Biuletyn Peryglacjalny, no. 38. 1999

STUART A. HARRIS*
Calgary

HEAT TRANSFER PROCESSES AT THE LOWER LIMIT OF ALPINE PERMAFROST, MARMOT BASIN, JASPER NATIONAL PARK, CANADA

Abstract

In March 1979, a hole was drilled to a depth of 16.8 m through 1 m of till and loess into the clast-supported boulders of a fossil Pre-Late Wisconsin rock glacier just above tree line (2,195 m elevation) at the Marmot Basin Ski Area, Jasper National Park (latitude 52°47' 36.5" N, 118°06'45.9" W). Vegetation cover consists of a moist alpine meadow, on a bench which slopes gently to the south on the otherwise steep mountainside. The mean daily temperature for the site has been recorded continuously, and monthly readings have been made of ground temperatures using a thermistor string. Depth of the snow pack has been recorded on each visit, while November to April snowfall data are measured daily at 1,985 m elevation. The mean annual air temperature (–2.1°C) has decreased by about 0.7°C since 1979, while the average winter snowfall (c. 310 cm) has increased by about 70 cm.

A water table occurs at about 12 m depth with a water temperature of about 1.1°C. Permafrost up to 6.8 m thick was present for almost all the period of study, with the active layer ranging from 2.2 to 4.3 m in depth. Temperatures in the permafrost varied seasonally, ranging down to –2.5°C. The ground temperatures are constantly changing and are more closely related to thickness and duration of snow cover rather than air temperatures. The geothermal gradient averages 0.45° per metre in the upper 10 m, and there is a "thermal offset" of about 4.1°C. Heat conduction is dominant at the surface, being replaced by closed-cell convection between the blocks. In the early summer, warmer snow melt from the adjacent mountain side increases the thickness of the active layer and sometimes pierces the permafrost so that the warm water descends to the main water table below.

INTRODUCTION

In March 1979, a series of four boreholes were drilled at Marmot Basin Ski Area, southwest of Jasper, Alberta (Fig. 1). They were part of a cooperative program between Dr. ROGER BROWN of the Building Research Division of the National Research Council of Canada and the author in the Department of Geography, University of Calgary. Ground temperature cables were emplaced in the holes, measurements being made every month, and the results were used in delimiting the lower limit of permafrost in southwest Alberta (HARRIS and BROWN, 1982) and along the

*Department of Geography, University of Calgary, Calgary, Alberta T2N 1N4, Canada

Eastern Cordillera of North America (HARRIS, 1986). Three of the boreholes penetrated permafrost, but one temperature cable soon failed. The other two have worked satisfactorily to date, and this paper will describe the results from MB #2, which lies on a bench below the upper slope of Marmot Mountain. It is of particular interest because of the materials it penetrates, the topographic situation, and its proximity to the lower limit of continuous permafrost (Pl. 1). The 20-year record permits the identification of two new processes in heat transfer in the ground which have not previously been demonstrated in the natural environment.

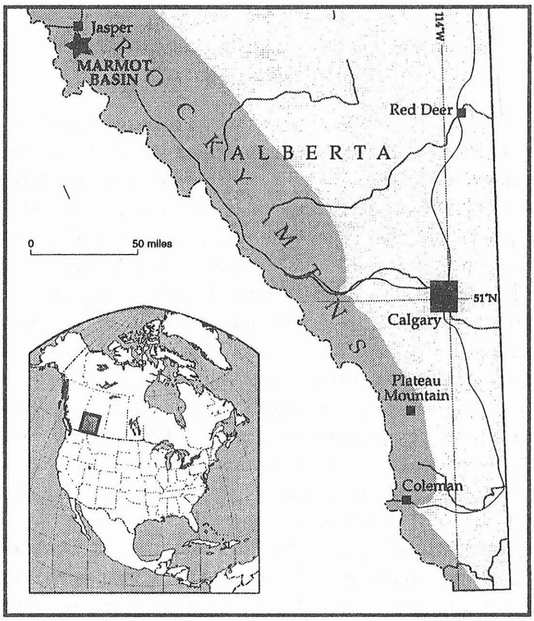
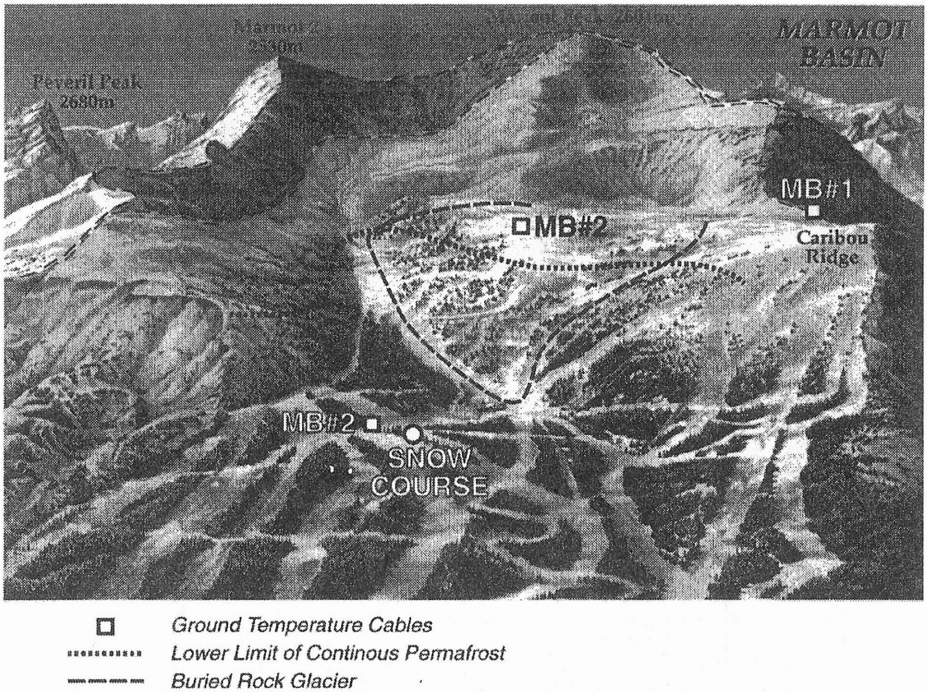


Fig. 1. Location of the Marmot Basin Ski Area in southwestern Alberta

SITE CHARACTERISTICS

MB #2 was drilled on a bench sloping at 6° southeast and below the steep (28–33°) upper slopes of Marmot Mountain (Pl. 1), at latitude 52°47' 36.5" N, longitude 118°06' 45.9" W. The elevation is 2,195 m and the drill hole lies about 10 m above the upper limits of tree line in the area. The vegetation consists of alpine meadow with a 90% ground cover of a wide variety of herbs, grasses, sedges and lichens (Tab. I).

