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GEO-HYDROLOGIC AND THERMOMETRIC OBSERVATIONS IN THE VICINITY OF THE COLUMBIA ICEFIELD, ALBERTA AND BRITISH COLUMBIA, CANADA

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Abstract. The Columbia Icefield rests upon limestones containing natural caves that drain waters from the glacier sole. The principal cave is sealed at one end by an extrusion of glacier ice 300 m below the icefield surface. The hydrologic regime of the cave indicates that the modern icefield is temperate in character and that water is present at the glacier sole throughout the year. An interpretation of the air temperature pattern in the cave suggests that the geothermal flux to the glacier is only 10–40% of the expected value because heat is abstracted by melt water circulating through the rock. U, Th and O isotopic analyses of calcite speleothems further indicate that the base of the icefield has probably been temperate throughout the past 150 000 years. The cave was inundated when glaciers expanded during the classical Wisconsinan–main Würm period. The inundation implies maintenance of a permanent water table at some hundreds of meters above the base in a valley glacier 750–800 m in depth.

Résumé. *Observations géo-hydrologiques et thermométriques au voisinage du Columbia Icefield, Alberta et Colombie Britannique, Canada.* Le Colombia Icefield repose sur des calcaires contenant des cavités qui drainent les eaux issues du lit glaciaire. La principale cavité est obstruée à une extrémité par une extrusion de glace de glacier à 300 m en-dessous de la surface du sol. Le régime hydrologique de la cavité indique que l'appareil glaciaire actuel est de caractère tempéré et que l'eau est présente sur le lit glaciaire tout au long de l'année. Une interprétation de la distribution des températures de l'air dans la grotte fait penser que le flux géothermique vers le glacier est seulement de 10 à 40% de la valeur attendue, parce que les calories sont évacuées par la circulation d'eau de fusion à travers la roche. Des analyses isotropiques de l'Uranium, du Thorium et de l'Oxygène dans la calcite des stalagmites et stalactites, confirment que la base de la calotte glaciaire était probablement tempérée depuis les 150 000 dernières années. La cavité fut immergée lors de l'expansion des glaciers pendant la période classique du Wisconsin–Würm central. Cette submersion implique l'existence d'un niveau d'eau permanent à quelques centaines de mètres au-dessus du fond d'une vallée glaciaire de 750 à 800 m de profondeur.

Zusammenfassung. *Geohydrologische und thermometrische Beobachtungen in der Umgebung des Columbia-Icefields, Alberta und British Columbia, Kanada.* Das Columbia-Icefield liegt auf Kalkgestein, in dessen natürlichen Hohlräumen Wasser von der Gletschersohle abfliesst. Die Haupthöhle ist an ihrem einen Ende durch eine Eisauspressung 300 m unter der Eisfeldoberfläche verschlossen. Der Wasserhaushalt in der Höhle deutet darauf hin, dass das heutige Eisfeld temperiert ist und dass Wasser an der Gletschersohle das ganze Jahr hindurch vorhanden ist. Eine Interpretation des Lufttemperaturgefüges in der Höhle lässt vermuten, dass der geothermische Wärmestrom zum Gletscher nur 10–40% des Erwartungswertes beträgt, da Wärme durch das im Gestein zirkulierende Schmelzwasser abgeführt wird. U-, Th- und O-Isotopenalysen von Kalktropfsteinen weisen ferner darauf hin, dass die Basis des Eisfeldes vermutlich während der letzten 150 000 Jahre temperiert war. Die Höhle wurde überschwemmt, als die Gletscher während der klassischen Wisconsin–Hauptwürm-Eiszeit vorstiessen. Während der Überschwemmung heilt sich der Wasserspiegel ständig einige hundert Meter über der Basis eines 750–800 m dicken Talgletschers.

In the southern Rocky Mountains of Canada a majority of modern glaciers are small and of the cirque, valley, or bench types. However, in the central area of the mountains resistant rock strata dip more gently than elsewhere, giving rise to some plateaus at high altitude. These in turn support larger ice masses, comprising small ice caps and highland ice expanses. The Columbia Icefield is the largest of them. Its massif central is 8 km in diameter and ranges in altitude 2 000–3 500 m a.s.l. (Fig. 1). It is drained by four of the largest surviving valley glaciers, the South, Columbia, Athabasca and Saskatchewan. The two latter have received substantial glaciological study (Meier, 1960; Paterson, 1964, etc.), but the Icefield itself is little known because it is comparatively remote and intensively crevassed in the southern portions. There are no published determinations of the thickness of the ice. The mean position of the firn line is not established: from our own sporadic observations it has lain between the 8 500 and 9 500 ft (2 600 and 2 900 m) contours in Figure 1 during a majority of the melt seasons since 1967.

