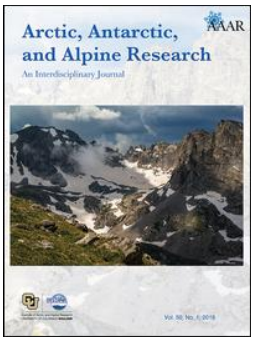

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C. C. Smart

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THE HYDROLOGY OF THE CASTLEGUARD KARST, COLUMBIA ICEFIELDS, ALBERTA, CANADA*

C. C. SMART†
Department of Geography, McMaster University, Hamilton, Ontario L8S 4K1, Canada

ABSTRACT

The Castleguard area is a classical alpine karst, characterized by high relative relief and partly overlain by glacier ice. Access to the subglacial aquifer is possible via a relict cave. Drainage of the glacier bed occurs at both the conduit and diffuse flow scales. The contemporary hydrology is dominated by supraglacial meltwaters, which have been traced into the groundwater system. More than 100 springs in the Castleguard Valley constitute a hierarchy based on size and relative elevation. Status within the hierarchy is demonstrated by behavior as "underflow" or "overflow" springs. Floods emerging from springs perched 300 m above the valley floor are an overflow probably derived from the Saskatchewan Glacier. Quantitative dye tracing has revealed complex spring behavior, largely in response to changes in the spatial distribution of recharge. The aquifer appears to be mature, with a well-developed conduit network. The numerous springs are a product of disruption by glaciation; disturbance appears to be the major effect of glaciers upon this karst aquifer. Ice may be drained karstically where sufficient pressure gradients exist; decreased ice velocities and erosion rates are possible effects of underlying karst upon glaciers.

INTRODUCTION

This paper examines the present-day hydrology of the glacierized alpine karst around Castleguard Mountain and uses the observations to consider the influence of glaciation on karst hydrology.

Karst systems are characterized by the loss of surface water underground where it is transmitted through discrete conduits (or caves) to springs. Water gradually dissolves the host bedrock, and consequently the aquifer evolves in character through time. However, the rate of evolution is slow compared to the characteristic frequency of hydrological events; the form of these events may therefore be considered fixed on the scale of a few years.

In alpine regions, karst processes are enhanced by the high relief and steep hydraulic gradients, although competing periglacial processes may militate against classical development. Snow and ice are also present and glaciers often dominate geomorphic and hydrological processes.

Glaciers are complex in their hydrological behavior. Although meltwater is mostly produced at the ice surface, a large part may be routed through subglacial tunnels to the snout. In some respects, therefore, the glacier is a karstic aquifer. However, the glacier conduits close in winter due to plastic deformation and, therefore, the characteristic life span of the glacial aquifer is far less than that of a conventional karst aquifer.

Glaciers have had considerable impact on many karst areas during the Quaternary (Ford, 1983a). Although the geomorphic impact of glaciation may be interpreted from

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†Present address: Department of Geography, University of Saskatchewan, Saskatoon, Saskatchewan S7N 0W0, Canada.

relict landforms, the nature of hydrological interaction remains somewhat speculative. At Castleguard, the glacier overlies an ancient, but active karst aquifer, providing a unique opportunity to study glacier-karst interactions.

Much or all of the interior of both karst and glacial aquifers is inaccessible, but their response to natural or artificial stimuli may be studied. Such a "black box" approach has proven successful (e.g., Brown, 1972) but is best applied to simple systems. At Castleguard, the cave allows direct access to part of the interior of the karst aquifer, providing useful qualifying data.

Both the glacial and karst components must be considered in a study of the modern hydrology of the area. This was attained through quantitative discharge measurements and dye tracing, and qualified by observations made within Castleguard Cave and on recently exposed Neoglacial surfaces.

HYDROLOGICAL COMPONENTS

Aquifer Units

There are four aquifer elements at Castleguard: the Cathedral Formation, the Eldon and Pika formations, unconsolidated deposits, and glacier ice.

The major karst aquifer is the Cathedral limestone which contains both Castleguard Cave and the presently active (inaccessible) systems, Castleguard II and III (see Ford, 1983b, this symposium). The limestone is strongly karstified where it is exposed, for instance in the northern Meadows area. Karstification is concentrated on the widely separated fractures that characterize the limestone. This accounts for the well-formed nature of shafts and

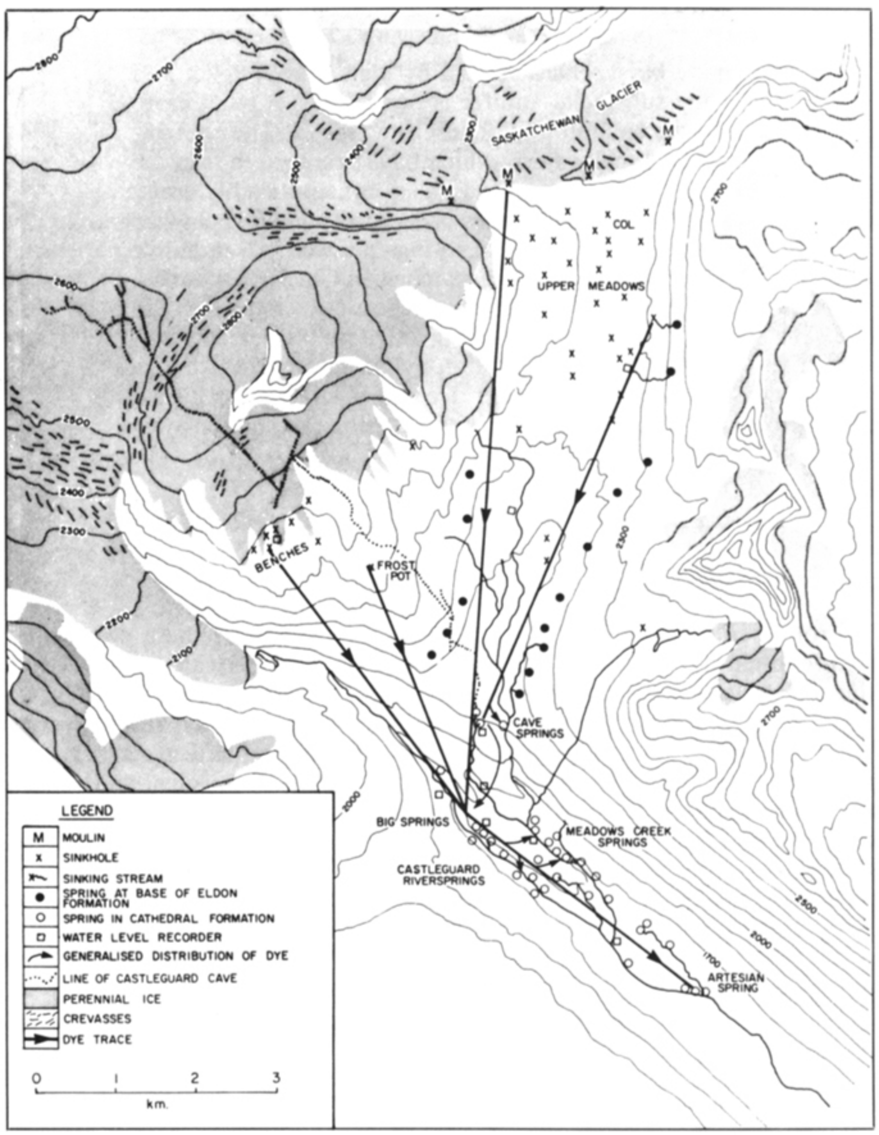

FIGURE 1. Hydrogeology of the Castleguard Karst. Flow recorders referred to in the text are numbered thus: 1—Castleguard River; 2—Meadows Creek (1); 3—Big Spring; 4—Cave Stream.

conduits explored in the Cathedral Formation. Stratigraphically above it lies the shaly Stephen Formation, which is generally impermeable. In places such as the South Benches, however, fractures provide localized high permeability. The Eldon and Pika formations constitute an upper aquifer which shows considerable variety in hydrologic behavior. Limestone components are readily soluble, but the frequent dolomite laminae less so. Numerous local-scale aquifers exist, discharging at the contact with less soluble beds. Well-formed shafts have developed only on major fractures and on sedimentary dikes found on the South Benches. These shafts are not enterable, but appear to provide hydraulic continuity through the Stephen shale into the underlying Cathedral limestone. Along the margins of Castleguard Meadows, in contrast, the Eldon-Stephen contact is marked by many springs (see Figure 1).