The mean annual air temperature (1979–1998) is –2.36°C, but there has been a gradual decrease of 0.7°C at this site during this period of time (Fig. 2). The mean annual snowfall (November to April inclusive) is 356 cm (1965–1998) at the Middle Chalet (see Pl. 1, 2 and Fig. 2). From 1977 to 1985 there was a period of low snowfall (c. 260 cm) and variability (±30



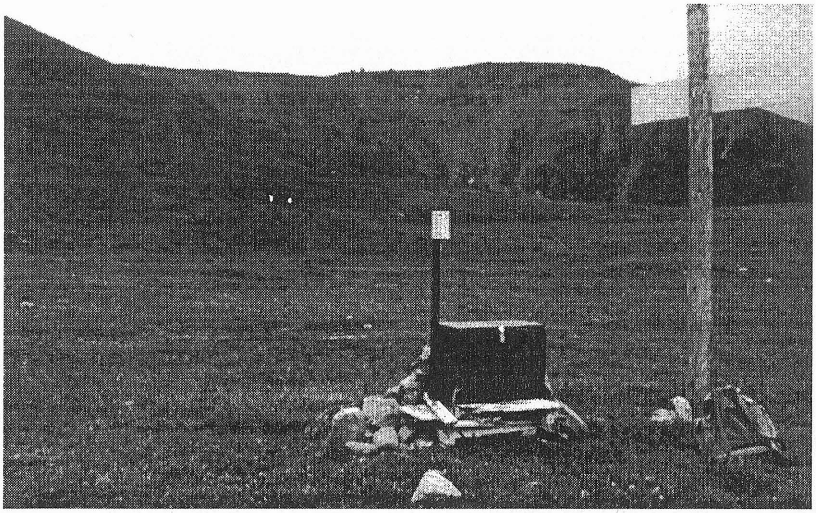
Pl. 1. The location of borehole MB#2 on Marmot Mountain, showing the approximate lower limit of continuous permafrost. The apparent shape of the pre-Late Wisconsin fossil rock glacier is also indicated

cm), followed by an increase to approximately 330 cm, accompanied by increased yearly variability (±90 cm). This variability changes independently of the snowfall amount.

The borehole passed through 1 m of Late Wisconsinan till and then entered large boulders lacking a mineral matrix. Drilling could only be achieved by alternate emplacement of cement and drilling, and this boulder pile extended to the bottom of the hole at 17 m. Sections nearby show a similar stratigraphy, and it appears that the boulders represent a fossil, pre- Late Wisconsin rock glacier whose extent is indicated in photo 1. Thus there is the possibility of interstitial advective/convective heat transfer by motion of moist air between the blocks.

METHODS USED

The hole was drilled using water as the lubricant and cooling agent, so that the ice and permafrost present were removed in the process. Addi-



Pl. 2. The weather station at MB #2

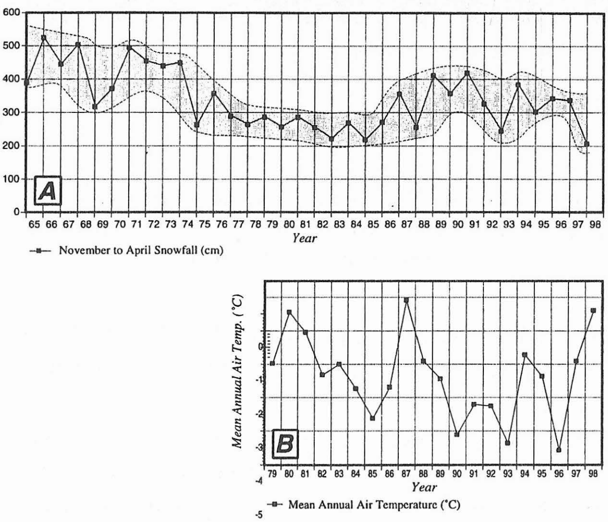


Fig. 2. A – Variation in mean winter snowfall (November to April inclusive) at the Middle Chaletbetween 1964 and 1998; and B – Variation in the mean annual air temperature at MB#2 during the study period

tional heat would also have been given out as the concrete used to fill the voids cured. A ground temperature cable was assembled using fifteen pair geophysical cable and six YSI 44033 thermistors. As soon as drilling was completed, the cable was put down the hole and allowed to freeze in place. The hole closed up around the cable in the upper 1 m during the following summer. Sealing of the thermistors inside the cable was achieved using silicone sealer and an outer sleeve cemented on with plastic solvent. This did not prove to be completely waterproof, so that the lowest two thermistors failed after about nine months, due to being below the main sub-permafrost water table in the blocks (approximately 12 m depth). Ground temperature measurements have been made every month since the drilling was completed. In addition, a Lambrecht three-level monthly soil temperature recorder was placed in a cedar box (Pl. 2). One level was used to record the air temperature inside an inverted can at 1.5 m, which had been painted white, while the other two channels recorded soil temperatures at 10 cm and 50 cm below the surface. The charts were changed every month, and the air temperature calibration was checked against the reading from a YSI 44036 thermistor which was also located in the inverted can. The resulting calibration curve was used to correct the air temperature recordings. The relative air humidity has been measured every 20 minutes by means of a data logger at MB#1 since 1994. On each site visit, the depth of snow above the borehole was noted, while the Marmot Ski Resort made measurements of daily snowfall between November and April at the Middle Chalet. The latter measuring site was located in the spruce forest at 1,985 m elevation and has not been affected by the resort development. It therefore provides a good record of the variations in total snowfall from year to year.

RESULTS

STRATIGRAPHY

The borehole penetrated one metre of Late Wisconsin till overlying a mass of boulders thicker than 17 m. The till serves to seal in the voids between the boulders, preventing air exchange with the atmosphere. A water table occurs at about 12 m depth, so that the air in the voids can form closed convection cells above this level capable of aiding in heat transfer through the boulders. Although the drilling thawed all the ice present originally, enough interstitial ice had accumulated in the upper 4.5 m by 1998 to produce a perched water table in the active layer.

THERMAL PROFILE

Figure 3A shows the geothermal envelope as measured by the upper four thermistors between 1979 and 1999. It is obvious that the thermistor placed at 4.75 m was incorrectly calibrated, but the long record permits the use of a suitable correction of –2.16°C. Figure 3B shows the effect of this modification.