Mount Castleguard abuts the eastern side of the Icefield. From its base at 1 740 m a.s.l. up to 2 500 m, the mountain is composed of Middle Cambrian formations of limestone and dolomite (Ford, 1971). The strata are singularly massive and resistant to mechanical erosion and compose a great staircase of benches and scarps. From the geological trends, these rocks must also underly most of the Icefield, which is highly crevassed because it rests upon staircase topography.

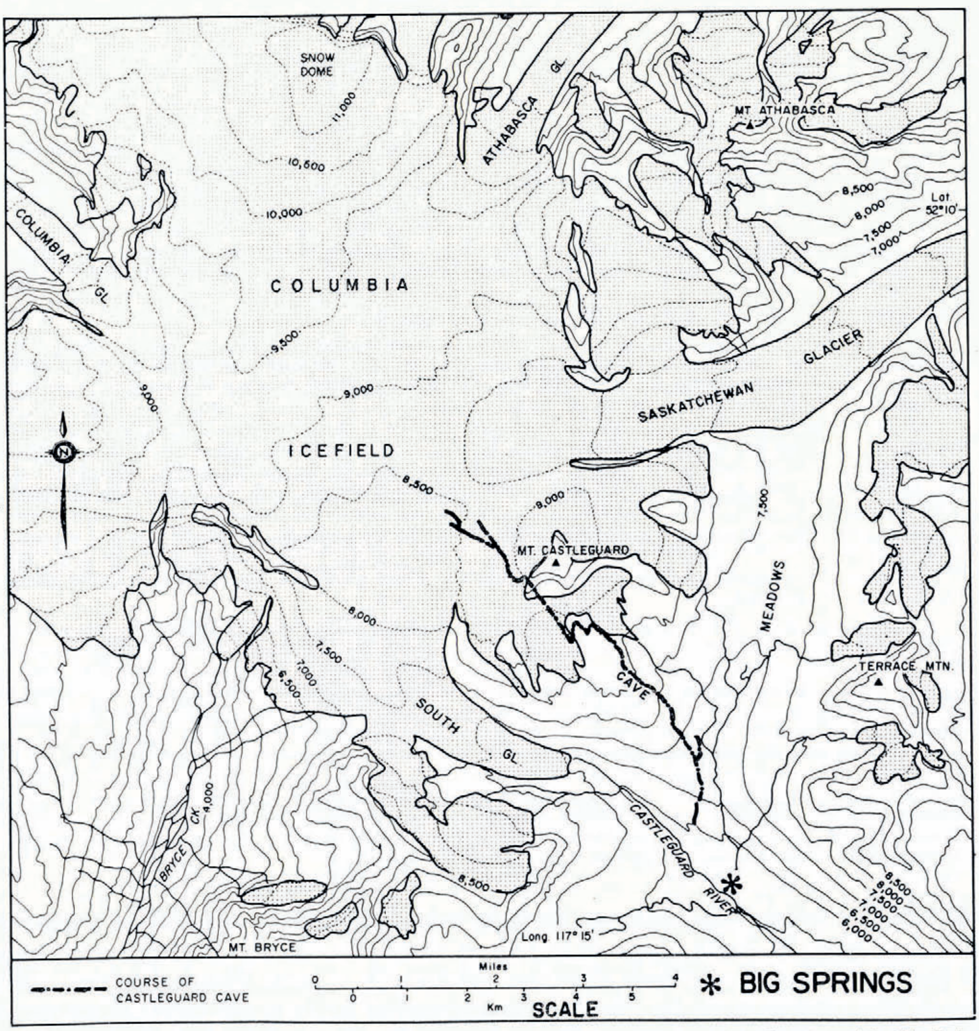

Fig. 1. The Columbia Icefield and environs showing the course of Castleguard Cave and the location of the Big Springs.

On Mount Castleguard the uppermost limestone benches support a series of small bench and cirque glaciers that surround the summit cone, which is composed of shales and sandstones. These glaciers are probably thin ($\leqslant$150 m). Recently they have receded an average of 500 m from prominent "neoglacial" moraines. Early photographs of the area (*c.* 1923) show the glacier termini still at the moraines, albeit downwasted.