Groundwater is also present in widespread unconsolidated deposits in the area. Neoglacial moraines lie within a few hundred meters of present ice margins, while older moraine bodies occur in the central Meadows and in the floor of the Castleguard Valley. The hydrological role of these features depends on their textural character. Usually, they are somewhat impermeable and function as aquicludes. The southern Meadows contain several extensive gravel fans which readily absorb water. Springs occur at their lower margins.

Temperate glacier ice is also an aquifer. The innate permeability of ice is very low (Raymond and Harrison, 1975), but surface meltwater may reach the glacier bed through crevasses and moulins. In spring, a buildup of basal water pressure leads to the development of integrated conduit networks leading to the snout (Röthlisberger, 1972; Shreve, 1972). Although developing under high pressure, once the network is fully formed it may fall to atmospheric pressure (Engelhardt, 1978; Hodge, 1979). As discharge falls at the end of the ablation season the conduits are closed by plastic deformation and must be reformed the following year. Reforming is generally governed by the basal pressure field but, more locally, bedrock irregularities control the paths of conduits (Nye 1973).

Water also occurs at the glacier bed in the form of a high pressure regelation water film (Weertman, 1972; Hallet, 1979a, 1979b). This is very poorly connected to the conduit network, although local flow takes place within the film under the influence of pressure fluctuation and phase changes (e.g., Robin, 1976).

Hydrologic continuity between glacier ice and groundwater has been inferred in several cases (Ford, 1971; Maire, 1977; Wildberger, 1981). Studies of deglaciated surfaces have shown that limestone karst develops beneath glacier ice (Ford, 1979; Walder and Hallet, 1979) and in some cases could only have functioned beneath glacier ice (Figure 2; Lauritzen, 1981).

The beds of many glaciers consist of a layer of "subsole drift" (Englehardt, 1978; Hodge, 1979). The hydrologic effect of drift depends on its texture and the hydraulic gradient applied. In many cases it appears to be quite impermeable (Hodge, 1979).

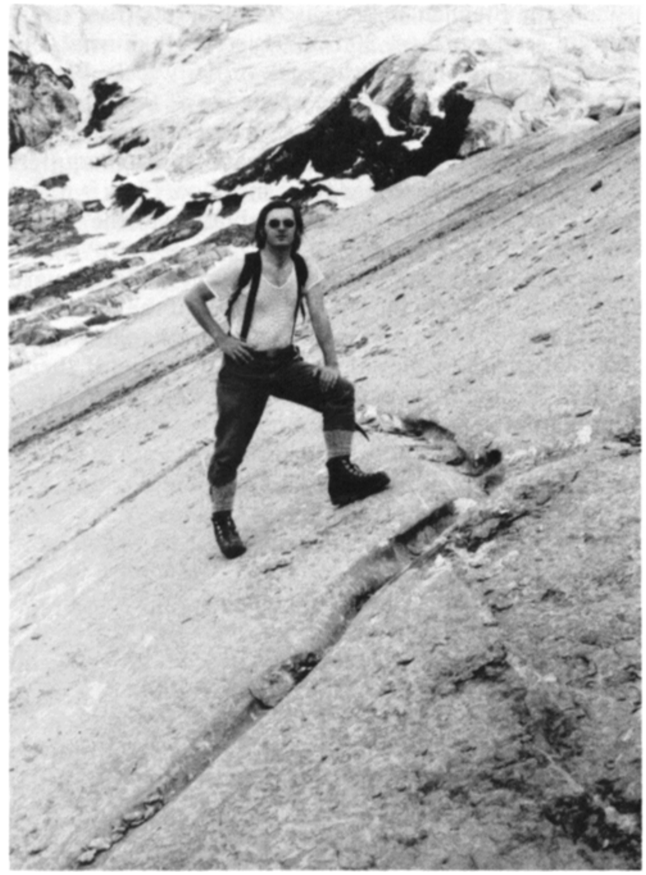

FIGURE 2. A Nye channel cut in bedrock on a recently deglacierized surface near the Castleguard area. Such a channel can only function beneath glacier ice.

It is concluded that, in the Castleguard area, karst subglacial drainage is likely where ice is in contact with the Eldon-Pika and Cathedral formations. Whether such drainage develops depends upon the hydraulic gradients available, and the local distribution and quality of subsole drift.

SPRINGS: DISCHARGE FROM THE AQUIFER

There are more than 100 springs in the Castleguard area (Figure 1). Their duration of flow ranges from a few perennial examples through seasonal springs to transient snowmelt and intermittent springs. A hydrogeological division into four groups may be made:

(1) Springs draining the Eldon Formation.
(2) Cathedral springs at the elevation of Castleguard Cave.
(3) Cathedral springs in the floor of Castleguard Valley.
(4) Springs draining unconsolidated aquifers.

Eldon springs are located on underlying aquicludes, usually the Stephen shale. Discharge varies from insignificant seepage to ca. 20 L s^{-1}, is strongly diurnal, but markedly lagged behind the peak of daily melt. Flow is short lived (<2 mon) on the west edge of the Meadows where recharge is from ephemeral snow patches. Springs along the eastern margin of the Meadows drain aquifers fed by glaciers and seepage from moraine. Flow is maintained during the 3 to 4 mon of ablation at the source altitude.

The Cathedral springs at the elevation of Castleguard Cave discharge from the Upper Member of the formation (Ford, 1983b, this symposium), and are hanging some 300 m above the valley floor. Flow from one of them, the Red Spring, is perennial; it is weakly diurnal during the ablation season, and responds to rain events. It is shown in Figure 3. Other springs including the Cave Mouth, are normally dry, but discharge major floods (up to 6 m^3 s^{-1}) during hot weather in summer. The nature and origin of these transient floods is considered in more detail later.

The Castleguard Valley floor contains some 100 springs with an aggregate discharge of up to 20 m^3 s^{-1}. They are concentrated along two axes, the Castleguard River and the lower course of Meadows Creek (Figure 1). Many emerge from exposed bedrock in hanging positions, while others discharge through till or fluvioglacial gravels. A few emerge in anomalous positions along the divide between the two streams.

Artesian Spring is the lowest known spring. It is perennial and discharges ca. 1 m^3 s^{-1} in summer. Farther up the valley, springs are no longer artesian and show progressively shorter flow durations. Visually dominant are Big Springs, a group of four springs perched 15 to 40 m above the Castleguard River, 3 km upstream of Artesian Spring. The largest of this group (Figure 4) had peak flows of ca. 5 m^3 s^{-1} in 1979 and 1980. The highest active springs noted in those years were approximately 110 m above the elevation of Artesian Spring. Above them were additional spring sites with channels clear of vegetation (suggesting recent activity) but no water and, finally, overgrown spring points that appear completely inactive.

Water quality is similar throughout the Valley springs: low suspended load, low dissolved load, and depleted PCO_2. This and their diurnal variations in flow suggest a supraglacial origin. Under low flow conditions, turbidity of the springs increases, suggesting a proglacial source normally masked by clear supraglacial water at high flow.

Springs from unconsolidated materials are quantitatively insignificant, but may provide perennial flow into the more rapidly depleted karst aquifer.

Sinks: Recharge of the Aquifer

The characteristics of aquifer recharge depend upon meteorological events and the hydrology and topography of sink points.

Parks Canada (1981) estimate >7 m annual snowfall on the Columbia Icefield, and at least 5 m falls on the lower Meadows area. Snowpack ripening occurs between April and August, depending on elevation. Melt is usually very rapid once a snowpack is ripened, except in cases of thick drifting and a northerly aspect.

Rainfall in mid-July to mid-September 1979 was 104 to 144 mm. In the same period in 1980, 140 to 250 mm fell. These figures reflect values from two strongly contrasting years. They also illustrate the order of spatial variability characteristic of mountain precipitation. The

FIGURE 3. The Red Spring: An immature spring just below the Cave Entrance in the limestone of the Cathedral Formation.