The profile shows an unusually steep geothermal gradient between 4° and 11 m depth of 0.45°C per metre. There is also a "thermal offset"

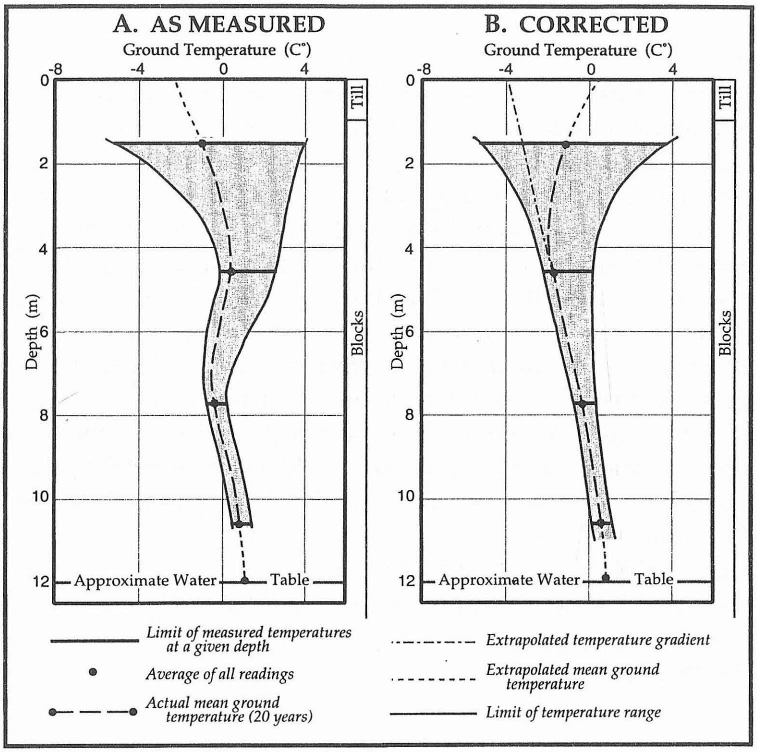


Fig. 3. The thermal envelope of ground temperatures: A – as measured by the thermistors, and B – after correction for a miscalibrated thermistor at 4.75 m depth

(KUDRYAVTSEV and MELAMED, 1972; BURN and SMITH, 1988) of about 4.1°C in the upper 4 metres. The bottom temperature is controlled by the presence of the water table, and the two thermistors below this indicated that the water had a fairly uniform temperature of 1.1°C. These thermistors operated for about nine months prior to failure due to water around them inside the protective coat.

The shape of the thermal envelope is also distinctive. In the upper 4 m, it takes the form of a bell-shaped curve which is normal for conductive heat transfer. Between 4 and 9 m depth, the envelope has the form of an inverted cone, which has been described from situations where heat advection occurs (HARRIS, 1996; HARRIS, PEDERSEN, 1998). Below 9 m, the width of the envelope increases slightly, presumably in response to variations in temperature of the water in the water table and heat conduction from it.

TEMPORAL VARIATIONOF READINGS OF INDIVIDUAL THERMISTORS

Figure 4 shows the temporal variations in the four thermistor readings over the 20-year period. The data for the thermistor at 4.75 m depth is shown after correction. There is clear evidence of a relatively uniform ground temperature at each level, punctuated by short waves of heating and cooling. Those waves coming from the surface are clearly related to the seasonal heating and cooling waves associated with summer and winter. Typically the cooling wave peaks in February at 1.52 m, in late July or August at 4.75 m, and in late July to November at 7.62 m, although there is some variability from year to year.

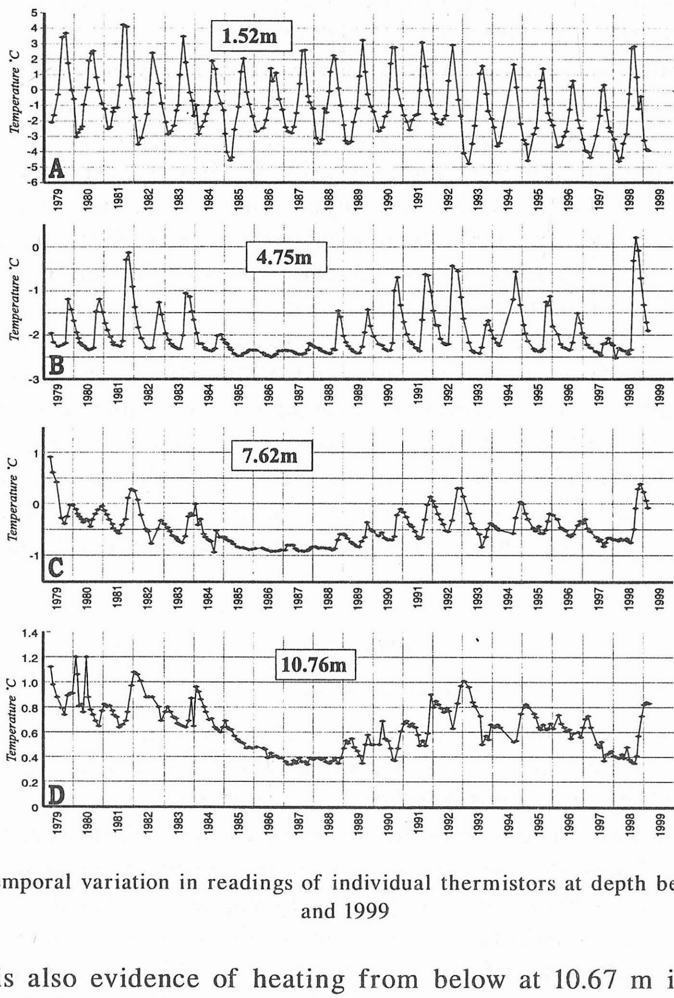


Fig. 4. Temporal variation in readings of individual thermistors at depth between 1979 and 1999

There is also evidence of heating from below at 10.67 m in late 1979, 1981, 1991, 1992 and 1998. The amount of the heating varies from event to event, and in 1998 the entire profile thawed briefly. These appear to be fairly random events and are independent of the size and duration of the heating wave in any one year. Intense surface heating waves not reflected in temperature changes at depth were encountered in 1983, 1987, 1988 and 1989.

TEMPORAL VARIATIONS IN GEOTHERMS

Figure 5 shows the temporal changes in geotherms in relation to snow cover and mean annual air temperature. There is a better qualitative correlation between depth of winter snow cover and ground temperature than with air temperature. In general, low or late winter snow covers permit greater heat loss from the ground, e.g. during the winter of 1997–1998, which was actually one of the warmest winters on record. Conversely, a deep or early snow cover minimizes heat loss, resulting in reduced winter cooling of the surface layers of the ground.