The course of Castleguard Cave is shown on the plan in Figure 1 and in schematic section in Figure 2. It is a solution cavern in the limestones. The only entrance accessible to explorers is at the south-east end, in the flank of the Castleguard River valley at 1 970 m a.s.l. From there, the principal passage extends north-west for 9.0 km, ascending 320 m. It passes beneath Mount Castleguard and its glaciers and at the limit of current exploration has penetrated beneath the central Icefield for a distance of 1.5 km. No other cavern that is known makes so extensive a penetration beneath extant glaciers. Castleguard Cave is ancient. It affords opportunities for observations of past and present conditions prevailing at the sole of the Icefield and the Mount Castleguard glaciers, insofar as these are recorded at some depth in the underlying rock. The purpose of this paper is to present pertinent results of a series of expeditions since exploration began in 1967.

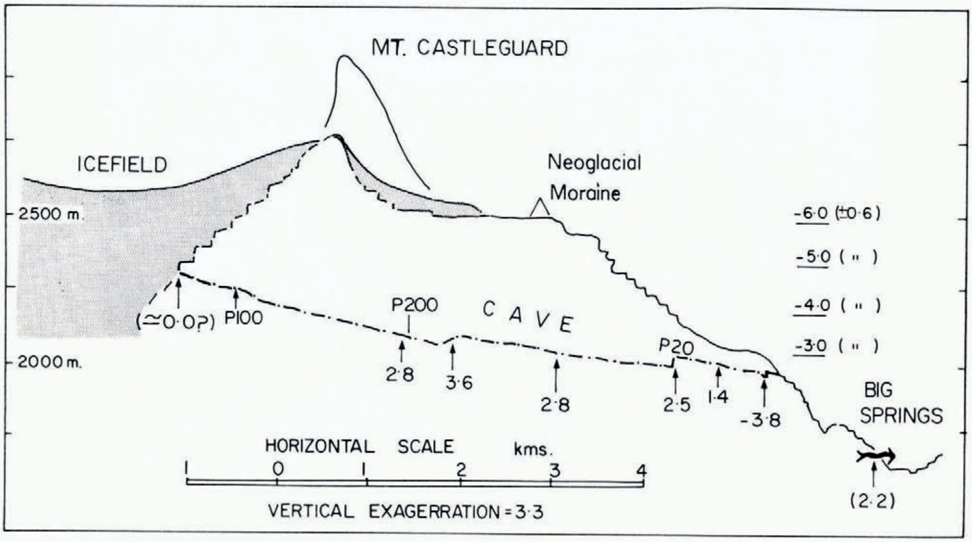

Fig. 2. Schematic north-west–south-east section through Mount Castleguard and Castleguard Cave. 2.8, etc. = dry bulb air temperature in the cave, degrees C. −4.0, etc. = calculated mean annual temperature of external air, degrees Celsius.

1. The cave and ground-water hydrology

Castleguard Cave is a simple, linear river cave created by groundwaters flowing north-west to south-east, following the direction of stratal dip. The principal passage is an alternation of vadose and phreatic parts; vadose caves are created by gravitational streams entrenching the floors of air-filled galleries; phreatic caves enlarge under conditions of complete and permanent water-fill. Dimensions of the passage are varied but average 15–20 m². A characteristic cross-section is shown in Figure 3.

At the up-stream (north-west) end the cave bifurcates at P100 (Fig. 2). The principal passage continues north-west for 800 m where it is sealed by ice (Fig. 4) which appears as a flat wall across the passage. The ice is slightly ablated around the perimeter where it contacts the rock, but it maintains an air-tight seal. No melt water was observed on the three occasions

Fig. 3. The principal passage of Castleguard Cave in the vicinity of P200. The circular cross-section is characteristic of phreatic solution. A trench 12 m deep has been carved in the floor by later vadose invasion waters. Laminated fines deposited during the flood event (see text) infill both phreatic and vadose sections and are now sapped down into the entrenchment. (Photograph by D. C. Ford.)

it has been visited (November 1970, April 1973, April 1974). The ice is very coarsely crystalline and appears to support one or more contained blocks of local limestone. It is entirely unlike the cave ice formed from ground waters or water vapour that is common in other caves of the Rocky Mountains and is interpreted as glacier ice of the Icefield extruded into the cave passage for an unknown distance. Its altitude is 2 295 m a.s.l. and the surface of the Icefield directly overhead is at 2 592 m, suggesting that the glacier is 297 m deep at this point.

