THE HYDROLOGY OF A GLACIERISED ALPINE KARST: CASTLEGUARD MOUNTAIN, ALBERTA

By

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ABSTRACT

Alpine karst throughout the world has been affected by past glaciation, and yet little is known of the interactions between glacier ice and karst. This dissertation attempts to gain some understanding of the problem through the study of the Castleguard Area, Alberta, where a karst aquifer is presently overlain by temperate glacier ice.

Quantitative fluorometric tracing and hydrometric measurements generated a broad data base on aquifer behaviour. Tracer breakthrough curves were interpreted using a new systematic approach which considers an explicit set of processes likely to affect the particular tracer under the given experimental conditions. Non-linearity in aquifer behaviour and rapid groundwater velocities demonstrated the aquifer to be an extreme conduit type. Conduit springs are elements in a vertical hierarchy, in which the topmost springs are "overflows", and exhibit greater flow variability than their associated "underflows". A numerical model was developed to simulate a conduit aquifer. It demonstrated that pulse train and recession analysis, widely accepted methods of karst

aquifer investigation, could be rather misleading when applied to conduit aquifers.

Interactions between ice and groundwater were observed at two scales: regelation water appeared to feed a diffuse percolation system, and supraglacial melt passed into subglacial conduits which entered open vadose shafts. Karst is unlikely to be entirely subglacial in origin, because of the limited aggressiveness of subglacial waters.

The Castleguard karst appeared to have originated preglacially in response to the breaching of impermeable caprock. Glaciation re-ordered the landscape and produced abundant clastic debris which subsequently blocked or obstructed karst conduits. Much of the resulting karst is paragenetic and comparatively immature due to glacial disruption and slow growth rates. Geomorphic and hydrologic interactions between ice and karst depend intimately upon the relationship between the geographic zones of the glacier and the aquifer.

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This thesis is dedicated to the late Professor E.K.Tratman, who so generously and enthusiastically showed me the joys of karst research.

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The force that drives the water through the rocks
Drives my red blood,
That dries the mouthing stream, turns mine to dust,
And I am dumb to mouth unto my veins
How at the mountain spring the same mouth sucks.

D.M.Thomas; The Poems.

CHAPTER ONE

INTRODUCTION

The characteristic caves, closed depressions and underground drainage of the karst landscape can be observed to some extent in every physiographic environment. Alpine karst is one distinctive type developed in regions of great local relief where the geomorphology and hydrology are significantly affected by snow and ice. Many alpine karst areas are presently ice free, although they may have been intensively glaciated during the Pleistocene. Their landforms correspondingly reflect the impact of past glaciation and are presently relic forms. While the evidence for past glaciation is clear, the past behaviour of the karst when it was beneath ice is much less obvious, and as a result, rather speculative inferences have been made concerning glacier-karst interaction. This study attempts to establish the nature of active processes of glacier-karst interaction. Field work took place in the Castleguard karst, a location where temperate glacier ice presently overlies hydrologically active karst.

The Castleguard karst lies at the northern end of Banff National Park in the Main Ranges of the Canadian Rocky Mountains. Local relief exceeds 500 m, and the adjacent Columbia Icefield is the largest extant ice body in the Rocky Mountains. Drawing on hydrochemical, hydrological and geomorphological evidence, Ford (1971a) inferred that the Columbia Icefield furnished the majority of the water emerging from karst springs in the Castleguard Valley. In other known glacierised karsts (eg. Maire 1978) the glacier ice obscures the karst and little direct interaction can be observed. At Castleguard, an extensive cave system passes at shallow depth beneath the Columbia Icefield, providing a unique opportunity to study subglacial hydrology. Unfortunately, parts of the cave are flooded in summer, and direct observation is not possible during the peak ablation season. However, recent retreat of glaciers in the area has exposed extensive "neoglacial pavements" which provide comparatively undisturbed examples of a karst glacier bed.

The inaccessibility of the subglacial karst meant that a direct study of hydrological and hydrochemical processes was not possible. Instead it was decided that a study of karst aquifer behaviour might reveal something of the role of the glacier ice, and that observations from the cave and an awareness of current work on glacier hydrology could supplement this data. The uncertainties of tackling the problem in this indirect manner were balanced by the advantages of the short-lived ablation season and the diurnal flow regime. These meant that a great deal of hydrological information could be gathered in a relatively short but intensive field season.

The majority of the hydrologically active karst is inaccessible at Castleguard and, like most karst aquifers must initially be considered an undifferentiated "black box". Preliminary results showed that aquifer behaviour was considerably more complex than had been initially supposed, and it was difficult to distinguish the effects of the aquifer from the effects of the overlying glacier. As a result, the study became increasingly committed towards an attempt to resolve the nature of the aquifer. This involved both passive hydrological measurements and active water tracing. These studies revealed the aquifer to be an extreme type of karst system, dominated by rapid conduit flow. Novel approaches had to be developed for its study. These methods centered around high precision water tracing, and the development of a new numerical model for simulating conduit aquifers. The result of these studies was a reasonably satisfactory model for the conduit system, which then provided a basis for evaluating the past and present impact of glacier ice on the Castleguard karst.

The present work constitutes the most intensive hydrological study to date in a glacierised karst, and has developed techniques applicable to karst systems in general. However, in a broader sense it provides some basic data on the hydrological behaviour of a karst catchment in the Rocky Mountains. Many of the rivers rising in the Rockies supply water to more arid areas of Western Canada. Both glaciers and groundwater are important in moderating flow in these rivers (Drake and Ford 1974, Ozoray 1977, Young 1977), but have so far received relatively little detailed study.

CHAPTER TWO

THE CONTEXT:

GLACIER HYDROLOGY, ALPINE KARST

AND

THE CASTLEGUARD AREA

Alpine karst is characterised by high relative relief and the presence of snow and ice. Castleguard is an outstanding example of active alpine karst; the local relief exceeds 1000m and exposed karst features adjoin an ice cap of 325 square kilometres.

2.1. GLACIER HYDROLOGY

2.1.1. SUBGLACIAL WATERS

The dominant hydrological feature of glacier ice is the release of winter precipitation during a relatively brief ablation season. In general, there is an inverse temporal relationship between ablation and precipitation which means that glaciers are important in regulating stream flow by acting as natural dynamic reservoirs (Young 1977). The glaciers of interest here are "temperate" in that the bulk of the ice is always at pressure melting point throughout the year, although the surface will be colder in winter, and "cold spots" may develop on the glacier bed. Conditions beneath a cold based glacier or a major ice sheet will be quite different, but are not of direct concern here.