The temporal changes also clearly show the independence of the normal seasonal temperature fluctuations originating at the surface from the random heat input from below. Figure 6 shows the detail of the thawing and refreezing event in 1998 and early 1999. Thawing commenced at the ground surface, reaching 1.52 m in August and 4.75 m in October, and penetrating completely through the permafrost in November and December. By November, the ground at 1.52 m and 4.75 m had refrozen, and this freezing back is gradually extending downwards.

Also shown in figure 5 are the variations in thickness of the active layer and permafrost between 1979 and 1999. When the borehole was drilled, the pre-existing permafrost around the drill hole was melted. For figure 5, permafrost was defined as ground that is cryotic for more than two years. Substantial variations in thickness of both the permafrost and active layer occur, which are summarized in table II. Also shown in figure 5 and table II are the variations in temperature of the permafrost with time.

DISCUSSION

RELATIONSHIP TO ENVIRONMENTAL CONTROLS

The site is typical of benches on the mountain sides below steep slopes and above tree line in the Cordillera of North America. Actual winter snow cover (Fig. 5) is considerably less than the mean winter snowfall (Fig. 2) due in part to redistribution of snow by wind from these relatively exposed sites. Snow is also removed from the more exposed parts of the steep mountain slopes, to be re-deposited in gullies, depressions and in the upper part of the subalpine forest. This results in earlier thawing of the snow cover on the south-facing exposed slopes. Meltwater moves downslope, tending to warm as it moves, and then flows more slowly over the ground surface on the gently sloping bench underneath its snow pack. This helps in thawing the upper part of the active layer as well as melting the

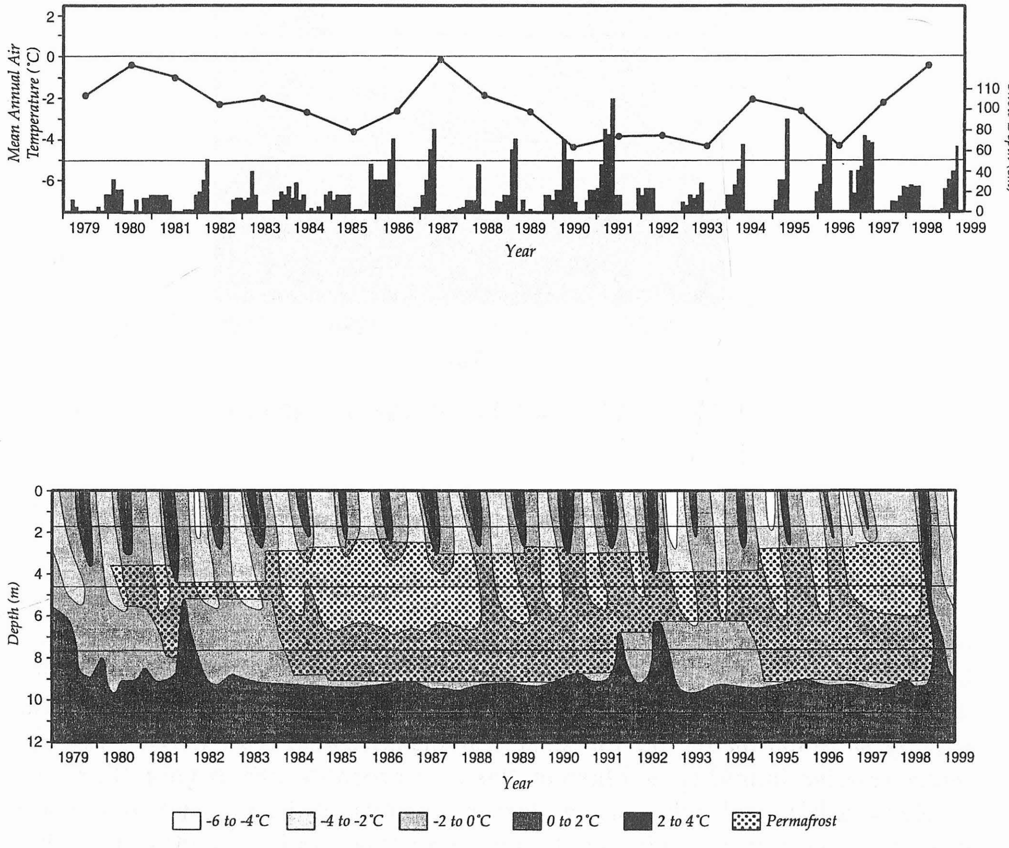


Fig. 5. Temporal variation in geotherms between 1979 and 1999, compared with snow depth and mean annual air temperature at MB #3. Also shown are the apparent variations in thickness in permafrost based on the definition of two years with temperatures continuously below 0°C

overlying snow pack. The warm water tends to disappear into the active layer in the process.

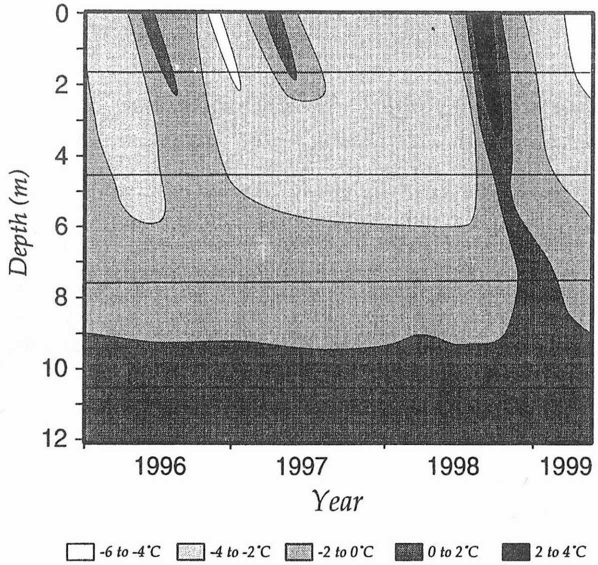


Fig. 6. Detail of the temporal variations in geotherms from 1997 to 1999 showing the evolution of the thawing and refreezing of the permafrost at MB#3

The other factor controlling the actual thickness of the snow pack is the sublimation of snow. The mean relative humidity of the air between October and April (1994–1998) is 87%, varying between 69% and 99% for a given month. HARRIS (1973) showed that 50% of the winter snowfall was lost by sublimation in the forest at Kananaskis, near Banff, Alberta. The mean relative humidity at Marmot Basin is probably 10% higher than that at Kananaskis, and sublimation during periods of high wind and lower relative humidity is undoubtedly an important cooling factor. In subliming, 1 gm of ice at 0°C would absorb about 620 calories of heat, so that this process will also help cool the surface of the snow pack and hence the ground underneath. In comparison, evaporation of 1 gm of soil moisture in summer only removes 540 calories of heat, but this has been shown to be an important cooling process for the soil in summer (OUTCALT *et al.*, 1998).