Within 150 m of the ice blockage, stalactites of calcite are growing actively, i.e. are furnished with flowing water. Considering this observation and the condition of the ice itself, it is probable that the temperature of the extruded ice is 0°C or a very little below.*

* Attempts to get an accurate thermometer in working condition to this point have failed.

Fig. 4. The glacier ice seal at the north-west end of the principal passage, Castleguard Cave. The ice is slightly ablated around the perimeter but maintains an airtight seal. There is no melt water. The limestone block in the centre is supported by the ice. P. Thompson seated at left. (Photograph by A. C. Waltham.)

During the winter months a strong current of air flows to the north-west throughout the cave to P100. Here it leaves the principal passage and passes across a 30 m shaft into an inaccessible fissure. From there it must either pass into the base of the Icefield through exits that are not sealed air-tight by ice or follow an unknown gallery traversing further beneath the glacier.

Castleguard Cave has been abandoned by the ground-water river that created it or, alternatively, the river has been eliminated by growth of the Icefield in its catchment. From U series dating of speleothems (below) abandonment occurred $\geqslant$155 ka (1 ka = 10^3 years B.P.). The genetic river has been replaced by innumerable "invasion" waters. These are later affluents that chanced to drain across the pre-existing cave; in scale, they range from tiny seepages depositing stalactites and stalagmites to freshets discharging 1 $m^3 s^{-1}$ at maximum. Invasion waters appear everywhere along the principal passage save in the "cold zone", the first kilometre at the south-east end where a condition of permafrost limits seepage through the overlying rock. However, this zone is subject to complete flooding during the summer when waters rise up from a deeper tributary passage.

The discharge of all invasion waters is at a maximum in the summer: most have ceased to flow by the end of the winter season (April). On geological grounds, the sources of these waters lie north and north-west of the cave, i.e. they derive from the soles of the Mount Castleguard glaciers and from the eastern margin of the Icefield.

In the Castleguard River valley, 280 m below the south-east entrance of the cave and aligned upon it are the "Big Springs", which discharge from impenetrable bedding planes in the lower limestones. From considerations of the volume and carbonate chemistry of the waters, Ford (1971) has shown that the principal source of the springs ($\geqslant$80% of discharge) must be the base of central portions of the Icefield, probably much of the area lying between the 8 500 and 9 500 ft (2 600 and 2 900 m) contours in Figure 1. The waters must flow through a second limestone cave ("Castleguard II"), approximately underlying the known one. The known cave was therefore abandoned as a consequence of re-routing of ground waters rather than the alternative, their elimination by Icefield growth, that was mentioned above. Substantial active karst sink holes must exist beneath $>$300 m of ice today. On three of the four occasions that the Big Springs have been gauged (August 1969) the Icefield component of their discharge was estimated to exceed 8.5 $m^3 s^{-1}$. Most of the supposed source area was overlain by firn at the time. In winter, discharge is reduced to $<$0.1 $m^3 s^{-1}$ or *c.* 1% of known summer peak flows. Winter reduction of the invasion waters in Castleguard Cave is similar in magnitude.

The ground-water hydrological observations indicate that there is abundant water at the sole of the central Icefield and beneath the smaller Mount Castleguard glaciers during the summer months. Waters must derive from the melt of firn and surficial ice because discharge is very greatly reduced during the winter freeze up. However, it is not entirely eliminated: some waters are present at the sole of the Icefield even at the end of the winter. Melt waters enter the glaciers both above and below the firn line and are able to pass through $\geqslant$300 m of glacier ice in either situation to reach the sole. These points suggest that the modern ice masses considered have temperatures uniformly at the pressure melting point or very close to it.

2. Thermal features of the cave

Figure 2 displays a variety of thermal data pertaining to the cave and its environs. Temperatures within the cave are a representative selection of dry-bulb air temperatures measured in a traverse of April 1973 that was extended from the south-east entrance as far as P200. Measurements in April 1974 confirmed the pattern. The value $\simeq$0.0°C at the ice plug is that inferred above. Water emerging at the Big Springs in August 1969 had a mean temperature of 2.2°C with little variation.