Most water discharged from a glacier is derived from surface melt; this may amount to several metres per year. If the ice is at pressure melting point, then geothermal heat and viscous flow will also generate some 6 mm/a of water (Weertman 1972). Much of this water is routed through the glacier to the portal. Little observational information is available concerning this flow route, but studies on deglaciated surfaces (Walder and Hallet 1979), and using boreholes in ice (e.g.Hodge 1976, 1979, Engelhardt 1978) have supplemented more theoretical descriptions (e.g.Nye 1976, Rothlisberger 1972, Shreve 1972, Weertman 1972). The resulting general model suggests that there is an integrated network of discrete conduits, more or less weakly connected to an extensive, but discontinuous, thin subglacial water film.

The conduit network is at approximately atmospheric pressure, and serves primarily to transport surface meltwaters through the glacier. The network is a response to the relative disposition of recharge and discharge points. The bulk permeability of glacier ice is pressure dependant, but permits a water flux of less than 0.1 mm/a at depth (Raymond and Harrison, 1975), so that most surface melt water therefore reaches the glacier bed through cracks and fissures, and the more obvious moulins and crevasses. In general the topology of subglacial channels is controlled by the basal pressure field which is a reflection of ice thickness and basal shear stress. In addition, subglacial conduits represent an equilibrium between enlargement by melting and closure by ice deformation. This results in a competitive advantage for larger channels so that a dendritic network will tend to evolve (Shreve 1972).

Subglacial channels may be formed either in ice or in bedrock. The former are referred to as Rothlisberger channels, and are prone to closure by ice deformation and flow around obstacles. The latter channels are called " Nye-channels" and are cut in bedrock so that they are more resistant to closure by ice flow. This will encourage water flow to be renewed along the same axis in succeeding years, despite changes in glacier conditions. In addition, Weertman (1972) has shown that Nye channels have the ability to drain water from the subglacial water film under flowing ice.

The subglacial water film is derived from basal and frictional meltwaters, and the occasional entry of water from overpressurised conduits. The water film appears to be a few micrometres in thickness, but on occasion may thicken (Hallet 1979). These thickening events have been associated with spring runoff and uplift of the glacier surface (Iken et al.1979), and with occasional transient rainfall events exceeding conduit capacity (Engelhardt 1978). Walder and Hallet (1979) suggest that there are also larger discrete water bodies of a few square metres in area at the glacier bed. The pressure in the basal film varies from 50% to over 100% of ice overburden pressure. This variability is explained by the presence of areas of direct ice-bedrock contact (Goodman et al.1979) and the the occasional ablity of glacier ice to confine water under artesian conditions (Baranowski 1973, Boulton et al.1974).

The glacier bed is not usually a clean ice-bedrock contact. Many lithologies are readily abraded, producing a layer of heterogeneous and plastic subsole drift (Engelhardt 1978), which may infill and occlude conduits and generally limit the hydraulic continuity implicit in the above model.

2.1.2. SUBGLACIAL HYDROCHEMISTRY

By definition, chemical processes are dominant in the development of karst, thus in order to understand subglacial karst, it is necessary to consider the nature of subglacial solution chemistry.

The low partial pressures of carbon dioxide in glacial meltwaters (Ek 1964,1965, Ford 1971) result in a relatively low solution capacity. Flow route, residence time and mixing determine to what extent this feeble solution potential is realised. Turbid subglacial and proglacial waters may rapidly gain total hardnesses of 30-40ppm (Collins 1979, Ford 1971, Reynolds and Johnson 1972) due to the ready solution of the crushed crystal lattices typically produced by glacial abrasion. Much of the solutional capacity is expended on colloidal and suspended sediment which considerably reduces the geomorphic impact.

Chemical conditions in the subglacial film remain poorly understood, but they are dominated by the high overburden pressure and the cyclic solution-precipitation induced by regelation processes. Hallet (1976a) has described chemical conditions for a carbonate bed exhibiting local solution-precipitation features. The high ambient pressure at the glacier bed allows a large quantity of carbon dioxide to be held in solution. The critical questions are how much carbon dioxide is available at the glacier bed, and what the realised partial pressures are.

The evolution of polycrystalline ice from snow results in a gradual purging of impurities over time. Physical exclusion by sintering, removal in percolating waters and purification by zone melting result in an impressive reduction of impurities in the passage of ice through a glacier (Berner et al.1977, Glen et al.1977). In the basal zone of a glacier these processes are complicated by the refreezing of squeezed interstitial water and pressure melt (Souchez et al.1978). On freezing, the low partition coefficient between ice and water for all solutes results in supersaturation and precipitation of the solutes, either as subglacial precipitates, solid inclusions, or gas bubbles in ice (Hallet 1976b, Hallet et al.1978). Freezing therefore provides a mechanism for concentrating both dissolved solids and gases at the base of a temperate glacier.

Gas bubbles in basal ice (Kamb and LaChappelle 1968) demonstrate that conditions of gaseous supersaturation occur occasionally in the basal water film. These bubbles exist under high pressure, and high partial pressure of the component gases is possible.

The composition of basal gas bubbles is unknown, but measurements on gases in bulk temperate glacier ice show a depletion that is compatible with meltwater stripping (Weiss et al.1972). Carbon dioxide, the gas of concern here, is an exception. The proportion of carbon dioxide increases in relation to total gas content down glacier, although the bulk concentration decreases. Stauffer and Berner (1978) attribute this to dissolution within the ice, and possibly to clathrate formation (Miller 1969, 1973).

It appears therefore that carbon dioxide is relatively resistant to englacial removal, and may become somewhat concentrated in the basal ice of temperate glaciers. The crucial difference between conduit and film waters is the relatively high ambient pressures and the low water flux in the latter.

The isotopes in glacier ice provide a valuable natural tracer, although they are affected by numerous processes. The deuterium content of precipitation decreases with altitude and latitude, and in winter at a given location (Krouse 1970, Moser et al.1972, Moser and Stichler 1981) Summer precipitation is generally lighter than ablation water (Ambach et al.1976). Furthermore, the age of glacier ice results in an extremely low tritium content (Ambach et al.1973). Within a particular glacier the isotopic signature of the source area is modified by firnification and ice metamorphism. Data rather weakly support the flow concept of Reid (1896) in which high altitude accumulation is subducted beneath the equilibrium line to reappear at lower altitude in the ablation zone (Epstein and Sharp 1959, Hambrey 1974, Krouse 1970, Lawson and Kulla 1976).

The isotopic effects of ice metamorphism are complicated because the fractionation depends largely on the degree of completion of freezing (Jouzel and Souchez 1982). Melting is known to have little isotope effect, compared to partial refreezing which can cause significant fractionation. This effect is probably responsible for the isotopic enrichment of basal ice (e.g.Lawson and Kulla 1978).