HEAT TRANSFER PROCESSES

Examination of figure 5 shows that timing, the amount of snowfall and the thickness of the actual snow cover are more important than small changes in actual mean annual air temperature in determining the degree of winter cooling of the ground. The shape of the ground temperature envelope (Fig. 3B) indicates that the upper 4.5 m of the profile is cooled dominantly by conduction of heat, producing a bell-shaped curve. Conduction of heat through the till is no problem, but heat conduction through the upper part of the blockfield requires an icy matrix in this zone.

Below 4.5 m but above 9 m, the thermal envelope takes the form of an inverted cone, suggesting advective cooling by air circulation (see HARRIS, 1996; HARRIS, PEDERSEN, 1998). This is probably by a convection movement of air within the spaces between the boulders, moving heat from the water table at 12 m upwards into the lower part of the permafrost zone. There it would deposit ice, providing a source for the interstitial ice in the zone above. This would be a closed-cell convection of the type being tested for cooling road beds in Alaska (GOERING, 1998). Growth of this interstitial ice would be limited to the lower limit of permafrost at any given time. The result is the extraordinarily steep geothermal gradient (0.45°C per m) which is two orders of magnitude steeper than most geothermal gradients resulting from conductive heat flow, e.g. LACHENBRUCH, MARSHALL (1969) and GOLD, LACHENBRUCH (1973) in Alaska. It is also quite different from the open-cell, advective heat exchange documented in the surface of a blockfield south of Plateau Mountain by HARRIS and PEDERSEN (1998). There the geothermal gradient is far lower than that resulting from conductive heat flow in the adjacent soil.

At 12 m, the regional water table is encountered. Since the water has a fairly constant temperature of about 1.1°C, this acts as a relatively constant source of heat. Above it, there are signs of heat flow by conduction as well as by air convection between 9 and 12 m depth, based on the ground temperature envelope.

WATER

In a permafrost environment, water represents a source of heat, and since it is mobile, it can represent an important form of heat flux. At Jasper, we have to deal with three forms, viz.: summer rainfall, meltwater from snow, and groundwater. Summer rainfall was not measured at the site, but some data is available for Jasper Park East Gate (Monthly Weather Normals A.E.S.). The total rainfall between May and October is usually twice the total winter precipitation as snowfall, so it must be considered. OUTCALT *et al.* (1998) noted that it could either warm or cool the ground. Meltwater from snow has already been described.

Water tables and groundwater are often disregarded in heat flow studies, e.g. LACHENBRUCH, MARSHALL (1969) and GOLD, LACHENBRUCH (1973). In wet, lowland sites, this may be valid since the entire profile is waterlogged. Even so, differences in water content in the profile may well cause differences in thermal properties which should affect the geothermal gradient. LUNARDINI (1998) regarded convective heat transfer or the thawing of frozen soil as being dependent on the water velocity, but concluded that it generally should have little effect on the thaw process. This conclusion is probably limited in application to lowland arctic situations.

In hilly or mountainous areas, the water table is usually present at depth. TOLSTIKHIN (1940) divided groundwater in permafrost regions into suprapermafrost water, intrapermafrost water and subpermafrost water. Suprapermafrost water consists of the perched water table at the base of the active layer or on the surface of the thawing front. It is almost always present and represents an important reserve of water for tundra plants, e.g. on the Tibetan Plateau. Evaporation of this water also cools the soil in summer (OUTCALT *et al.*, 1998). Intrapermafrost water has been found in an active rock glacier in the Swiss Alps (VONDER MÜHLL, 1992), where it produces an intrapermafrost talik.

Subpermafrost water is related to the regional water table. At Marmot Basin #2 it lies at 12 m and represents a relatively stable source of heat at depth. The stable temperature is presumably due to the integration of geothermal heat flow from depth, surficial heat flow from the surface, and advection of heat in water from upslope. In order for this stable temperature to alter, a change in one of these controls is required, and certainly the most likely variable is the advection of heat in water from upslope or above.

SOURCES OF HEAT

There appear to be two main sources of variations in the geothermal heat flow regime (Fig. 5). The surface source consists of the seasonal variations in heat flow in and out of the ground, in response to summer and winter. These variations regularly extend down through the profile, although the variations are small below about 4 m. The second source is related to apparently random variations in the temperature of the groundwater.

A peculiarity of the data on annual active layer thickness is its high variability. This is far more than is seen in the same years at the other borehole (MB #1, see Pl. 1). Furthermore, the mean active layer thickness is almost double that at MB #1, although there is only a minor difference in summer air temperature. When the active layer thickness is plotted against the square root of the thawing index (°C/days), the data plot in a line virtually parallel to the axis representing the active layer thickness (Fig. 7). Linear regression analysis shows a reasonably good correlation and indicates that the active layer thickness is not dependent on the thawing index. It is quite different from the results of HINKEL, NICHOLAS (1995) and NELSON *et al.* (1998, Pl. 1), who found a reasonable correlation at lowland study sites in Alaska. These results imply that heat conduction

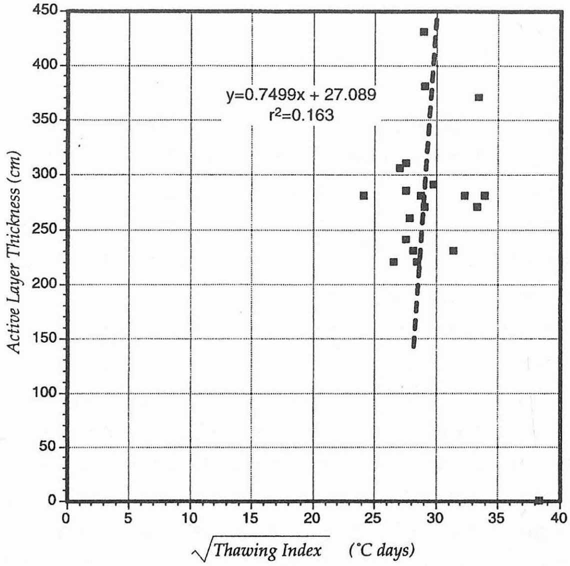


Fig. 7. Relationship between active layer thickness at MB #2 and the square root of the thawing index (°C/days)