There are no meteorological stations of long record close to the Icefield. Mean annual external air temperatures (Fig. 2) were calculated by a multiple linear regression analysis of data from 74 meteorological stations within a 280 km radius of the cave.*

A majority of known caverns are shorter and at shallower depth than Castleguard Cave. They possess an interior air temperature that changes very little from season to season and is close to the mean annual external temperature of the locality. Seasonal variation of cave air temperature (and, therefore, possible deviation of the annual mean from that prevailing outside) may be introduced by seasonal inflows of external air or by large streams of water. The problem has been analysed by Wigley and Brown (1971) who show that external air effects may penetrate 4–5X_0, where X_0 is a relaxation length that is a function of cave diameter and velocity of airflow. The relaxation length of water effects is some six times greater than that of air effects.

Throughout the winter season at Castleguard Cave there is a strong and constant flow of air from the south-east entrance up through the cave to P100 where, as noted, it passes into an inaccessible fissure. During the summer, this air flow is reversed. Water effects may be neglected because of the very small volume of the invasion waters compared to that of the cave.

* The relationship obtained was

$$\bar{T} = 10.9 - 0.00293z - 0.0283y \text{ (°C)}$$

where $\bar{T}$ is mean annual dry bulb temperature, z the altitude in ft and y the latitude in degrees N of 49° N. The multiple correlation coefficient is 0.977; the standard error is 0.6 deg.

Taking representative values of passage diameter and velocity of air flow at Castleguard Cave, $X_0 \approx$ 200–500 m, and the limit of penetration of seasonal thermal effects is therefore 800–2 500 m. The greater part of the cave, lying between P100 and P80, is insulated from them. If the cave were shallow, the air temperature in this interior sector would be expected to approximate the mean annual external air temperature of $-3.0\ (\pm 0.6)$°C. The record of April 1973 shows the temperature rising sharply through a zone of seasonal effects at the south-east end of the cave that is 1 800 m in length and more slowly thereafter to a maximum of $+3.6 \pm 0.2$°C.*

The only feasible source of heat to explain the warmth of the cave interior is geothermal. In Figure 5, geo-isotherms are drawn to accord to the measured and calculated temperatures, assuming the basal ice temperature of $\simeq 0$°C that is suggested by the hydrological observations. The geothermal gradient between the centre of the cave (+3°C) and the moraine 510 m above it (-5.9 ± 0.6 deg) is found to be 0.02 deg m^{-1}. This value is bracketed by the widely quoted range of 0.01–0.05 deg m^{-1} for geothermal gradients in non-volcanic areas (Carslaw and Jaeger, 1959). The schematic geo-isothermal pattern of Figure 5 therefore appears reasonable.

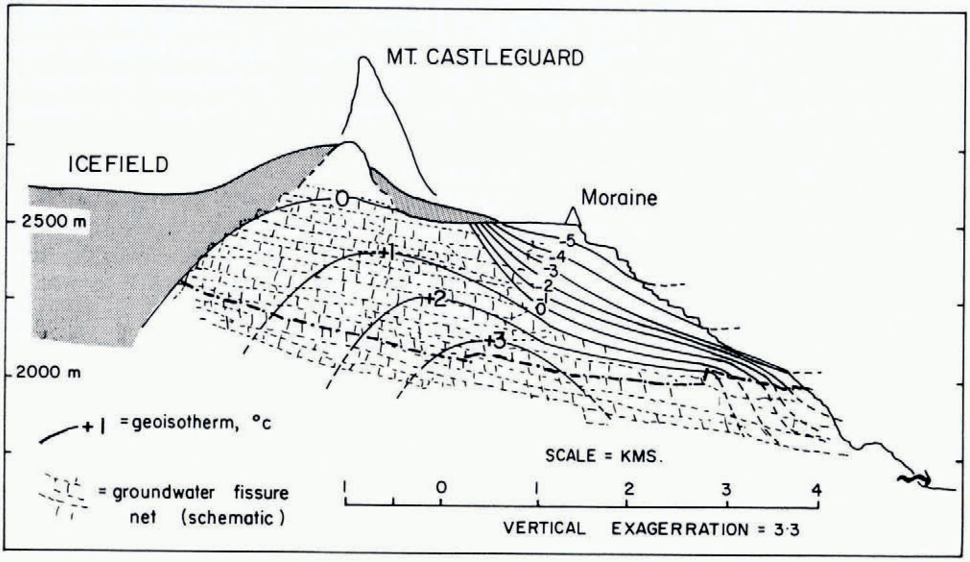

Fig. 5. Geo-isothermal section of Mount Castleguard and the cave, inferred from the data of Figure 2.