Runoff from alpine glaciers shows both seasonal and diurnal variations in isotopic composition. Seasonally, spring runoff is lightest, with marked variations in summer and a steady heavy base flow (if any) in winter (Ambach et al.1976). Diurnally, the peak flow is heaviest; this is conventionally explained by a rather unsatisfactory ablation flow-base flow mixing model (e.g.Behrens et al.1971, Ambach et al.1976). Summer precipitation events have a marked effect on the composition of glacier runoff, because of their distinct isotopic composition and the reduction in ablation during precipitation events.

2.1.3. GLACIAL GROUNDWATERS

The nature and behaviour of groundwater beneath glaciers is a highly speculative subject. It is difficult to differentiate groundwater from subglacial water with a similar residence time on any a priori physical or chemical basis. This is a major criticism of those who attempt flow separation of glacier runoff by conductimetry (e.g.Collins and Young 1979).

Clayton and Moran (1974) have developed a groundwater model for continental ice sheets overlying permeable material. The groundwater potential gradient reflects the basal pressure field, and is strongest near the glacier margin. This also corresponds to the frozen zone, which dramatically decreases substrate permeability. These conditions are believed to have a particular landform association (Moran et al.1979).

The continental model is inappropriate for conditions beneath temperate alpine glaciers, such as those at Castleguard. Ford (1979, 1983) has presented a model for the alpine situation in which subglacial groundwater potential is enhanced beneath ice caps located above unglacierised valleys. When ice occupies these valleys the potential is dramatically decreased, and groundwater flow may stagnate. Several examples of active subglacial karst have been described by Maire (1978), but Ford et al.(1976) provide the most detailed account of conditions in such an aquifer. The main effect of subglacial groundwater flow appears to be the diminution of basal geothermal heat flux, because of heat abstraction by circulating water. Further observations on subglacial conditions are presented later in Chapter Seven.

The potential for groundwater flow beneath glaciers is only realised when suitable flow routes exist. It is speculated that karst may develop beneath temperate ice, but in many examples subglacial waters are probably following pathways developed prior to glacierisation.

2.2. ALPINE KARST

An appropriate definition of the karst aquifer is necessary before the subcategory of alpine karst can be identified. Bakalowicz and Mangin (1980) define the karst aquifer as one in which a characteristic heterogeneity is developing from the organisation of underground flow routes into a structured hierarchy. They identify three zones: the epikarstic (or subcutaneous) aquifer, the unsaturated zone (characterised by the vertical movement of water), and the saturated zone in which water in conveyed in predominantly horizontal conduits and storage occurs in semi-isolated voids termed "systemes annexes". A convenient extension of this definition for the hydrogeomorphologist is that Karst is a landscape assemblage evolving towards an organised routing of surface waters into a karst aquifer as defined above. Abandoned caves; palaeokarst and micro-features do not fall into this definition, although glaciers may be considered to be karst aquifers.

Within this definition alpine karst may conveniently be defined as a karst system characterised by high hydraulic gradients and a marked seasonality in the availability of water in response to the accumulation and ablation of snow and ice.

The classical literature on alpine karst comes from the mountainous regions of Europe. The Muotatal valley in Switzerland and the Picos de Europa are considered type localities (Bogli 1964, Miotke 1968), although both areas are now devoid of glacier ice. Much of the European literature is concerned with geomorphological classification, and inevitably, a major concern in the Alps is with geologic structure. The two classical types of glaciated karst are Schichtreppen (stepped) karst and Rundhocker (Roche Moutonee) karst, formed respectively on gentle and steeply dipping strata (Bogli 1964). Nicod (1974,1976,1978) is a leading proponent of the geomorphological school, writing primarily descriptive reports centred on geomorphic mapping and concentrating accordingly on the classification of form. This approach encourages implicit assumptions concerning landform evolution, and contributes little to our understanding of the systematic function and evolution of karst and the karst aquifer. The definition of karst stated above would be roundly rejected by this group.

Geomorphic observation has provided some indications on the interaction between ice and karst, however. Miotke (1968) suggested that closed depressions were preferred locations for the initiation of glaciers, and that perennial ice forms at lower altitude in karst areas. In contrast Barrere (1964) noted that glaciers overlying karst appeared to have been less active than those on adjacent non-karst rocks.

A variant on the geomorphic approach is the speleological bias of Maire (1976, 1977a,b, 1978). Alpine cave systems are interpreted largely in terms of geologic structure.

Exceptions in the European work are the thesis of Wildberger (1981) and recen collective research on the

Muotatal in Switzerland (Behrens et al.1981). Wildberger draws on considerable hydrological and hydrochemical data to characterise the alpine karst of the Rawil region. The Muotatal study describes a systematic tracer study undertaken by a large team. This region proved to have a most complex hydrology, including multiple spring groups, tracer-dependant divides, and flow routes passing beneath valleys and rising at springs above the intervening valley floor level. The tracer work followed that initiated by Zotl (1974) in Austria, and the methods and locale are similar to those of the present work.

The interactions described between alpine karst and glaciers are complex. The local effects are some compromise between solutional enhancement caused by proglacial discharge and suppression by the debris characteristically produced in the glacial environment. Observations in glaciated karst (Horn 1935, Lauritzen 1981, P.L.Smart and P.W.Williams pers comm.) show that certain conduits could only have functioned subglacially: ponors on divides or on roches moutonnes, and ascending passages in valley walls are examples. Walder and Hallet (1979) describe previously subglacial karst from exposed proglacial bedrock areas. The irregular subglacial drainage system inferred from surface morphology was not clearly integrated with the karst.

The karst aquifer may be dramatically affected by glaciation. In particular, Ford (1979, 1983) identifies nine effects, which are broadly classed under destruction, derangement, inhibition, preservation and enhancement. Ford points out that while some of these effects have been observed, others remain hypothetical. In the alpine situation, glacial dissection and deposition have resulted in extensive destruction and derangement of karst aquifers. The impact of extant ice remains speculative, but depends on the distribution and extent of the ice body. An example of postglacial derrangement is described by Brown (1972) from the central Canadian Rockies, in which karst has been buried at the sinks by a postglacial landslide, and at the risings by fluvioglacial and fluvial sedimentation. Where postglacial entrenchment has occured, the karst may be rejuvenated. Palmer (1977) and Mylroie (1981) confirm these effects in lower relief areas.

The ability of caves to survive glaciation has made them a useful palaeoenvironmental resource. There appears to have been a decrease in speleothem production during glaciations (Atkinson et al.1976, Gasgoyne 1979 p246, Harmon et al.1977), which suggests reduced vadose percolation. Speleothems from beneath glacier ice in Castleguard Cave suggest that this need not be an absolute cessation. However, major infilling events in caves appear to date from glacial epochs.

Ford (1971a) and Ford et al.(1976) have shown that temperatures beneath the Columbia Icefield are insufficent to maintain permafrost. Areas beyond the "protection" of the ice are heavily frost shattered, although groundwater appears to be able to flow through pre-existing conduits (Ford 1971a, Tolstikhin and Tolstikhin 1976). In this context glaciers may be viewed as affording protection from the permafrost regime affecting extraglacial areas.