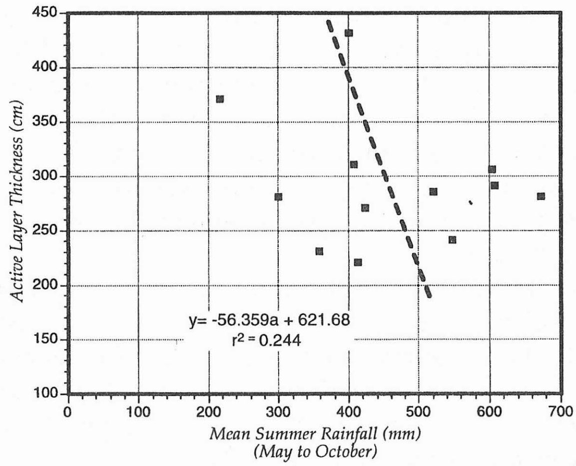


Fig. 8. Relationship between active layer thickness at MB #2 and mean summer rainfall (mm – May–October) at Jasper Park East Gate

through the soil is not responsible for the high annual variability, even though it is always an important process in the upper layers of the soil.

Two other explanations are possible. As suggested by OUTCALT *et al.* (1998), the summer precipitation could be involved in some way. In figure 8, the active layer thickness has been plotted against the mean summer rainfall (May–October) recorded at the Jasper East Park Gate weather station. Unfortunately the data is limited, but linear regression suggests that the summer rainfall does indeed affect the active layer thickness. When the summer precipitation is high, the active layer is thinner, presumably due to the higher moisture content in the active layer increasing evaporative cooling and resisting heating of the ground due to the high sensible heat capacity (specific heat) of the water.

In figure 8 there is considerable scatter, indicating that at least one other factor is involved. This is almost certainly the warm melt water from the adjacent mountain side, which would also explain the much deeper active layer on the bench than at MB #1, which is on the ridge-top. The sudden, random variations in the groundwater cannot be explained by the normal temperature regime in the active layer above it. If changes in geothermal heat flow were occurring, all the temperature cables in the area should show the same changes. MB #1 does not show these same events, so the variations must be specific to this site. The only reasonable explanation is the advection of limited quantities of warmer water into the groundwater, so producing a local thermal anomaly. This will slowly disappear and also move downslope with the groundwater flow. Above it, the base of the permafrost would rise substantially and quickly, only to sink to its former position when the heat source has gone.

It was unclear how this happened until 1998, when a pool of warm water perforated the permafrost table around the cable. When the thawed zone reached about 4.5 m, the warm water drained down to the water table and the hole through the permafrost table rapidly closed up (Fig. 6). The heating from below resulted in the rapid rise of the permafrost base by up to 3 m in two to three weeks, and this was followed by a very slow downward growth of this base, in spite of intense cooling from the ground surface during the winter.

This process is shown diagrammatically in figure 9. The random rapid warming events which affected the base of the permafrost in 1981–82, 1991–92, and 1992–93 presumably indicate that similar perforation of the permafrost table occurred a short distance upslope from the temperature cable in those years. The frequency of the perforation of the permafrost at a given place appears to be 1 in about 20 years at this site, while abrupt heating from below appears to occur once in every four years on average. This will obviously be site-dependent.

IMPLICATIONS FOR THE CALM PROJECT

The present study shows the complexity of the processes affecting the thickness of the active layer at a given site in the mountains. It suggests that they are closely related to the local topography, geology and hydro-

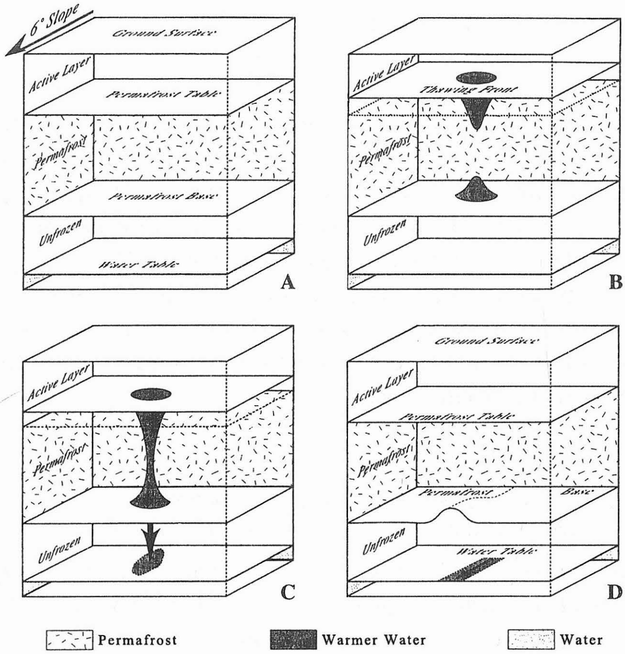


Fig. 9. Diagram showing the evolution of piercing of the permafrost by a pool of warmer water: A – The normal situation without the warmer water; B – The pool of warmer water thawing downwards through the permafrost; C – The pool of warmer water after it has drained down to the water table. Note that the perforation in the permafrost table is healing. D – The thawing of the permafrost base above the pool of warmer water as the latter moves slowly downslope

logy, in addition to the factors identified for arctic and subarctic lowlands (see for example, HINKEL, NICHOLAS, 1995; *et al.*, 1998; OUTCALT *et al.*, 1998; ANON, 1996). The difference in seasonal variations in ground temperatures between the lowland arctic sites and the MB #2 site is almost 2 orders of magnitude. It seems doubtful whether the data from a single site would be representative of a given mountain area. Multiple study sites related to the different parts of the landscape would be essential in order to obtain a clear picture of active layer thickness in the area. Furthermore, lowland sites which receive meltwater that has travelled downslope over the frozen ground may also show similar effects where the warmer water descends through the soil to a water table, e.g. on a flood plain or over terraces. Thus the processes found at MB#2 may occur at numerous places wherever the terrain is undulating.