As a consequence, it is significant that the geothermal gradient between the cave and the glacier base is much gentler, 0.004 deg m^{-1}; geothermal heat flow towards the glaciers appears to be retarded. It is suggested that this is caused by melt waters (invasion waters) abstracting heat as they pass down through small fissures. In Figure 5, hypothetical flow paths of invasion waters are drawn to accord to the observed geological and hydrogeological conditions in the environs of the cave. There is an inverse correspondence between the frequency and distribution of fissures and the geothermal gradient.

Abstraction of geothermal heat is independently illustrated by the case of the Big Springs. At least 90% of their discharge is melt water from glacier soles, with an initial temperature of 0°C. Allowing for mixing of the $\leqslant$10% of warmer, extraglacial waters it is estimated that Big Springs waters measured in August 1969 had gained 1.5 deg from geothermal sources.

* This particular measurement is a mean of six measurements at the site over a 40 h period in 1973. 1974 measurements reproduced this value.

From the Castleguard examples, it is suggested that geothermal flow to glaciers that rest upon fissured rocks such as many limestones, dolomites, sandstones, etc., may be reduced to 10–40% of the expected flux. This result is consistent with a rough calculation of the effect of seepage water upon the heat balance in the rock. We are specialists in speleology and would emphasize that the frequency of fissuration in the Castleguard limestone is *lower* than in almost any other cavernous rock known to us.

3. Chronologic and palaeo-temperature studies of speleothems

From Figure 5, a lowering of the mean annual external air temperature and/or the basal ice temperature by 4 deg is required to place all or almost all of the cave below 0°C. This would halt the flow of most ground waters except possibly the largest, which in turn would halt the deposition of calcite speleothems (stalactites and stalagmites) from seepage waters. It is widely accepted that depression of temperature during maxima of the Last Glaciation greatly exceeded 4 deg in extra-tropical areas. The speleothem record of Castleguard Cave is therefore of interest.

North-west of P80 (the limit of seasonal thermal effects) the cave is plentifully decorated with speleothems. They grow in both vadose and phreatic sections and may be divided into two classes: (a) speleothems, tending to be large in dimension, that grew before a flooding event that effected considerable re-solution of the calcite; (b) speleothems, tending to be small, that have grown since the flooding event and display no re-solution: most appear to be actively growing today.

Calcite speleothems may be dated by Uranium series methods if they were deposited with traces of uranium in sufficient amount but no thorium (Thompson and others, 1974). Castleguard speleothems are satisfactory in this respect and the sample record of their ages is presented in Figure 6. When collecting in the field there is no way of determining that the complete age-range of growth has been sampled (or of determining whether precipitation occurred in conditions of equilibrium or kinetic fractionation: see below). It will be appreciated that the age record of speleothem growth in the cave is therefore necessarily incomplete.

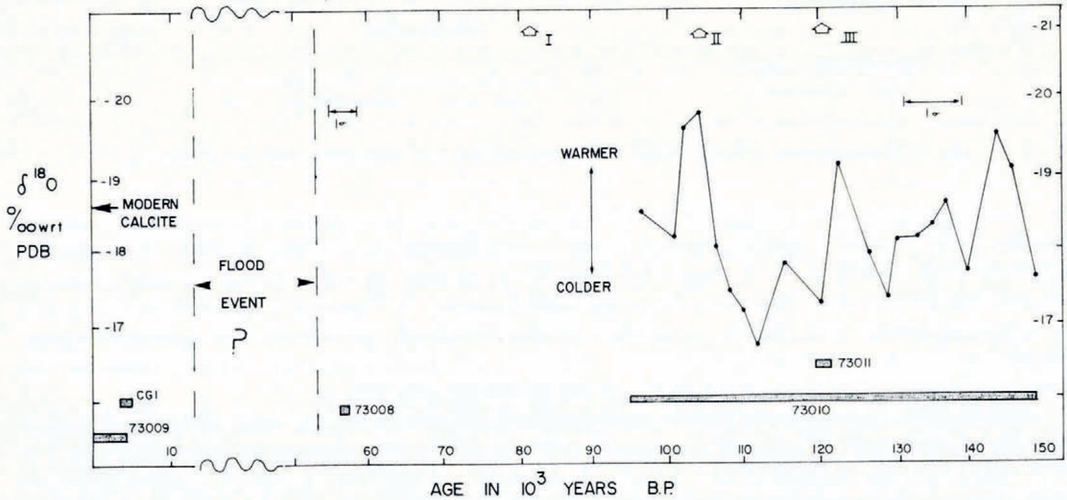

Fig. 6. Uranium series chronology and $^{16}O:{}^{18}O$ fractionation of speleothem specimens from Castleguard Cave. 1σ = error margin of dating (1 standard deviation). I, II, III = temporal position of the Barbados high sea-levels (Broecker and Van Donk 1971).