2.3. THE CASTLEGUARD AREA

Sweeting (1972) describes Castleguard as a type locality of Schichtreppen karst. It is still largely overlain by glacier ice, so that the association between the ice and karst may be actively observed, rather than inferred from relic landforms, as has been the case in most other studies.

The Castleguard area lies in the northern extremity of Banff National park, Alberta, adjacent to the continental divide (Fig.2.1)

The landscape is dominated by Castleguard Mountain (3083m, Fig 2.1), a horn peak cut into upper Cambrian clastics and resting on a broad plinth of resistant limestones and dolomites. The plinth supports small cirque glaciers to the south and east with broad proglacial benches in front of them. On the north and west sides of the mountain lies the Columbia Icefield. This supplies the Saskatchewan and the South Castleguard Glaciers. The Icefield and these glaciers define the northern and western boundaries of the area. The glaciers become deeply crevassed where they flow over the edge of the Castleguard Mountain pedestal and occupy deep glacial troughs in their lower courses.

Castleguard Meadows is a broad transection valley extending between the Saskatchewan and Castleguard valleys. It descends 200m from north to south where it hangs 300m. To the east of the Meadows an elongated bench, matching that beneath Castleguard Mountain supports a series of variously evolved horn peaks and residual ice bodies called the Terrace Mountain Range.

Boreal forest occupies much of the Castleguard valley up to 2050m. Where a stable substrate is available, alpine meadow extends up to 2300m. Above this and in recently deglacierised areas there is little if any vegetation.

2.3.1. GEOLOGY

The lithology and structure have been described by Ford (1971, 1983) and are summarised in figure 2.2. The major karst rocks are platform carbonates of Cambrian age, with greater than 560m aggregate thickness.

The basal Cathedral Formation may be divided into two members, the Main and Upper. The Main Member is a very uniform, well bedded, massive black crystalline limestone, whereas the Upper Member includes more shallow water facies, and is more variable, including coarse and porcelaneous beds.

The Stephen Formation overlies the Cathedral Limestone and is a 60m sequence of more thinly bedded dolomites, limestones and calcareous shales. The Eldon and Pika Formations are fine grained limestones with regular dolomite laminae separating the 6-10cm limestone beds. South of Castleguard Mountain occasional 10-80cm thick, vertical sedimentary dikes dissect the Eldon and Pika Formation. Similar dikes are seen in Castleguard Cave, 350m below in the Cathedral Formation. It is not known whether there is vertical continuity between these features. Above the Pika, the Arctomys Formation commences a sequence of middle to late Cambrian shales, sandstones and dolomites.

The Cathedral, the Eldon and the Pika Formations are exceptionally massive limestones, notwithstanding their laminated composition. These are cliff-forming limestones and the infrequent, but extensive joint faces are important in this respect. The Stephen Formation is more akin to the upper shaly units, being mechanically weak and yielding abundant talus.

Hydrogeologically, the Cathedral is the predominant aquifer. Most active conduits occupy the Main Member, while Castleguard Cave occurs in the Upper Member. Where it is exposed, this limestone is clearly prone to solutional attack. The Stephen Formation is generally an aquiclude, the exceptions probably associated with joints or the sedimentary dikes which may traverse the whole unit. The frequent dolomite laminae prevent large scale karstification of the Eldon and Pika Formations. However, joints have proven suitable flow routes, and have clearly been solutionally enlarged. Where the underlying Stephen Formation acts as an aquiclude, numerous springs occur, such as along the eastern edge of Castleguard Meadows.

The local structure is unusually straightforward. The Saskatchewan and Castleguard valleys occupy the only obvious fold axes. In between, the dip is 5.5-6.5 degrees SSE. In the Castleguard valley, dip increases suddenly to 35 degrees, occasional tear fractures are evident, and structural discontinuities are seen across the Castleguard River. Extensive fractures parallel to the fold axis occur in the Castleguard Valley. These now guide surface streams and where exposed show evidence of karst solution. Traces of a north-south fracture set are seen on aerial photographs. Their hydrogeological significance is unknown.

2.3.2. GEOMORPHOLOGY

The morphology of the area is typical of alpine glaciation, although all the glaciers now appear to be retreating (Luckman and Osborn 1979, Reid and Charbonneau 1981). The Pleistocene history is poorly understood.

The central Meadows are occupied by a moraine complex, presumably Wisconsinan in age. Henoch et al.(1979) have dated postglacial material from the upper Meadows at 9600a B.P. The implicit age of retreat is supported by a date of 6000a B.P. obtained from the South Benches of Castleguard Mountain for a conventional flowstone within 50m of the present glacier. The Cavel advance of the last century has been the most major Holocene advance. Ice has retreated from 0.5 to 2.0 km in the last 90 years. Extensive moraine deposits up to 50m thick occur in the upper Castleguard valley.

Castleguard Cave holds geomorphic information extending well back into the Pleistocene (Gascoyne et al.1983). Certain speleothem samples from the cave are magnetically reversed, suggesting that it was drained at least 700,000 years ago (Gascoyne, et al.1983). This implies that local base level was below the cave elevation at that time.

The present depth of the icefield varies from 100 to 365m (Waddington and Jones) and Meir (1961) gives a thickness of from 100 to 200m for the Saskatchewan Glacier.

The morainic material left by past glacial advances now mantles large areas. At high altitudes periglacial processes are reworking this and solifluction appears to be active. Felsenmeere have developed on exposed bedrock areas beyond the neoglacial limits, except for a few resistant beds of the Upper Cathedral Formation.

Fluvial activity is largely limited to the reworking and redistribution of glacial debris. Much of the lower Meadows is mantled by alluvial fans which are active only during spring snowmelt. The Castleguard River and Meadows Creek have both cut significant gorges through thick moraines and into the Cathedral Formation in postglacial times.

The only surfaces relatively free of debris are the recently deglacierised benches around Castleguard Mountain. These surfaces host extensive fields of subglacial precipitates; solution-precipitation phenomena produced by regelation at the bed of the vanished glacier (Ford et al.1970, Hallet 1976a) The various microforms created by subglacial regelation remain imperfectly understood. Hallet (1976a,b, 1977, 1979) has demonstrated the importance of these features in the understanding of glacial processes. The general geomorphology of the area is illustrated in Figure 2.3.

