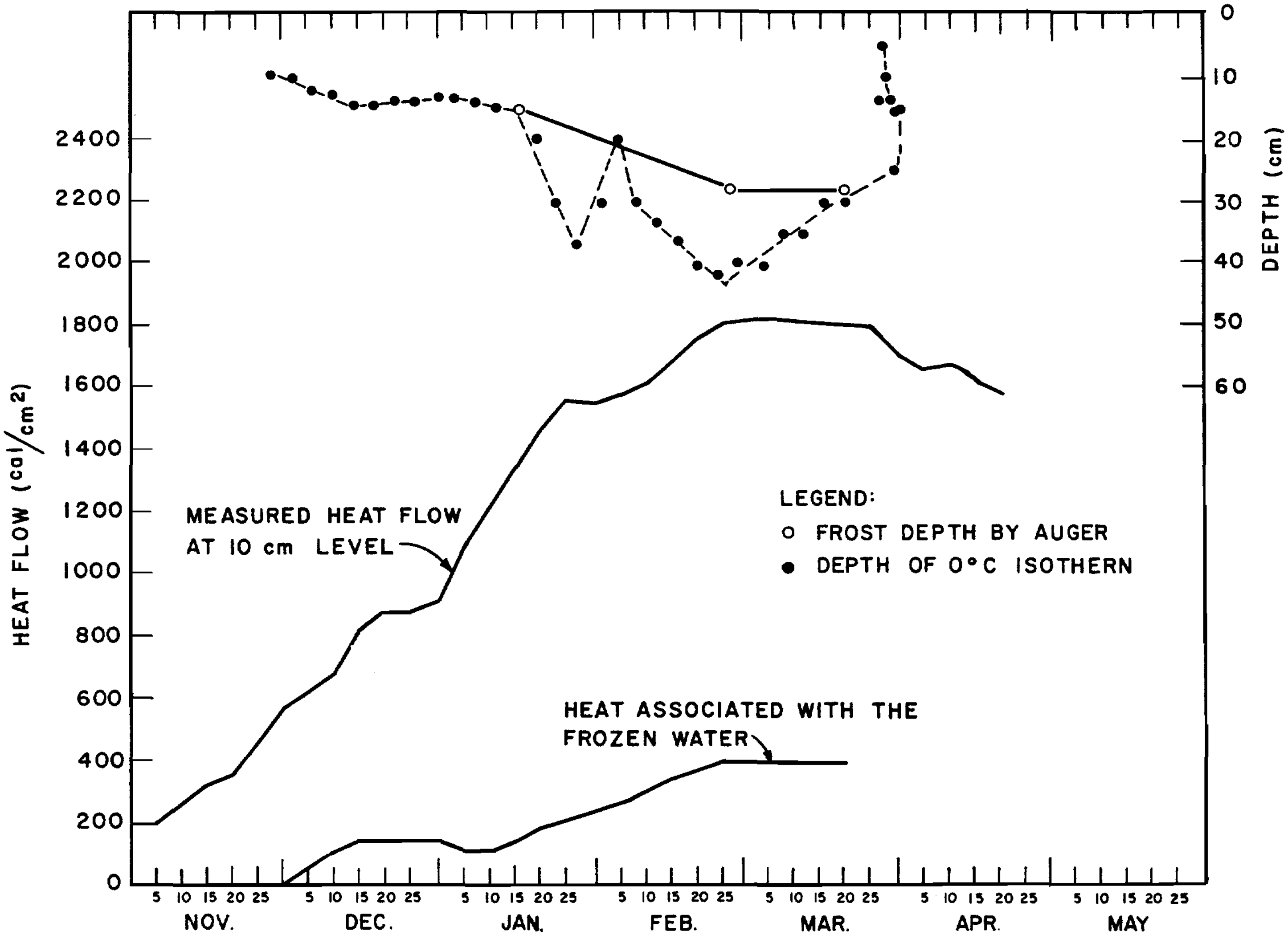


FIGURE 5
HEAT FLOW AND DEPTH OF FROST OBSERVATIONS FOR 1957-58