Ages of speleothems growing before the flood event range from 155 ± 9 ka to 56 ± 2 ka. The oldest post-flood specimen is 4 ± 0.2 ka in age. Growth has been generally continuous since that date and probably commenced somewhat earlier in other speleothems that were not sampled.

The most important specimen in 73 010. This was a sample from the stalagmite shown in Figure 7. The stalagmite is two metres in height, standing in the centre of the principal passage at a place where the cave cross-section was reduced by earlier sedimentation. 73 010 was taken across growth layers at the base of the stalagmite. The entire feature has suffered severe re-solution by flowing water which filled the passage and preferentially removed the outer (younger) growth layers. The stalagmite therefore grew for some time after the youngest date obtained from it, 93 ± 3 ka. From a reconstruction of its dimensions before re-solution it is reasonable to suppose that speleothem growth was continuous until 56 ± 2 ka, the time of the youngest dated pre-flood material (73 008).

Fig. 7. Eroded stalagmite standing in the centre of the principal passage 800 m south-east of P200. The feature is 2 m in height. It has suffered re-solution by waters of the long-sustained flood event (see text). It is partly buried by fines deposited by the floods and decorated with some subsequent calcite precipitate, including the "false floor" seen at bottom left. Specimen 73 010 was taken from the base of this stalagmite. (Photograph by D. C. Ford.)

U/Th ages therefore demonstrate continuous speleothem growth from *c.* 155–93 ka, and probably until <56 ka when it was halted by a flood event. It recommenced before 4 ka.

Hendy (1971) has shown that oxygen isotopes in cave calcite are fractionated in one of two ways: either under conditions of equilibrium resulting from the slow outgassing of CO_2 from aqueous solution or kinetically if CO_2 outgassing is rapid and/or if evaporation occurs.

Under conditions of equilibrium deposition, the oxygen isotope variations in a speleothem are an indicator of palaeoclimate change through the effect of temperature change on the calcite–water fractionation, as in oceanic foraminifera (Emiliani, 1955; Broecker and Ku, 1969), and the Greenland and Antarctic ice cores (Dansgaard and others, 1969).

Three speleothems have been analysed for C and O isotopic variations. Speleothems younger than the flood event proved to be kinetically fractionated. This implies that the modern pattern of summer and winter air flow in the cave has existed for much of Holocene time. Very young, fast-growing straw stalactites in recesses where the drafts are not effective and measured relative humidity is >95% yield a $\delta^{18}O$ value of $-18.6‰$ PDB, compatible with equilibrium fractionation at approximately +3°C.

Specimen 73 010 displays equilibrium fractionation; the record is included in Figure 6. It is considered that the $\delta^{18}O$ variation probably represents a variation of temperature of approximately 5 deg, that is 2 deg warmer to 3 deg cooler than the present value of +3.5°C at the site. An increase of 2 deg in the mean annual external air temperature would not suffice alone to destroy the Columbia Icefield. Probably this persisted at approximately its modern dimensions during the period 155–93 ka. Three warm peaks and one cold trough occur in the 73 010 record. The peaks correlate well with $\delta^{18}O$ peaks that we have measured in specimens which grew at the same time in caves in Bermuda and Kentucky. There is also good agreement with the two older of the three raised coral reefs in Barbados taken to represent climatic optima during the Last Interglacial (Broecker and Van Donk, 1970). This suggests that the deep cave speleothems of Castleguard Cave (and therefore the hydrogeothermal state there at a given time) record climatic events of greater than continental scale.

The speleothem chronological and palaeotemperature record so far obtained is discontinuous and is, perhaps, over-interpreted here. However, the trend of findings is to suggest that at no time during the past 155 000 years have supplies of seepage water to the cave been discontinued by freezing at the glacier base and in the cave rock. If mean annual external air temperatures declined more than 4 deg below the calculated values (as would seem probable at peaks of Wisconsinan glaciation), the rock was shielded by contemporaneous expansion of glacier ice temperate at the base.