2.3.3. KARST GEOMORPHOLOGY

No karst features are large enough to appear on the 1:50,000 topographic map of the area (NTS. 83C/3), because most surface forms are relatively young or relic. The Cathedral Formation in the upper Meadows contains over 100 ponors, most of which can boast little or no catchment, having apparently developed from subglacial or proglacial streams during previous ice advances. Earlier this century, Freeman (1925) showed the Saskatchewan Glacier to hold an ice-dammed lake on the upper Meadows, which presumably drained karstically. The ice now lies at least 50m lower, the lake has gone and the karst drainage has been abandoned. Further south are some depressions of several hectares in extent. These appear to function during snowmelt and rainstorms, but are lined with reworked clastic sediments. A few minor sinks on the eastern edge of the upper Meadows are nourished by springs at the Eldon-Stephen contact.

Castleguard Cave (Ford 1971, 1975, 1980, Ford et al.1983, Thompson 1976) is the major karst feature. It extends some 6.5km north east from its only known entrance which hangs some 300m above the Castleguard Valley (Fig 2.1). For most of its length the passage is a remarkably linear, discrete conduit with little branching. This in part reflects the massive nature of the bedrock, but also suggests a highly organised and mature karst aquifer. The linearity of the cave reflects its function in conveying water beneath an impermeable caprock from the more complex headward area, which presently lies beneath 100-300m of perennial glacier ice.

The cave is now a relic feature, having been abandoned for 700,000 years. It is thus not considered a karst feature in the functional definition given above. However, several sequences of flooding, filling and rejuventaion have occurred since its initial abandonment. Ford and Schroeder (1983) interpret extensive silt-clay rythmites as representing obstruction of springs by ice advances in the Castleguard Valley, while subsequent dissection and collapse are assumed to represent "interglacial" reactivation.

Vadose invasion shafts become more common towards the headward complex of the cave, beneath the glacier ice. These shafts frequently penetrate the cave, but appear discordant, passing directly through to greater depths. The walls and ledges of these shafts are clean washed and may be littered with erratics from the overlying Arctomys formation, although they are dry in winter (when the cave may be safely visited), but clearly very active in summer. The erratics indicate some connection to the surface above the Pika Formation.

The relative warmth of the cave in winter induces a convective draught (Wigley and Brown 1976) to pass up through the cave from the entrance to the shafts, and into inpenetrable passages (Atkinson et al.1983). In summer, this airflow is reversed, but is periodically occluded when floodwaters fill parts of the entrance passages.

Apart from the alluvial deposits in the cave, a diamicton (possibly glacial till) occurs in the Headward Complex. If it is till, then it is in marked contrast to the clear, intruded glacial ice which terminates a nearby passage (Ford et al.1976 p223).

The temperature profile of the cave warms to 3 degrees celsius 2.5 km beyond the entrance zone (Ford et al.1976). Beneath the glacier ice, temperatures decline and reach zero once more at the ice blockage (Atkinson et al.1983, Worthington 1983 pers. comm.). There are abundant drip waters in this inner section, many of which are producing speleothems. The conventional models for speleothem growth cannot be applied beneath glacier ice, and the mechanism inducing precipitation has excited some speculation (Atkinson 1979, 1981, 1983, Dreybrodt 1982, Smart 1981). Atkinson (1979) initially postulated evaporation as a significant process. Subsequent measurements negated this hypothesis, but also showed insufficient partial pressure of carbon dioxide to cause calcite precipitation. Atkinson (1981, 1983) has since suggested that precipitation is induced by dissolution of gypsum or pyrite by waters previously saturated with dolomite. Dreybrodt proposes a temperature effect which Atkinson (1981) regards as unimportant.

2.3.4. HYDROLOGY

The climate of the Columbia Icefield region is poorly documented. Meteorological stations at Sunwapta (50km) and Parker Ridge (10km) provide scant winter data, but the nearest established stations are Jasper (106km), Lake Louise (103km) and Banff (153km).

In general, mean daily temperatures above freezing occur from May to September inclusive, and rain may be expected during this time. The great altitudinal range of the area implies a strong variation in temperature and precipitation with elevation. Snow may fall above the treeline throughout the year.

Ford et al.(1976) estimated mean annual temperatures by interpolation from neighbouring stations. The calculated lapse rate of 1.0 degree celsius per 120m gave a mean annual temperature of 0 degrees at 1600m (the level of the springs in the Castleguard valley) and -7 degrees at 2500m (the highest ice-free karst area).

The winters of 1978-1979 saw 433mm and 407mm water equivalent snowfall at Parker Ridge. Both winters produced relatively low snowfalls at established sites.) The relationship between these data and snowpack status at the initiation of spring runoff is unknown. Meteorological data were collected on Castleguard Meadows and the Benches in the summers of 1979 and 1980, and these are presented in Chapter Four.

Surface runoff occurs from May-October with peak flows in June-August, depending on catchment altitude and ice-snow status. Overnight temperatures below zero have a significant effect in reducing streamflow. More sustained flows may occur in rivers drawing on groundwater.

Surface streams drain the South, Southeast and East Castleguard Glaciers and Terrace Mountain Glacier. In spring, surface runoff drains the central Meadows area where till deposits are then wet and impermeable. In summer this area drains internally through constricted sinkholes. Catchment areas are therefore seasonally variable. The lower Meadows are occupied by extensive gravel fans which act as local aquifers. Small sinks occur in an adjoining alluviated area here, and the fans may also have some underground leakage. Although the south east Benches drain surficially in spring, the residual ice in the upper catchment drains underground in summer.

The upper Meadows and the south Benches are karst areas; all runoff disappears down ponors. The upper Meadows are snow free from mid-July and sinking streams are fed only by springs or storm runoff. The south Benches are a proglacial area, so that sinking streams are active all summer. However, the ablation is brief both daily and seasonally because of the high altitude and limited storage. In summer, sinking streams engulf less than 2 cubic metres per second over the entire Castleguard area.

There are numerous springs which may be classified as follows:

1. Springs flowing from the Eldon-Stephen contact.

2. Springs in the upper Cathedral Formation at the level of Castleguard cave.

3. Springs in the lower Cathedral Formation in the floor of the Castleguard Valley.

4. Springs draining unconsolidated alluvial and glacial deposits.

The Eldon-Stephen springs are mostly found along the eastern edge of the Meadows. A few on the lower western margin flow only during snowmelt on the terraces above.

The few springs in the upper Cathedral Formation occur at the lip of the Meadows, 300m above the Castleguard River. The Red Spring lies some 30m below the entrance to Castleguard Cave. It is unusual in being a perennial spring. The Cave entrance acts as an intermittent spring, disgorging substantial floods on hot days in summer (Ford 1971a, Thompson 1976) The floods rise from a permanently flooded shaft inside the cave in an area known as Boon's Blunder. Floods also have to rise up a further 8m shaft before discharging from the entrance.