South Benches were consistently the wettest area, and the Upper Meadows the driest. Precipitation ranged from hard storms (25 mm in a few hours) to gentle drizzle and snowfall at higher elevations throughout the summer. By mid-September 1980 mean daily temperature was less than 0°C on the Benches. Over 60 cm of new snow had accumulated there, much of which was not measured by rain gauge. This exemplifies the difficulty of identifying the hydrologic behavior of source areas which cover a wide altitudinal range.

The sinkholes most favored by snowmelt and rainfall are those which have large catchment areas. Closed depressions are well developed in the Upper Meadows, where all snowmelt and precipitation is channelled underground through numerous shafts and fissures. Farther south, morainic material and alluvium have plugged shafts so that the capacity to absorb water is limited, and runoff from intense storms overflows to Meadows Creek. On the South Benches, an abundance of shafts absorbs all surface water, and only in Neoglacial moraine areas

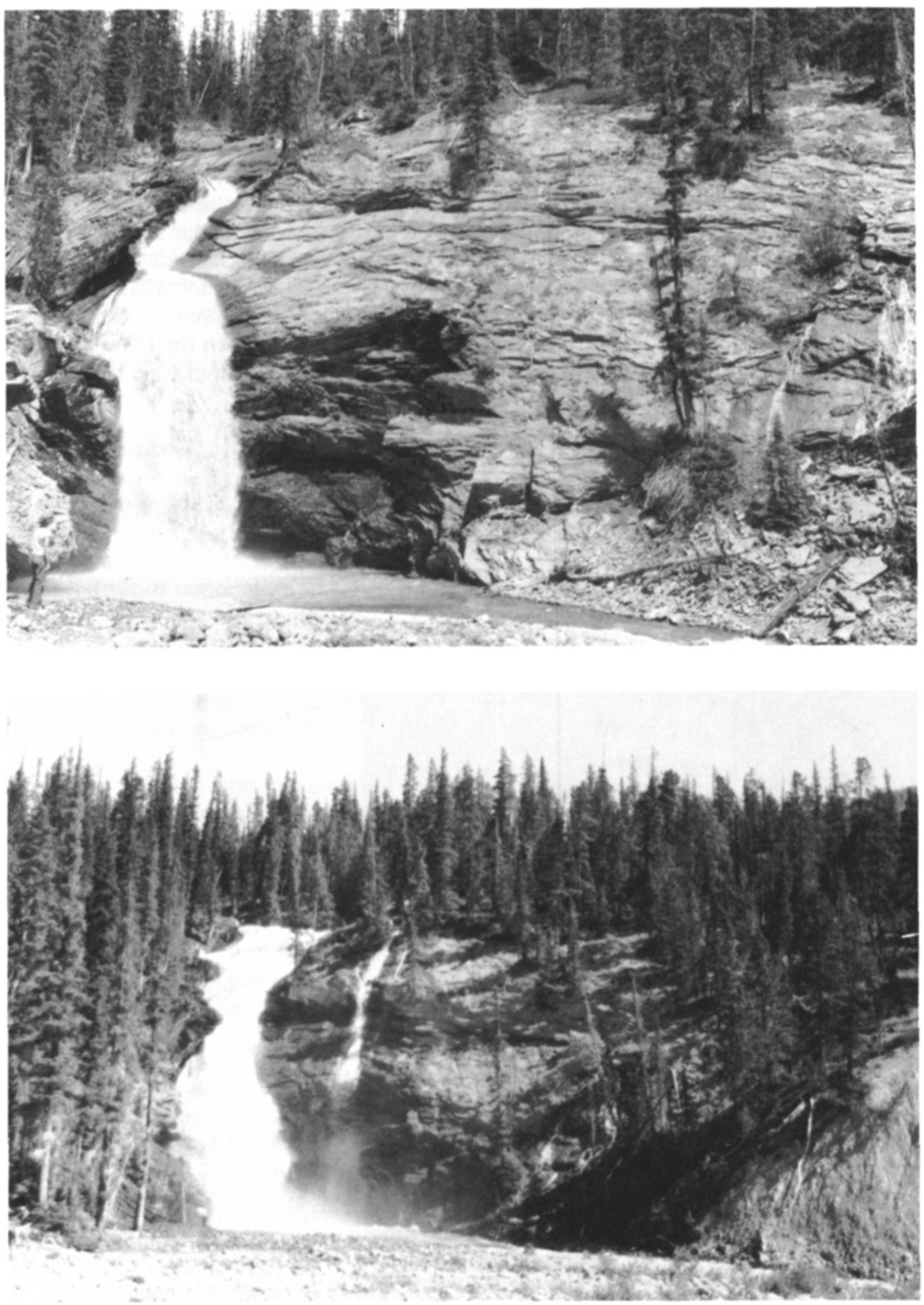

FIGURE 4. Big Spring is the visually dominant member of the Spring hierarchy of Castleguard Valley, but flows for only 3 to 4 mon per year. Discharge changes rapidly; these pictures were taken three days apart.

does surface runoff occur. In one actively eroding shaft, 30 m from the ice front, a calcite flowstone was found. A sample of this material has been dated using the U/Th method at 6100 ± 200 BP (sample no. 80017).

The topmost springs along the Meadows Creek axis in the Castleguard Valley discharge only occasionally. At other times water may sink into the aquifer via these spring points: they are reversing features called "estavelles." The Castleguard River also flows across Cathedral limestone for a considerable distance upstream of known springs. There may be some aquifer recharge here, although the armored bed and high sediment load would hinder this. At low flows the spring water becomes more turbid, but is still markedly clearer than the Castleguard River. The Red Spring water all sinks underground again at low flow (ca. 50 L s^{-1}).

After spring melt many recharge areas are active only during rainfall events. The bulk of observed recharge is then at proglacial sinkholes where discharge is maintained by melting of perennial ice or snow, e.g., at the margins of the glaciers around Castleguard Mountain (Figures 1 and 5). The only other sustained recharge is a few sinks in the Col karst which are fed by springs from the Eldon-Stephen contact. Total *observed* recharge in mid-summer is estimated at 2 m^3 s^{-1}.

Despite the dramatic decrease in the apparently active catchment area after spring snowmelt, the springs in Castleguard Valley attain their full flow of some 20 m^3 s^{-1} in mid-summer. The remaining 18 m^3 s^{-1} of discharge can only be accounted for by recharge of meltwaters at the bed of the Columbia Icefield and channel losses in local streams.

Water quality in valley springs and the diurnal flow regime suggest a dominantly supraglacial origin. Moulins and crevasses are sites where surface meltwaters access the glacier bed. These are shown in Figure 1. The potential source area runs in an arc from the upper South Glacier around the northern flank of Castleguard Mountain and down the Saskatchewan Glacier. Castleguard Mountain stands on a broad pedestal of the Eldon-Pika rocks. Crevasses are concentrated on the northern rim, forming a marked icefall 2 km north of Castleguard Mountain. The geological structure (see Ford, 1983b, this symposium) is such that the waters descending these crevasses will readily access the Cathedral limestone. The unentered hydrological link conjectured between here and the Castleguard Valley is titled Castleguard II.

North of Castleguard Meadows, the Saskatchewan Glacier is in direct contact with the Cathedral limestone. There are numerous crevasses and moulins, although no direct recharge to bedrock is observed at the glacier margin. The hypothetical link between the Upper Meadows karst and the Cave has been called Castleguard III (Ford, 1971).

THE CAVE: INTERIOR OF THE CATHEDRAL AQUIFER

Castleguard Cave is an ancient, long-abandoned trunk conduit which drained the upper member of the Cathedral Formation. Today it is hydrologically active only around the entrance and as a route for small scale vadose waters. Nevertheless, it demonstrates the types of conduit developed in the Cathedral limestone and their style of organization.