IMPLICATIONS FOR SOIL CLASSIFICATION SYSTEMS

The extreme variability in active layer thickness from year to year in table II suggests that a classification system based on the different types of soil forming processes and the resulting horizonation (e.g. SOKOLOV, KONYUSHKOV, 1998) would be easier to use than one based primarily on active layer thickness (see for example, SLETTEN, 1997; KIMBLE, AHRENS, 1994; U.S. D.A., 1998). Using presence or absence of permafrost in the upper 1 or 2 m of the soil as a basis for a major soil order, gelisols, leads to the need to reclassify the soils from year to year in soils showing this degree of variability in depth of thaw in summer. It is not generally possible to re-examine depth of thaw everywhere on a map sheet every summer, so the use of the Russian system with modifiers to indicate presence of permafrost would be preferable. A separate modifier could be used for soils with intermittent permafrost.

DETERMINING EQUILIBRIUM CONDITIONS AND ESTABLISHING CHANGES IN GROUND TEMPERATURE

Figure 6 indicates that for sites where there is heat input into the ground by meltwater from upslope, it is necessary to watch the base temperatures between seasonal heat inputs. The data shows a tendency of the ground to cool over the study period by about 0.4°C at all depths studied. The 0.3°C differential between the air and the ground is probably due to the higher winter precipitation since 1985, producing greater quantities of snow melt and runoff in the spring. It would also reduce the heat loss during winter. This emphasizes the importance of monitoring winter precipitation as well as mean annual air temperature. Only by having monthly ground temperature measurements can equilibrium conditions be recorded. Establishing whether ground temperatures are in equilibrium with a given climate on mountain slopes may take more data and more years than on ridge-tops or at lowland arctic sites.

CONCLUSIONS

Marmot Basin #2 is a borehole through 1 m of till into a mass of boulders. The water table occurs at 12 m depth and continuous permafrost extends down to about 9.5 m depth. Seasonal thawing of the active layer is up to 4.5 m, but varies widely from year to year. The depth of thawing appears to be controlled by a combination of the amount and warmth of meltwater descending the mountain side onto the bench in Spring and also the quantity of summer precipitation. Higher rainfall correlates with a thinner active layer, accounting for 24% of the variation, but there is a very poor correlation between thawing degree days and active layer thickness. The meltwater appears to explain 60% of the variation from year to year.

Processes of heat transfer include dominant conduction in the upper 4.5 m, suggesting that the boulders above this level are in a matrix of ice. Below this, closed-cell convection of air efficiently transfers heat, producing a geothermal gradient of 0.45°C/m. Periodically, thawing of the lower 3 m of the permafrost may occur due to perforation of the permafrost table by small pools of warmer meltwater. The perforation quickly closes, but the pool of warm water will keep the permafrost base elevated above it until it either moves away downslope or the excess heat is dissipated.

Time and amount of winter snow cover are the dominant controls of degree of winter cooling, i.e., if the mean annual air temperatures are suitable for permafrost to develop, it is the precipitation regime (summer and winter) which determines the degree of permafrost development.

Multiple study sites related to the differential parts of the landscape would appear to be essential in order to obtain a clear picture of active layer thickness in an undulating area, unlike the sampling pattern of the CALM Project as applied to arctic and subarctic lowlands. The extreme variability in annual depth of thaw also makes it difficult to use the new U.S.D.A. (1998) classification of the gelisol order in mountainous areas.

ACKNOWLEDGEMENTS

Funding for the research was provided by the former Building Research Division of the National Research Council of Canada and by N.S.E.R.C. operating grant #7483. Marmot Basin Ski-Lifts Ltd. kindly provided logistical help as well as the data on winter snowfall. Parks Canada are thanked for their permission to carry out the project and collect the appropriate data.

References

ANON, 1996 – Circumpolar active layer monitoring (CALM). *Frozen Ground,* 20; p. 15–16.

BURN, C. F., SMITH, C. A. S., 1988 – Observations of the "thermal effect" in near-surface mean annual ground temperatures at several sites near Mayo, Yukon Territory, Canada. *Arctic*, 41; p. 99–104.

GOERING, D. J., 1998 – Experimental investigation of air convection embankments for permafrost-resistant roadway design. *Proceedings of the 7th International Conference on Permafrost, Université Laval. Collection Nordicana*; p. 319–326.

GOLD, L. W., LACHENBRUCH, A. H., 1973 – Thermal conditions in permafrost: A review of North American Literature. *Permafrost: 2nd International Conference*, North American Contribution. National Academy of Sciences, Washington, D.C; p. 3–23.

HARRIS, S. A., 1973 – Annual soil moisture regimes in the rooting zone across the prairie-forest boundary in south-west Alberta. *Journal of Soil Science*, 25; p. 448–460.

HARRIS, S. A., 1986 – Permafrost distribution, zonation, and stability along the eastern ranges of the Cordillera of North America. *Arctic*, 39; p. 29–38.

HARRIS, S. A., 1996 – Lower mean annual ground temperature beneath a block stream in the Kunlun Pass, Qinghai Province, China. *Proceedings 5th Chinese Permafrost Conference, Lanzhou;* p. 227–237.

HARRIS, S. A., BROWN, R. J. E., 1982 – Permafrost distribution along the Rocky Mountains in Alberta. *Proceedings of the 4th Canadian Permafrost Conference*, Calgary, Alberta. National Research Council of Canada, Ottawa; p. 59–67.

HARRIS, S. A, PEDERSEN, D. E., 1998 – Thermal regimes beneath coarse blocky materials. *Permafrost and Periglacial Processes,* 9; p. 107–120.

HINKEL, K. M., NICHOLAS J. R. J., 1995 – Active layer thaw rate at a boreal forest site in central Alaska, U.S.A. *Arctic and Alpine Research*, 27; p. 72–80.

KIMBLE, J. M., AHRENS, R. J., 1994 – Proceedings of the meeting on the classification, correlation and management of permafrost-affected soils. United States Department of Agriculture, Soil Conservation Service; pp. 232.

KUDRYAVTSEV, V. A., MELAMED, V. A., 1972 – Asymmetry of the enveloping curves of temperature fluctuations in soil in relation to calculations of depths of seasonal and perennial freezing [in Russian]. *Permafrost Research*, 12; 308 p. Moscow State University.

LACHENBRUCH, A. H., MARSHALL, B. V., 1969 – Heat flow in the Arctic. *Arctic*, 22; p. 300–311.

LUNARDINI, V. J., 1998 – Effect of convective heat transfer on thawing of frozen soil. *Proceedings of the 7th International Conference on Permafrost. Collection Nordicana*; p. 689–695.

NELSON, F. E., OUTCALT, S. I., BROWN, J., SHIKLOMANOV, N. I., HINKEL, K. M., 1998 – Spatial and temporal attributes of the active-layer thickness record, Barrow, Alaska, U.S.A. *Proceedings of the 7th International Conference on Permafrost, Université Laval. Collection Nordicana;* p. 797–802.