Springs in the Castleguard Valley are visually dominated by the Big Spring which is the largest of a group of 5-6 springs hanging 5-40m above the valley floor and discharging from a front of 300m. These springs combine into a turbulent stream which immediately joins the Castleguard River. Freeman (1925) estimated the springs to have a flow of 1,500 cubic feet per second (43 cubic metres per second), but Ford (1971a) established a more reasonable figure of 9-11 cubic metres per second. Further exploration of the Castleguard Valley revealed at least 80 springs extending 3.5km downvalley from the Big Spring.

Water sinking in the upper Meadows was traced to the Big Spring (Ford 1971a). However, the imbalance in the observed groundwater budget led Ford to suggest the base of the Columbia Icefield as the most probable source for the majority of the Big Spring Water. This was supported by the relative clarity of the water when compared to the turbid Castleguard River into which it flowed. Furthermore, the waters are saturated with reference to calcite at a partial pressure of carbon dioxide far below ambient atmospheric at any possible source altitude (Ford 1971). This is characteristic of glacial and nival meltwaters (Cogley 1972, Ek 1964, 1966, Miotke 1974). A large moulin (approx. 100m x 75m) engulfing a substantial quantity of water has been recently reported from the Columbia Icefield, almost directly overlying the headward zone of Castleguard Cave (B.Gadd, pers. comm., 1981). It has never been closely examined, and its hydrogeological significance is unknown.

Ford suggested that water from the Columbia Icefield travelled through an active analogue of Castleguard cave, graded to a lower valley floor. The active system was referred to as Castleguard II in contrast to the relic cave, Castleguard I. The traced link from the upper Meadows to the springs was called Castleguard III. The springs cease flowing in winter, but, like sinkholes in the area, are not explorable for any distance.

Flow from unconsolidated materials is quantitatively unimportant, because of the apparently low permeability of till in the area.

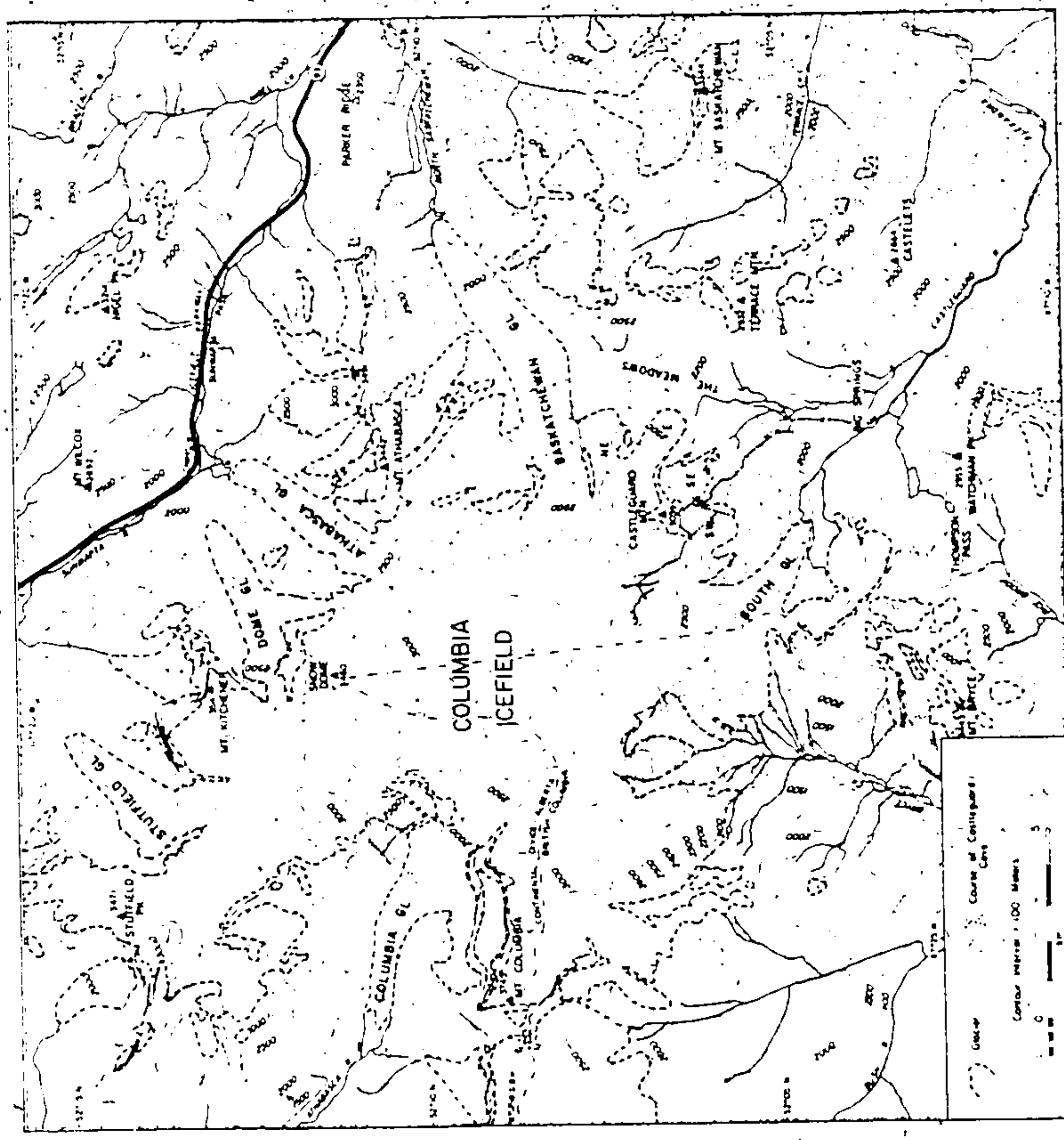


Figure 2.1. Topography of the Castleguard Area (Ford 1983)