The conduit is large and unbranched with a cross-

FIGURE 5. A typical proglacial stream sink on a fracture in the Pika Formation on the South Benches. The massive nature of the limestone and the immaturity of the sink is apparent. Castleguard Mountain, composed of younger clastics, rises in the background.

sectional area of 4 to 9 m^2. It runs essentially downdip with occasional tributaries (see Ford et al., 1983, this symposium). The cave has long been abandoned (>700 ka; Gascoyne et al., 1983, this symposium), although subsequent gentle flooding by glaciogenic waters has introduced extensive rhythmic clay-silt deposits into the cave (Schroeder and Ford, 1983, this symposium). The headward complex passages are blocked by debris, some of which appears to be glacial till; glacier ice blocks one conduit.

Vadose invasion waters are active in much of the Cave and dissect sediments and bedrock features before descending to Castleguard II down inaccessible shafts. In the central cave invasion waters are concentrated on faults or major joints, occasionally marked by ancient speleothems. In the headward complex there are numerous youthful shafts which bear little relation to the cave, often driving directly through it to lower levels. Erratic pebbles are seen in occasional channel segments fed by them.

During winter expeditions in April, there is little hydrological activity in the cave. However, the reports of summer explorers (e.g., Thompson, 1976) and high water marks indicate greater hydrological activity in summer.

The strong flooding at the cave entrance is caused by water flowing up a shaft at the head of Boon's Blunder. The provenance of this water is unknown. Floods occur only during sustained warm weather. Flood waters emerge initially from springs some 20 m lower than the entrance. Further data on the cave floods are presented in the quantitative hydrology section of this paper.

A characteristic feature of the cave is the persistent draught which flows in at the entrance in winter, and out in summer (see Atkinson et al., 1983, this symposium). The winter destination of this air appears to be the shafts in the headward complex.

Coexisting with the actively eroding shafts at the upper end of the cave are fresh calcite speleothems. There is clearly a marked contrast in the chemical and hydrological character of the waters forming these adjacent features.

Small caves are also found within the Neoglacial limits of the South Glacier. They appear to have been active as "marginal channels" running parallel to the valley wall with occasional entrances, some blocked by glacial debris. Their anomalous position suggests that they must have developed relatively rapidly during glaciations, perhaps exploiting preexisting karst openings. Similar forms are possibly active today along the present margins of the South and Saskatchewan glaciers.

THE CATCHMENT

In karst areas the hydrologic boundaries do not necessarily coincide with surface divides. The first task of the karst hydrologist is to define the catchment of particular springs and the destination of influent streams. This is done using various techniques: point-to-point dye tracing, analysis of regional water table trends, regional geology, etc. (e.g., Quinlan, 1982).

At Castleguard it is not possible to define a simple regional water body. Furthermore, glacier ice obscures much of the catchment area. An attempt to define the catchment area was made by:

(1) Dye tracing from accessible sink points.
(2) Investigating subglacial bedrock topography.
(3) Comparing runoff with an instrumented glacier.

Point-to-point dye traces were made from the Benches, Frost Pot, and the Upper Meadows (Figure 1). A trace was also made from a moulin on the Saskatchewan Glacier. All traces were positive to the Big Spring and every other spring sampled in the Valley groups.

The trace from the Upper Meadows emerged strongly in the springs around the cave, appearing less strongly at the Big Spring. The trace from the Saskatchewan Glacier did not appear at any of the Cave Springs; the cave was not in flood at the time.

The catchment thus defined is a broad arc some 4 to 6 km in radius with an apex of ca. 45° at the Big Spring. The nearer portion of the arc is drained surficially by the Meadows Creek. There is no confirmation of the extent of the glacial component of the catchment.

Figure 6 shows the estimated subglacial bedrock topography of the area. It was constructed using data from radio-echo sounding (Waddington and Jones, 1977) and seismic profiling (Meier, 1960). The surveyed elevation of the ice blockage in the cave (2305 ± 10 m) provides a further datum, and the entire headward complex a minimum elevation. It is assumed that crevasse lines represent major bedrock steps and that there is concordance between ice and bedrock divides.

It can be seen that the subglacial topography is typical of alpine glaciation. The headward complex of the cave underlies a cirque-like feature, possibly a closed depression. Part of the bed of the Saskatchewan Glacier trough is a closed depression. This is higher than the Big Springs (1740 m), although the gradient between the two points (ca. 0.019) is much less than that from the headward complex (0.058). Direct contribution from areas north and west of the cirque heads is likely to be limited by the high elevation and absence of crevasses.

Peyto Glacier 70 km south of Castleguard is part of an International Hydrological Decade (IHD) study basin (Young, 1977). The glacier area of 13.9 km^2 and representative summer daily ablation discharge of 8×10^5 $m^3\ d^{-1}$ yield a specific runoff of 57.6 mm d^{-1}. Comparable figures of 60 to 67 mm d^{-1} are obtained from a small glacier in the Purcell Mountains, 180 km southwest (Weirich, pers. comm., 1982). In contrast, the specific runoff of the topographic catchments of the South Glacier and Meadows Creek is only 13 and 23 mm d^{-1}, respectively. This suggests either significantly reduced ablation or substantial loss to groundwater.

An estimate of the total discharge from the karst aquifer is 20 $m^3\ s^{-1}$ at high stage. Taking Peyto Glacier as a standard with its specific runoff of 57 mm d^{-1}, the catchment area of the valley springs is estimated to be 30 km^2. This corresponds to the area of all glaciers around Castleguard Mountain, the upper South Glacier, and the Saskatchewan Glacier above the Meadows. It is probably an underestimate, because the area is largely above the firn line and ablation is correspondingly reduced when compared to an entire glacier. Here, it is assumed that there is no contribution from Meadows Creek and Castleguard River.

If the catchment area is taken as 30 km^2, and the range of dissolved load taken at extremes of 20 and 30 mg L^{-1}, an approximate erosion rate may be calculated. Total measured spring flow is calculated at 2.3×10^7 m^3 for 1980. Given that only about half the springs are included in this figure, that runoff during the spring was not measured, and that 1980 had a very cool summer, a figure of 4×10^7 $m^3\ yr^{-1}$ is probably more reasonable. Taking these extreme values, erosion rates of between 6 and 15 mm ka^{-1} are calculated. These are far lower than glacial erosion rates of 82 to 1400 mm ka^{-1} given in Embleton and King (1975: 313), but are only for solutional erosion of carbonate. Nevertheless, the clarity of the spring water supports this low erosion rate for the spring catchment area.

QUANTITATIVE HYDROLOGY

Measurements of physical hydrology allow an understanding of the behavior of the aquifer. The models generated may be evaluated by quantitative dye tracing.

Hydrologic Measurements

The inaccessible karst aquifer may be considered to be a black box in which spring discharge is a response to impulses applied to the system. It is therefore necessary that the inputs and outputs be documented in order to characterize the aquifer.

In alpine regions runoff is generated by both snowmelt and precipitation. Neither process is simple. Melt is governed by surface energy balance and snow and ice distribution. Precipitation is spatially and altitudinally variable in both quantity and form. Given that the catchment area of the karst aquifer is undefined and that ice topog-