4. The flood event

At some time after 56±2 ka (specimen 73 008) the cave was filled entirely with water. This is demonstrated by the re-solution of speleothems (reported above) and by the deposition of up to 15 m of laminated silts and clays in the passages. These deposits have suffered much subsequent erosion by the freshest waters but sufficient are preserved to indicate that deposition was simultaneous throughout the great length and vertical extent of the cave. Rhythmic colour banding is common in the deposits but is too variable for them to be categorized as varves. However, their nature indicates settlement from a semi-continuous supply of sluggishly circulating water that persisted for a long period.

The underlying cave, Castleguard II, will also have been flooded. Therefore, a head of water of at least 600 m was contained within Mount Castleguard. The agent impounding such a large body of water in well-drained, cavernous rock whilst permitting its continual re-supply would appear to be a valley glacier infilling the Castleguard River valley. From the U/Th age dating this glacier was, broadly, "classical" Wisconsinan in age. The caves were incorporated into the glacio-hydrologic system of the Icefield and its valley-glacier offshoot.

The glaciers circulated water not merely at the base; from considerations of hydraulic gradient in the rock, the permanent glacier water table must have been some hundreds of metres above the base of the ice in the Castleguard valley where trim-lines and other evidence suggest a Wisconsinan glacier that was 750–800 m in depth.

There is no evidence to suggest that the water-fill froze at any time within the cave. We have seen the effects of such freezing in other caves situated above Wisconsinan glacial trim-lines in the Canadian Rockies. Wall rock and speleothems display very dense patterns of shatter. These are absent in Castleguard Cave.

Conclusions

Most parts of the Mount Castleguard glaciers and a substantial portion of the Columbia Icefield belong to a hydro-geological system discharged through Castleguard Cave and the Big Springs. The modern ground-water hydrology suggests that surface glacial melt water penetrates to the bases of these glaciers during the melt season, both above and below the firn line. Winter discharge from the glacier soles is reduced to *c.* 1% of the summer volumes. The geothermal flux to the base of the ice is significantly lower than expected because the heat is abstracted by melt waters passing through the fissured rock. These conditions may have persisted for most of the past 155 000 years. During some part of the last glaciation ("classical" Wisconsinan) when the ice mass was greatly expanded, the glacial water table was some hundreds of metres above the sole in a valley glacier draining the Icefield; the caves were flooded as a consequence.

Acknowledgements

These researches were supported by grants from the National Research Council and the Department of Energy, Mines and Resources, Canada. Dr M. C. Brown, University of Alberta, kindly supplied data on wind-flow and temperatures at the entrance of Castleguard Cave. Dr J. J. Drake, McGill University, contributed to the many discussions of hydrology in the field.

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DISCUSSION

W. D. Harrison: Where is the air circulating through the cave going?

D. C. Ford: It passes into an inaccessible fissure at P100, 800 m south-east of the termination of the cave at the glacier blockage. From there is may either (a) discharge into the base of the glacier through open sink holes or (b) follow some unknown cave passage beneath the glacier. On geological grounds, this would trend north-west to outlets to open air in limestone benches north of the head of the Columbia Glacier, i.e. it would pass entirely across the ice field. The second alternative seems improbable. However, there are difficulties with the first, also. Melt waters discharging at the Big Springs are saturated with respect to calcite at concentrations only one-third as high as those expected. This implies a chemical evolution of the waters in a system closed to the addition of CO_2 from any subglacial atmosphere which, in turn, suggests very little void space at the glacier sole into which air may be discharged in winter or from which it may be drawn in the summer. Winter discharge rates for the cave air are 36 to 50 $m^3 s^{-1}$. Summer withdrawals will be similar in amount.

T. J. Hughes: Is it possible to date at least the duration of the flood event from sedimentation rates in the clay deposits?

Ford: It has not been attempted as yet. There are difficulties. Modern invasion waters have removed the basal areas of many sections; others are inaccessible because the infilled fissures are too narrow at the base to be entered.

B. Hallet: Do you have any data on the composition of the ice that you believe to be extruded into the cave from the base of the ice field?

Ford: We have been able to recover a sample of some 40 ml as water. It will be difficult to extract any sort of core as a solid. The water sample has been analysed for $\delta^{18}O/^{16}O$ and proves to be identical in this respect to the melt water of the high, small glaciers surrounding the summit of Mount Castleguard and also to the water emerging at the Big Springs.