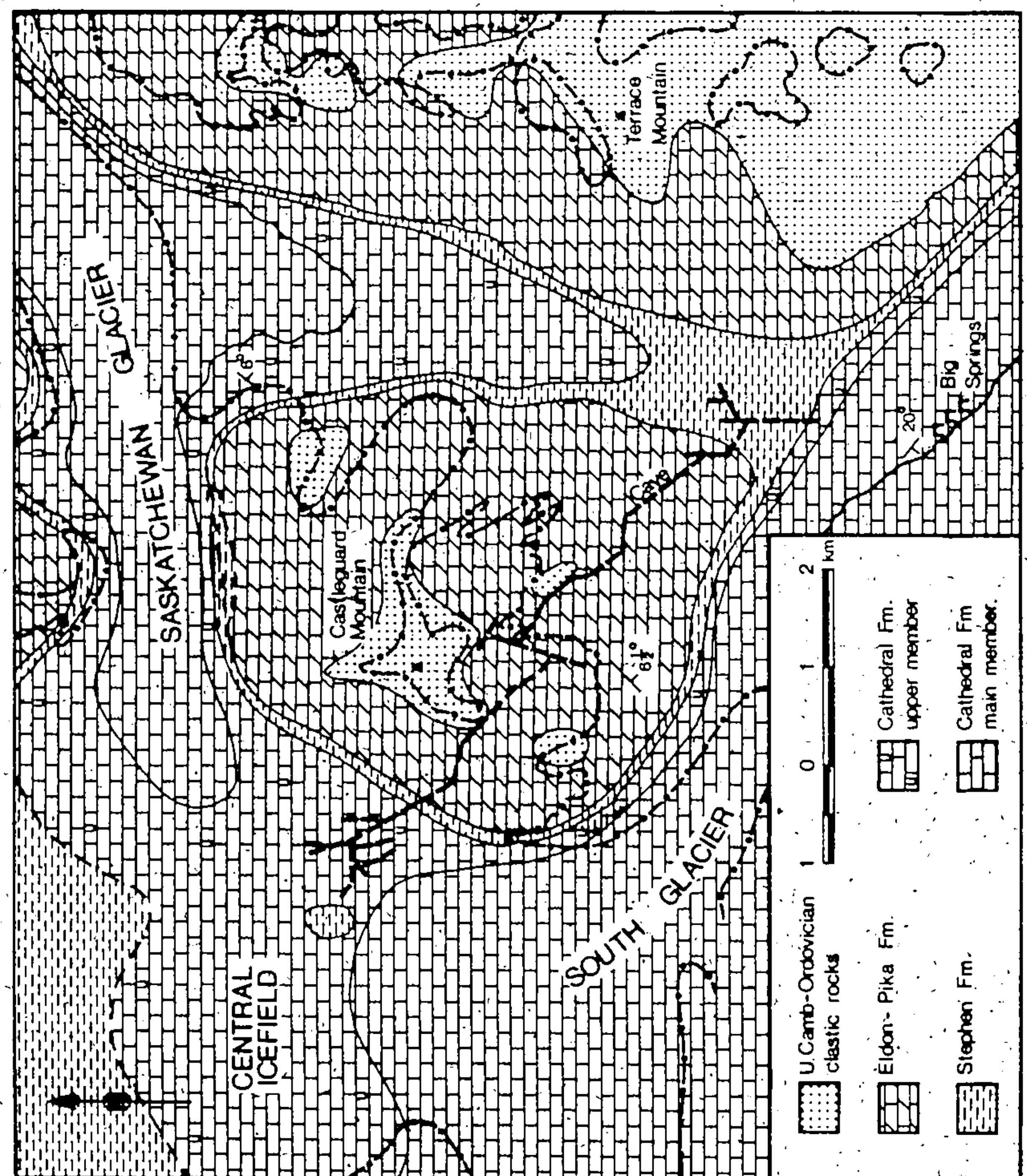


Figure 2.2. Geology of the Castleguard Area (Ford 1983)

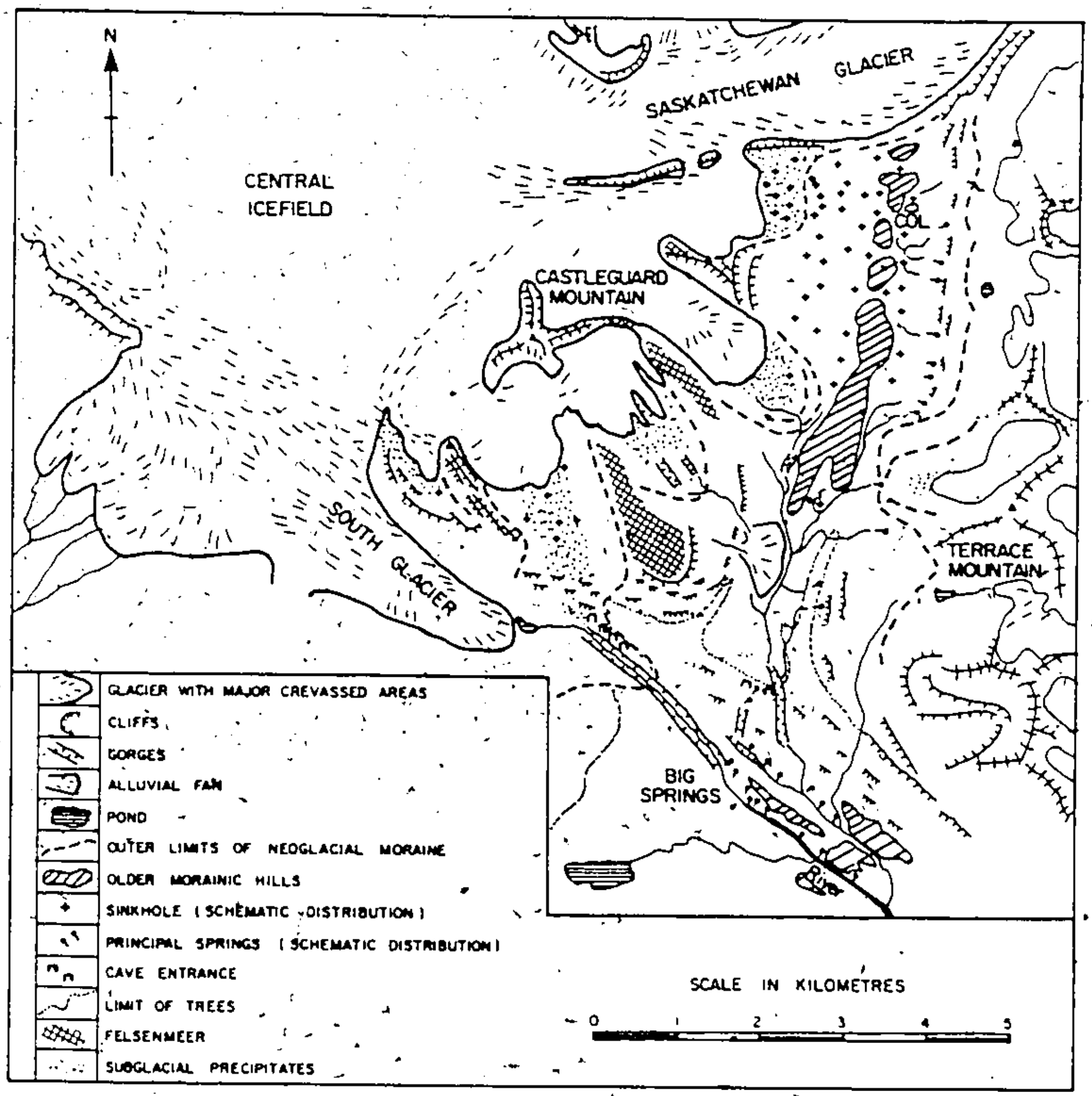


Figure 2.3. Geomorphology of the Castleguard Area (Ford 1983)

CHAPTER THREE

THE FIELD PROGRAM:

APPROACH, DESIGN AND IMPLEMENTAION

3.1. APPROACH

3.1.1. THE APPROACH TO KARST HYDROLOGY

Conventional hydrologic investigations are usually concerned with the behaviour of water within a selected catchment that is topographically defined. In karst areas groundwater is of great importance, and the unpredictable pattern of groundwater flow means that the catchment is not immediately apparent. The principal task of the karst hydrologist is thus to define the catchment. This is usually undertaken by an arduous program of tracer delineation (e.g.Atkinson 1977, P.L.Smart 1977), although quite apart from the physical and logistic limitations of point-to-point tracing, results may be ambiguous and dificult to interpret. An impressive example of such a study is from Kentucky (Quinlan 1982), where water level data from over 1700 wells and results from over 400 point-to-point traces were combined to define 28 catchments. However, even here divides were not absolute: dye might go into two "catchments" from a single injection point, or the boundaries might shift with changes in discharge.