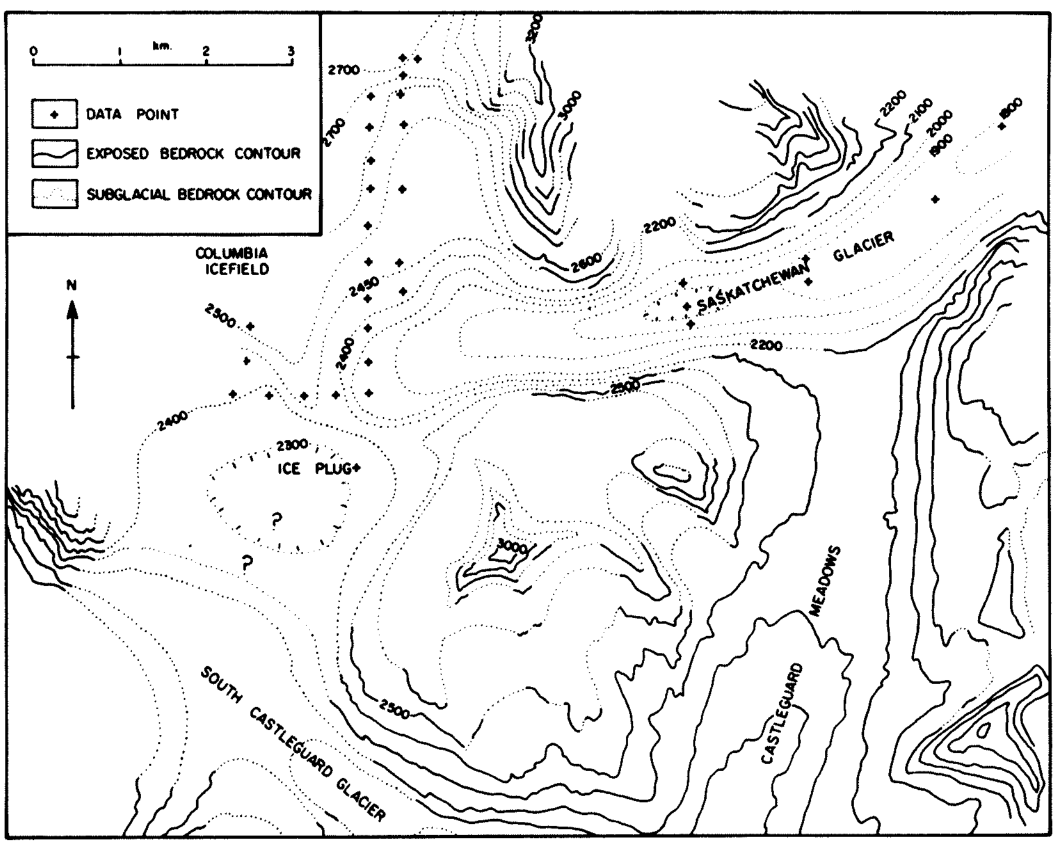

Figure 6. Bedrock topography of the Castleguard area. Although completely mantled by glacier ice, the underlying morphology is dominated by cirque forms. Note the closed depression in the Saskatchewan Glacier. It is probable that another overlies the headward complex of the cave.

raphy does not define hydrological divides, a hydrologic budget approach is not suitable. Some form of analog input is needed.

It is possible to simulate watershed runoff using a suitable watershed model (e.g., Quick and Pipes, 1977). This has been done for Peyto Glacier with some success (Power and Young, 1979). However, the time resolution is only to mean daily discharge which is insufficient for the present purpose. Instead, two glacial meltwater streams were continuously gauged to provide a crude analog input function. One is fed by the East Castleguard Glacier on the northeast flank of Castleguard Mountain. The other is the Castleguard River flowing from the South Glacier (Figure 1).

It can be seen from the 1980 hydrograph (Figure 7) that the input series is complex and nonstationary. It is possible to identify three constituent elements: (1) a seasonal low frequency component reflecting hydrological storage and persistence in the daily temperature; (2) a diurnal component reflecting the daily fluctuations in solar energy flux; (3) a transient component derived from rain events. The duration of rains is brief and they are irregular in occurrence.

In the following sections the effect of this sort of input function on spring flow will be briefly and qualitatively discussed.

The Big Spring was gauged initially on the assumption that it was the only significant output to the aquifer. Figure 8 shows, however, that its response to the input is far from straightforward:

(1) There is scarcely any response to transient events.

(2) The diurnal component is lagged and disappears completely at high stage, but is extremely strong during "recession."

(3) The seasonal component appears to be a "step function."

(4) The high frequency irregularity at the flow plateau is real. It is a highly aliased representation of surging of the Big Spring with a period of ≤ 5 min.

A simple interpretation of this behavior is that the Big Spring occupies a place in a hierarchy of springs. During high flow conditions the active springs are unable to handle the input to the system. The water is stored in the aquifer and the "water table" rises. This will continue until

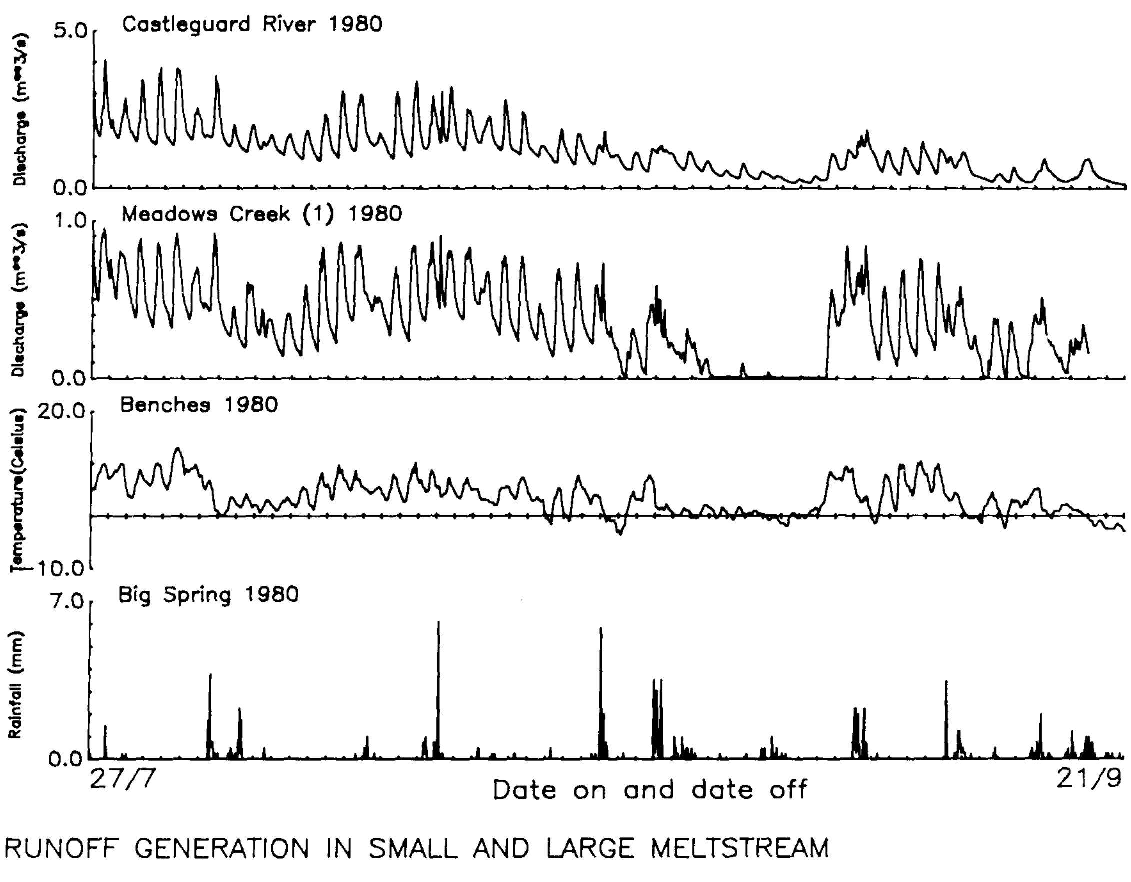

FIGURE 7. Castleguard River (recorder no. 1, Figure 1) and Meadows Creek (recorder no. 2, Figure 1) are glacial melt streams. The more subdued response of the Castleguard River to varying weather conditions reflects the lower altitude and larger size of its source glacier.

a higher level spring comes into operation. If all diurnal variations in flow are within the capacity of the top spring, then a uniform head will be maintained on lower springs and their flow will be steady. The top spring may be referred to as an *"overflow" spring* and the steady, lower level springs as *"underflow" springs.*

The deterioration of spring flow after 2 September is not at an exponential rate, suggesting that the Big Spring is also served by underflow members. The strong pulsing seen through 13 to 20 September is characteristic of overflow systems.

It is thus apparent that the relationship between input and the Big Spring is strongly dependent on flow conditions in the aquifer. The responses vary from relative amplification to complete attenuation.

Exploration of the lower Castleguard Valley revealed numerous underflow springs. A definite downvalley limit to them was never determined. However, the Stephen shale dips downvalley and is encountered in the floor some 5 km below the Big Spring. This would probably confine the aquifer, providing there were no permeable fractures.