OUTCALT, S. I., HINKEL, K. M., NELSON, F. E., MILLER, L. L., 1998 – Estimating the magnitude of coupled–flow effects in the active layer and upper permafrost, Barrow, Alaska, U.S.A. *Proceedings of the 7th International Conference on Permafrost, Université Laval. Collection Nordicana;* p. 869–873.

SLETTEN, R. S., 1997 – Permafrost-affected soils: consideration in classification hierarchy. In: *Cryosols in Classification Hierarchy. Russian Academy of Sciences, Novosibirsk, "Nauka"*; p. 23–29.

SOKOLOV, I. A., KONYUSHKOV, D. E., 1998 – Soils and the soil mantle of the northern circumpolar region. *Eurasian Soil Science*, 31; p. 1179–1193.

TOLSTIKHIN, N. I., 1940 – The regime of ground and surface waters in the region of permafrost distribution [In Russian]. *In*: M. I. SUMGIN *et al. General Geocryology*, Moscow. Akademia Nauk. SSSR; pp. 340.

U.S.D.A. 1998 – *Keys to Soil Taxonomy.* United States Department of Agriculture, Natural Resources Conservation Service. 8th Edition; pp. 326.

VONDER, MÜHLL, D. S., 1992 – Evidence of intrapermafrost groundwater flow beneath an active rock glacier in the Swiss Alps. *Permafrost and Periglacial Processes,* 3; p. 169–173.

Table I

Vascular plants and lichens found in the vicinity of Marmot Bassin # 2

Family	Species
Parmeliaceae	*Cetraria ericetorum* *Flavocetraria nivalis*
Peltigeraceae	*Peltigera aphthosa*
Lycopodiaceae	*Lycopodium alpinum* *Lycopodium annotinium*
Equisetaceae	*Equisetum scirpoides*
Graminae	*Calamagrostis inexpansa* *Festuca brachyphylla* *Festuca saximontana* *Poa alpina* *Poaarctica* *Poa cucksikii*
Cyperaceae	*Carex albo-nigra* *Carex phaeocephala* *Carex raymondii* *Carex saxatilis* *Carex spectablis*
Juncaceae	*Juncus drummondii var. drumondii* *Luzula parviflora* *Luzula spicata* *Luzula wahlenburgii*
Salicaceae	*Salix arctica ssp. arctica* *Salix reticulata ssp. nivalis*
Polygonaceae	*Oxyria digyna*
Portulacaceae	*Claytonia lanceolata*
Caryophyllaceae	*Minuartia rubella*
Ranunculaceae	*Anemone occidentalis* *Ranunculus eschscholtzii* *Trollius albiflorus*
Saxifragaceae	*Saxifraga bronchialis sp. austromontana* *Saxifraga caespitosa ssp. uniflora*
Rosaceae	*Dryas hookeriana* *Dryas octapetala*

	Potentilla diversifolia *Potentilla uniflora*
Leguminosae	*Oxytropis podocarpa*
Onagraceae	*Epilobium angustifolium* *Epilobium latifolium*
Ericaceae	*Arctostaphylos rubra* *Arctostaphylos uva-ursi* *Cassiope mertensiana* *Cassiope tetragona ssp. saximontana* *Kalmia microphylla* *Phyllodoce empetriformis* *Phyllodoce glanduliflora* *Phyllodoce intermedia*
Gentianaceae	*Gentiana glauca* *Gentianella propinqua*
Polemoniaceae	*Phlox hoodii*
Scrophulariaceae	*Castilleja occidentalis* *Veronica wormskoldii ssp. wormskoldii*
Valerianaceae	*Valeriana sitchensis*
Compositae	*Antennaria lanata* *Antenanria media* *Arnica diversifolia* *Arnica latifolia* *Arnica mollis* *Artemisia norvegica ssp. saxatilis* *Crepis nana* *Erigeron acris ssp. debilis* *Erigeron humulis* *Erigeron peregrinus ssp. callianthemus* *Erigeron speciosus* *Senecio pauciflorus* *Solidago multiradiata var. scopulorum* *Taraxicum ceratophorum*

Table II

Annual variations in the characteristics of the permafrost in relation to thawing index for borehole MB #2 from 1979 to 1998

Year	Permafrost Table (m)	Permafrost Base (m)	Permafrost Thickness (m)	Active Layer Thickness (m)	Temperature °C at 4.57 m depth max	min.	Thawing Index °C days
1979	n.d.	n.d.	n.d	3.7	−1.18	−2.26	1113
1980	n.d.	n.d.	n.d	3.05	−1.18	−2.33	734
1981	3.5–4.3	5.2–7.95	1.95–4.25	4.3	−0.13	−2.26	841
1982	4.3	5.2	0.9	2.85	−1.26	−2.31	c.760
1983	4.31–2.85	5.2–6.9	0.9–3.8	2.8	−1.05	−2.32	587.5
1984	2.85–2.8	6.9–8.8	6.05–5.8	2.2	−1.99	−2.36	813
1985	2.8–2.7	8.8–9.05	6.05–6.6	2.3	−2.34	−2.47	793
1986	2.7–2.8	9.05	6.6–6.8	2.4	−2.34	−2.50	760
1987	2.8–3.1	9.05	5.95–6.8	3.1	−2.19	−2.43	1317.5
1988	3.1	9.05	5.95	2.7	−1.45	−2.41	1111.5
1989	3.1–2.8	9.05	5.95–6.45	2.8	−1.42	−2.40	1147
1990	2.8–2.9	8.75–0.05	5.7–6.45	2.9	−0.69	−2.35	890
1991	2.9	6.8–8.75	3.8–5.7	2.8	−0.63	−2.36	831
1992	2.9–3.8	6.2–6.8	2.4–3.8	3.8	−0.43	−2.22	844.5
1993	3.8	6.2	2.4	2.2	−1.68	−2.41	705.5
1994	3.8–2.75	6.2–7.3	2.4–5.1	2.8	−0.57	−2.24	1243.5
1995	2.75	7.3–9.1	5.1–6.3	2.7	−1.25	−2.37	845.5
1996	2.7	9.1	6.5	2.6	−1.52	−2.34	778
1997	2.3	9.1	6.8	2.3	−1.96	−2.45	989
1998	Nil	Nil	Nil	Nil	0.2	−2.43	1474
Mean	3.13 m	7.31 m	4.88 m	2.86 m	−1.25°C	−2.36°C	928.9
n	17	17	17	19	20	20	20