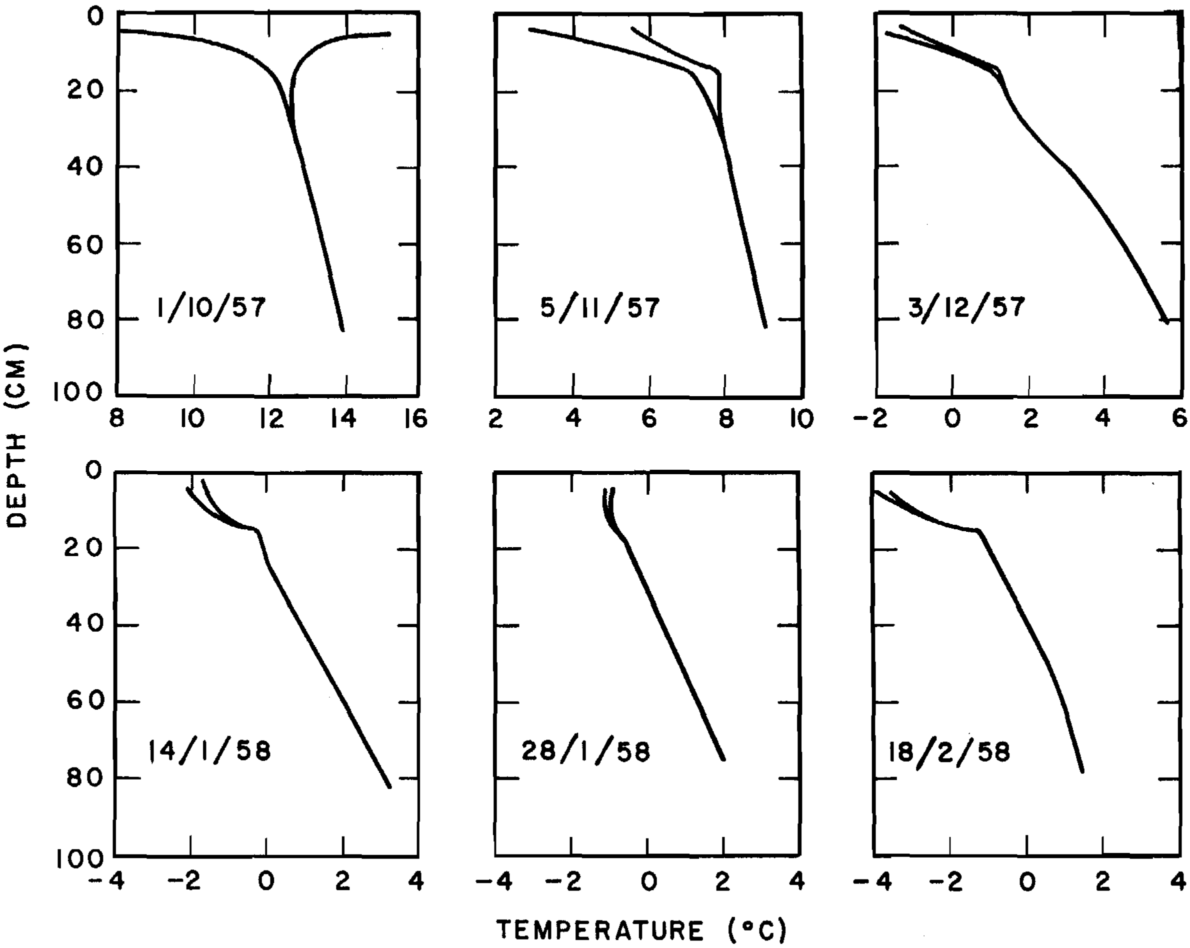


FIGURE 6

GROUND TEMPERATURE PROFILES

THE DAILY EXTREMES ARE SHOWN FOR EACH PROFILE