Subsequently, other springs downvalley were monitored, both individually and collectively. The collective monitoring was done by gauging Meadows Creek at two locations 1.5 km apart. Numerous springs appeared in this reach, adding up to 6 $m^3 s^{-1}$ to the flow. These springs have a complex behavior, acting as an underflow-overflow system (Figure 9). However, it can be seen that they also reflect the inflow discharge, suggesting that a component of the input discharge is derived from the upstream Meadows Creek. An approximate mass balance confirms that there is far less water flowing in the lower Meadows Creek than in the contributory streams in the upper and middle Meadows. This confirms that water is leaking into the aquifer through the channel bed. It is draining the Meadows Creek and Cave Creek where they run along the floor of the Castleguard Valley.

The overflow spring maintaining constant head on the Big Spring was recognized some 100 m away and at a similar elevation, but it was not monitored continuously. It was hypothesized initially that the floods issuing from Castleguard Cave were the overflow. These floods are highly characteristic of an overflow (Figure 8). However, it can be seen that the relationship is not consistent, cave floods occurring during both diurnal and steady regimes of the Big Spring. The 280-m rise in water level required is also very unlikely, given the apparently abandoned springs lying between the two levels.

The exact provenance of the cave floods is unclear. Early floods are associated with the disappearance of snow in the Upper Meadows, which suggests that this is

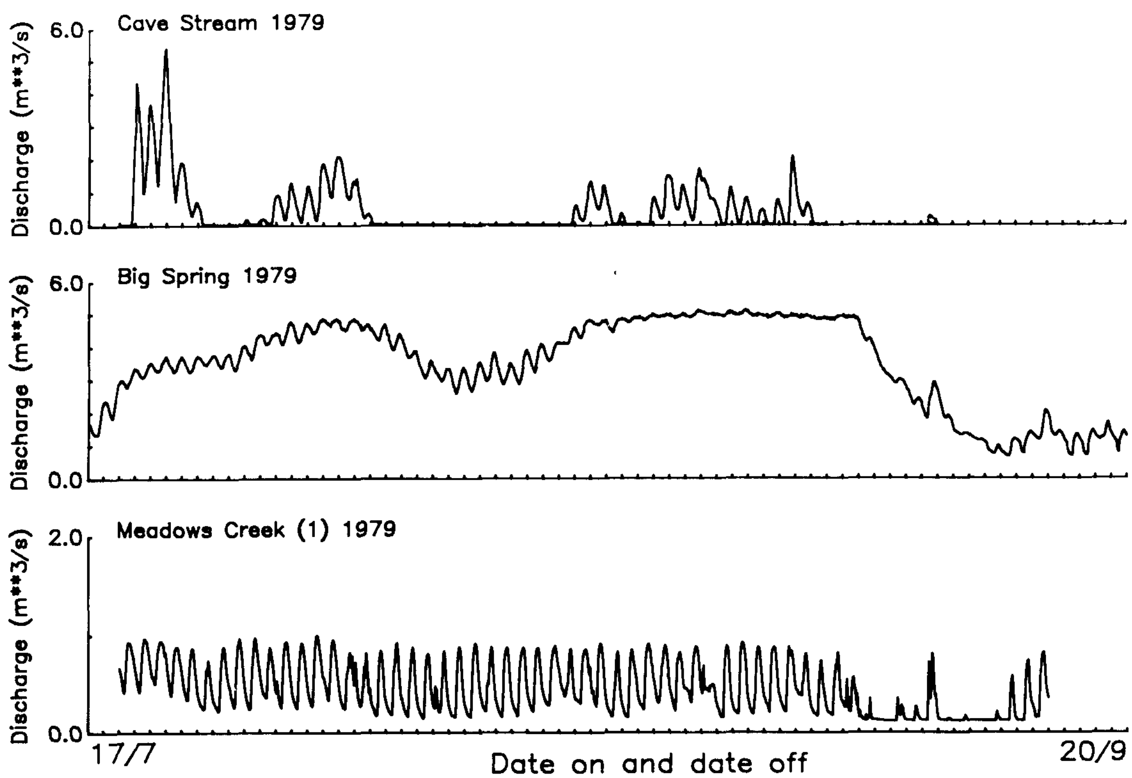

FIGURE 8. The nonlinear relationship between glacial melt and springflow results from the complex hierarchy of karst springs. The Big Spring (recorder no. 3, Figure 1) varies between an overflow and an underflow system. The Cave Stream (recorder no. 4, Figure 1) hydrograph is dominated by overflow flood events, but shows an inconsistent relationship to Big Spring discharge.

the source area. This is not possible in summer when the area has little surface water. Dye traces from the South Benches were made concurrent with cave floods and proved negative.

The Saskatchewan Glacier is the most probable source of the floods. The trace from the glacier was made when the cave was not flooding, and went to the Big Spring. The nature of the overflow mechanism is unclear. It is either a response to a rise in englacial water levels, or an overwhelming of the capacity of a low level conduit linking the glacier to the valley springs. The overflow route may well have developed when the glacier level was higher and an Upper Meadows-Cave conduit (Castleguard III) was exploited. This is not the same as the route dye-traced from the Col Karst.

From these data, the active karst aquifer appears to be somewhat of a contrast to the relict Castleguard Cave. The latter was a single discrete conduit, while the present system has dispersed inputs and outputs and restricted flow in the lower springs. A simple model of the modern aquifer may be proposed: a series of restricted shafts drain down into a complex reticulate network of fissures which discharges at the numerous springs. This model may be evaluated by quantitative dye tracing.

Fluorometric Dye Tracing

Fluorescent dyes have proven to be useful tracers in a wide range of environments (e.g., Atkinson et al., 1973; Smart and Smith, 1976), because of their relatively low cost and formidable field detectability (for Rhodamine WT. ca. 0.01 mg m^{-3} at Castleguard). Some 30 quantitative traces were made at Castleguard, and are summarized here (see also Smart and Ford, 1982).

The fluorescent dyes, Rhodamine WT, Lissamine FF, and Fluorescein were selected. Injections of up to 4.5 kg (dry weight) of dye were made into accessible sinkholes on the South Benches of Castleguard Mountain, the Col Karst on the Meadows, and in one case from a marginal stream on the Saskatchewan Glacier. Replicate traces were made from the South Benches. It was not possible to trace from the more outlying regions of the Columbia Icefield.

Sampling was with automatic water samplers collecting up to four samples (multiplexed) per hour. Continuous flow fluorometry was also employed. The instrument was a 12-volt-operated, Turner Designs® Model 10 Series Fluorometer. It was calibrated before each batch of samples and measurements were corrected for temperature (Smart and Laidlaw, 1977). Breakthrough curves

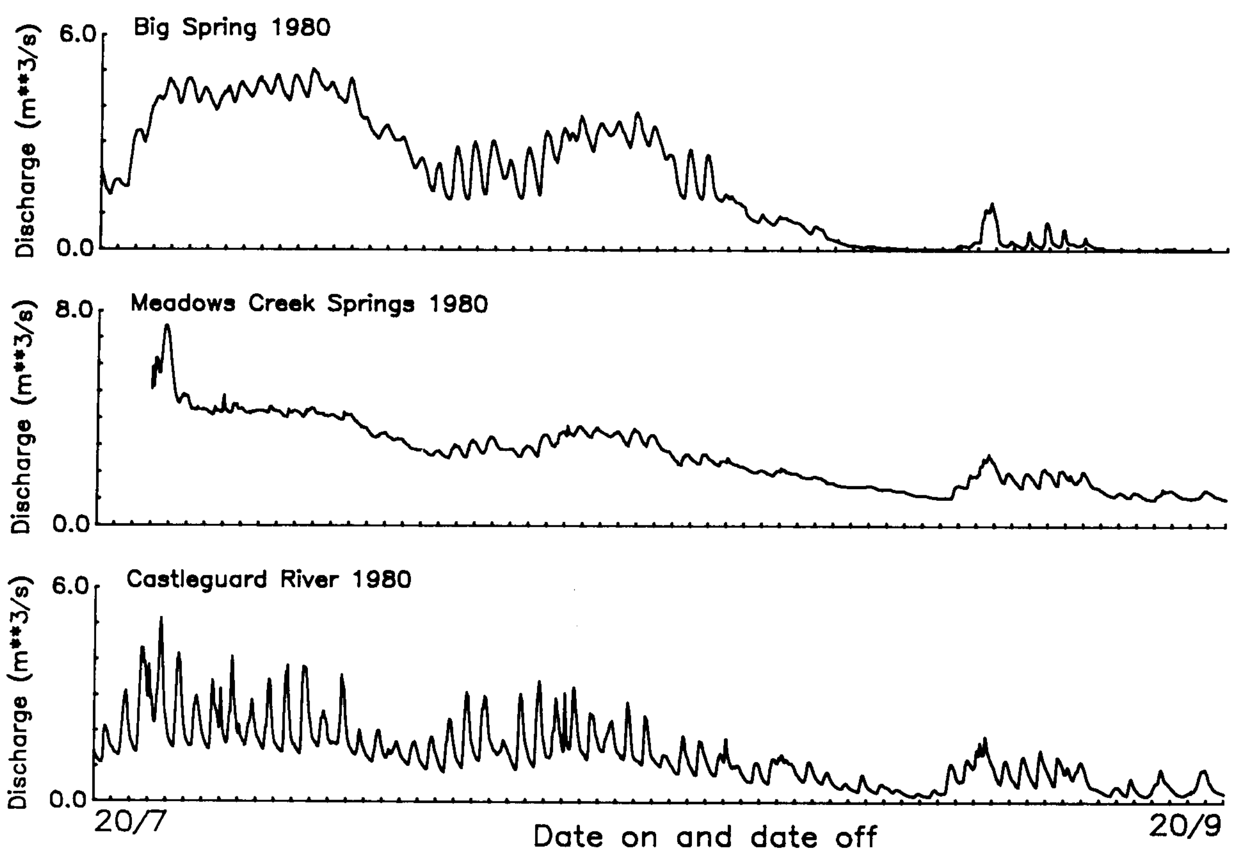

Figure 9. The Meadows Creek Springs discharge is the aggregate flow from numerous springs. Nevertheless, their behavior typifies an underflow, with steady discharge at high flow and more sustained discharge during recessions.

were standardized to a 1-kg injection for comparative purposes. As a backup procedure, activated charcoal detectors were used where a point-to-point result was of significance.

The catchment delineation by dye tracing has already been reviewed (Figure 1). The form of the dye breakthrough (time-concentration) curve and the time of travel are of concern here.

First arrival time from the South Benches varied from 3.5 to 12 h over the distance of 4 km and fall of 750 m, with average velocities of up to 650 m h^{-1}. The dye was not strongly dispersed; 90% of total recovery took 5 to 10 h depending on location and stage.

Spring groups may be identified by similarities in their breakthrough curves. Only time of arrival and dispersion will vary within a group, as these are properties of distance. The springs around Big Spring, and for 1 km down the Castleguard River behaved as one group. Those feeding Meadows Creek and extending down to Artesian Spring also produced breakthrough curves which were broadly similar, suggesting that they constitute a second group.

Figure 10 shows representative breakthrough curves from each group. The common bimodal form suggests that there was a splitting and recombining of flow underground before final distribution to the two groups. These are 1.4 km apart, but time of travel and dispersion are similar. The clear distinction between them is the comparative magnitude of the two curves. There is a relative *dilution* of the Meadows Creek springs when compared to the Castleguard River spring.

Figure 11 shows a comparable trace made under much lower flow conditions. The dye has traveled less rapidly and the distinctive second peak has disappeared; the alternative flow route has been abandoned. Note that the dilution effect is reversed.

The broad distribution of the dye among the springs supports the notion of a reticulate network. However, such a structure is contradicted by the high flow velocities and low dispersion of dye, which suggest a discrete conduit similar to the explored cave. A compromise structure is that of a discrete conduit with a complex distributary system. This could develop if a mature and well-developed conduit were blocked by debris or collapse. The extensive glacial deposits in the Castleguard Valley may have obstructed a preexisting cave. Continued flow in the system built up water levels within the aquifer and numerous outlets were exploited. The resulting multitude of outlets would have evolved and enlarged until competent to handle peak flows. Continued solution beyond this point would cause the highest springs to be abandoned progressively, creating the relicts observed today.

The dilution of Meadows Creek Springs at high stage may be explained by a contribution from an independent source area, or local aquifer recharge from the creek bed. At low stage this dilution does not occur, giving a tenfold increase in peak dye concentration. The consistent dye concentration in the Castleguard River Springs is less straightforward. Concentration should increase with decrease in discharge. The increase in turbidity of the spring waters at low flow suggests an increased contribution from a turbid stream like the Castleguard River. It is hypothesized that there is an increased influx of Castleguard River water at low flow, and that this has offset the increased dye concentration caused by low flow in the aquifer as a whole. This demands estavelles in the Castleguard River. The deep entrenchment of the channel into the Cathedral limestone might allow such an exchange to take place, although no estavelles were observed.

These dye traces demonstrate that the aquifer is well developed. Glacial disruption has produced the numerous springs of the Castleguard Valley. The complexity of aquifer behavior results from the varying contributions from different source areas: the central icefield, the Saskatchewan Glacier, channel losses, and proglacial sources. Only the latter influx was quantitatively traced.

DISCUSSION

The Hydrological Role of Glacier Ice

Most meltwater is produced at the surface of glaciers and must be routed to the bed through crevasses and moulins. Distribution of tensile stress within the ice is therefore the dominant variable controlling water penetration to bedrock, although other factors determine the quantity involved.

A great deal of water enters the karst aquifer through the headward complex of Castleguard Cave. The nearest crevasses are 200 to 400 m away, so that there must be interconnecting basal channels. The perennial draught demonstrates that these are always at atmospheric pressure. Therefore, in summer flow is in free-surface streams, and in winter plastic deformation beneath 200 to 300 m of ice fails to close the channels. Such channels must be partly contained in bedrock (called Nye channels, Figure 2); they are probably at prominent joint faces, making them promising locations for karst development. Furthermore, the low pressure in such channels will permit them to capture developing englacial channels (Röthlisberger channels).

Subsole drift constitutes the major local constraint on subglacial karst development, because of its low permeability. Water passing through carbonate drift will also exhaust much of its solutional potential. Some drift has possibly blocked conduits in the headward complex of the cave. However, in general the massive carbonates produce little drift. If the clear spring waters are removing the basal debris produced in their catchment (e.g., Collins, 1979), then remarkably little abrasion is taking place.

Proglacial and subglacial karst shafts on the South

DILUTION UNDER HIGH FLOW

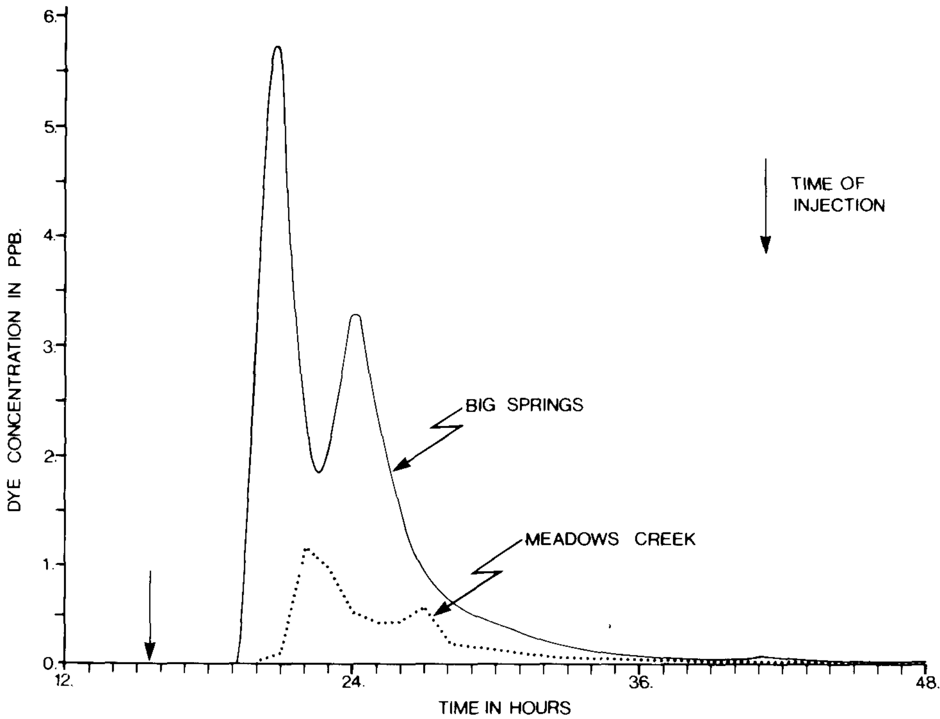

FIGURE 10. Breakthrough curves to two springs under high flow conditions. Meadows Creek Springs are being diluted, by flow from an independent conduit or by water sinking in the channel farther upstream.

DILUTION UNDER LOW FLOW

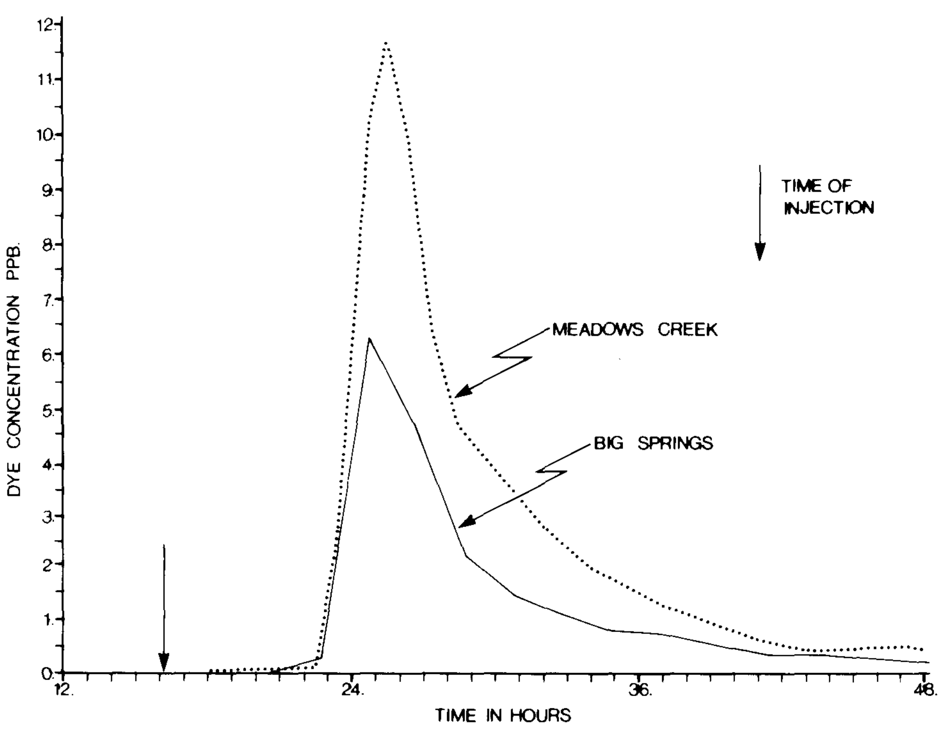

FIGURE 11. Breakthrough curves to two springs under low flow conditions. Compared to behavior shown in Figure 10, dilution is reversed. Possibly the Castleguard River is leaking into the aquifer above the springs.

Benches are able to engulf all of the locally produced meltwater. Joint faces oblique to the ice flow direction provide both sink points and potential subglacial channels to feed them. In the absence of debris, there is little constraint on their subglacial development. However, the low aggressiveness of subglacial waters should permit only very slow growth. This is demonstrated by the age of 6100 BP for a speleothem sample obtained from a shaft that is still enlarging. The shaft antedates the speleothem, yet it is still very constricted (ca. 1×0.1 m).

In marked contrast is the behavior of basal regelation water (e.g., Hallet, 1979a, 1979b). The benches around Castleguard Mountain host extensive subglacial precipitate deposits (Ford et al., 1970). These are formed by the melt-solution, freeze-precipitation of carbonate bedrock by clean basal ice (Hallet, 1976a). On vertical faces small "stalactite" forms occur where water has passed through tiny fissures in the bedrock (Smart, 1981). Hallet (1977) has suggested that localized regelation water may attain relatively high partial pressures of CO_2. It is therefore possible that regelation water is responsible for the speleothems growing in the cave beneath glacier ice. The high pressure characteristic of regelation waters (ca. 79% of ice overburden, Hodge, 1979) would enhance penetration into bedrock when compared to low pressure conduit water. It is possible that regelation water is responsible for the initiation of subglacial karst.

Glaciation and Karst

The geomorphic interaction between ice and karst have been discussed elsewhere (e.g., Nicod, 1976; Maire, 1977; Ford, 1979), and it is the hydrological effects which are of concern here.

The main effects of glaciation will be a reordering of relative elevations and a decrease of permeability through plugging of entries and exits of the aquifer. The net effect may be drainage, stagnation or stimulation of an aquifer.

Karst development can occur subglacially, as is clear from subaerial, relict conduits. Regelation waters may initiate the karst because of their high pressure. The volumes of regelation water available are insufficient for further development, however, and the involvement of conduit waters is necessary for continued growth.

In suitable locations, a functioning subglacial karst aquifer may drain the glacier bed, decreasing basal water pressures. It has been suggested that the loss of regelation and conduit waters to the ground will remove heat from, and reduce the geothermal heat flux to, the glacier bed (Ford et al., 1976; Drake, pers. comm., 1982). In addition, Hallet (1976b) has suggested that solutes in regelation water constrain basal sliding. The decrease in basal water pressure and removal of basal heat enhance this effect. A decrease in basal sliding and basal melt will decrease erosion rates (Boulton, 1974; Hallet 1979a). Thus the apparent "resistance" of carbonate rocks to glacial erosion may be as much hydrologically as lithologically based.

The behavior of a glacier "constrained" by basal drainage is hard to predict. Flow models may be inappropriate and the mass balance affected by decreased ice flow.

In postglacial conditions, karst aquifers have to adjust to the redefined topography. Springs are often no longer at base level and downcutting or backflooding may occur (Palmer, 1977; Ford, 1979; Mylroie, 1981; Ford, 1983a). The rate at which readjustment takes place depends on the erosional power of the water. At Castleguard, the depleted solutional capacity of meltwaters has resulted in relatively little change since deglaciation of the valley (ca. 10,000 BP).

In contrast to the karst processes, deep fluvial canyons have been cut during postglacial times in the Rockies. In places such as Castleguard, this has led to incidental interactions between surface streams and the karst aquifer.

CONCLUSIONS

Castleguard provides a remarkable juxtaposition of active karst beneath glacier ice and a spectrum of previously glaciated karst. The cave allows insight into present processes and the past history of the karst aquifer.

It is clear that three conditions are essential for karst aquifers to function subglacially:

(1) Penetration of surface meltwater to the glacier bed.

(2) Clean basal ice and the absence of subsole drift, obtained only on the more massive and conformably bedded rocks.

(3) Sufficient hydraulic gradient to permit circulation.

The generation of karst subglacially is more speculative. However, local development seems possible, providing that sufficient time exists for development, or if preexisting discharge routes are available within the aquifer. Otherwise, the limited solutional capacity of the waters is soon depleted, especially in vadose shafts.

The distinction between conduit and regelation waters made by glaciologists (e.g., Hallet, 1979b) appears valid in subglacial karst groundwaters also.

Glacial disruption and blocking of karst outlets and valley incision can create a spring complex typified by underflow-overflow behavior. The apparent youth of this feature is contradicted at Castleguard by the very rapid transmission of fluorescent tracers. An ancient trunk conduit, much like Castleguard Cave, must be feeding the springs in the Castleguard Valley.

The loss of subglacial waters to an underlying karst aquifer will generally retard basal sliding and erosion rates. In contrast, where groundwater discharges beneath glacier ice (contributing heat and increasing the basal pressure), an increase in ice velocity and erosion rates is to be expected.

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