In areas where the geological structure is complex, or the primary permeability of the limestone small, the water table has been rejected as a useful concept (e.g.Drew 1966, Lattman and Parizek 1964). As a result karst groundwater flow is difficult to model theoretically without considerable empirical information.

Attempts to go further and define water budgets for karst areas have met with limited success (e.g.Atkinson and Drew 1974, Chemin et al.1974, Markova 1970). Thus karst hydrology is often more concerned with basic definition than with strict resource evaluation.

3.1.2. AQUIFER MODELS AND THEIR IDENTIFICATION

1) Flow Analysis

The overall character of a karst aquifer is reflected in the behaviour of the discharge through springs which integrate the behaviour of the whole basin. Spingflow is amenable to quantitative analysis, but the model applied depends on the a priori conceptualisation of the aquifer.

White (1969) defined two basic flow types: diffuse flow and conduit flow, which Atkinson (1977) attempted to quantify for the Cheddar Catchment, Mendip, UK. White (1977) went on to suggest that the recession constant "a" might be used to characterise the aquifer in:

$$Q(t) = Q(0) \exp -a\ t$$

where $Q(0)$ is discharge at the start of recession

$Q(t)$ is discharge after time t

This approach may not always prove useful, because compound recessions may occur (e.g.Torbarov 1976), or exponential recessions may not be discovered, or be rather brief (e.g.Wildberger 1981). The former case is conventionally considered to result from the drainage of different storage media, but interpretation of recession behaviour has sometimes proven difficult.

A variety of chemical parameters may be used to characterise the history of spring water (e.g.Drake and Harmon 1973, Jacobson and Langmuir 1970, Schuster and White 1971, 1972, Wigley et al.1973). These authors all assumed a diffuse flow-conduit flow model. The effect of preconceived notions like this is illustrated by the interpretation of oxygen isotope data from the Pyrenees (that of Eberentz 1976) by Fontes (1981), who assumed altitudinal provenance as an explanation. Interpretaion of further data from the same system led a group of karst hydrologists to reject the altitudinal provenance hypothesis, and to give an explanation purely in terms of karst hydrology (Bakalowicz et al.1974, Bakalowicz and Mangin 1980).

Drawing on considerable data and experience the latter group have proposed a sophisticated and flexible model for the karst aquifer (Mangin 1973, 1975, Bakalowicz and Mangin 1980). The aquifer definition is that proposed in Chapter Two. It is the systeme annexe and the epikarstic aquifer which are the crucial differences between the French model and the diffuse flow-conduit flow model of English speaking countries. Diffuse flow is held to be the steady drainage of the systemes annexes, rather than drainage of fracture and pore storage. The complex relationship between the conduits and adjacent or distant, proximal or distal annexes accounts for many of the peculiarities and non-linearities observed in the discharge from karst aquifers.

In part the national differences in the modelling of karst aquifers reflects actual differences in the characteristic karst of different nations. However, the quantity and quality of data on which the French model is based has not been equalled in North America. It must be recognised that the level of explanation can only match the available data. More detailed data allow more sophisticated models.

A major criticism of Mangin's model is its rather broad generality. However, this is essential in order to encompass not only physical variability between particular cases, but the evolutionary nature of the karst aquifer. The internal evolution of the karst aquifer may be dramatically disrupted by quite independant geomorphic evolution of the surface environment. In this respect, a history of glaciation can have a major impact on the nature of an aquifer.

Magin (1975) attempted to develop a classification system for karst aquifers based on the recession behaviour and the dynamic storage compared to total annual flow. Unfortunately, the parameters are scale and climate sensitive, and require abundant data. A rational alternative is hard to conceive, however.

A possible method has been shown by Bakalowicz (1977) and Bakalowicz and Mangin (1980). The frequency distribution of chemical parameters in karst springflow has been found to be generally polymodal compared to that from diffuse and fracture aquifers. This shows that interpretaions based on the coefficient of variation (e.g.Shuster and White 1971, Ternan 1972) may be somewhat misleading when applied to such distributions, although the karst data are generally more dispersed than those from other aquifers.

An approach less demanding on data is the cross-correlation of inputs to the aquifer with spring discharge. Ashton (1966) developed the method initially, assuming certain relations between rainfall and the chemistry of karst water. Unfortunately, excessive interpretations of these supposed relations have been made (e.g.Christopher 1980). The hydrochemical identity of storm water is now recognised to be considerably more complex than initially thought (e.g.Friederich and Smart 1983, Miller 1981, Walling and Foster 1975).

A more robust procedure is the identification of the discharge response of springs to storm events. In some karst aquifers a sinking stream provides a simple input series. Where simple-throughput conditions hold, statistical analysis of the input-output series yields information on the system (e.g.Brown 1972). Williams (1977), perhaps drawing on earlier work (eg Tate 1879) describes an elegant use of artificaial flood pulses from a hydroelectric dam. Furthermore, this is the only known case where digital filtering of the output series was used to clarify the signal. In the present work the compound runoff response to both melt and rainfall also demanded digital filtering.

Unfortunately, influent streams may be only a small component of the total storm input (e.g.7-25% for the Mendip Hills, Atkinson and Drew 1974), and rainfall itself must be taken as the forcing function. This presents problems in identifying the areal effective rainfall, and becomes yet more complex where mixed rain-snowfall occurs. Dreiss (1982), expanding on earlier work by Knisel (1972), expended considerable effort in defining effective precipitation so that numerical instabilities could be avoided and catchment areas identified for particular springs. Mangin (1981a,b) did not recognise a problem, and used the cross spectral analysis of rainfall and springflow to identify water balances with apparent success (1981 pers. comm.).

2) Tracers

The application of hydrologic tracers to influent water may yield further information on the karst aquifer (Atkinson et al.1973, Brown 1972, Mangin 1975). Recently, repeated tracing has shown that the behaviour of an injected tracer is strongly discharge dependant (C.C.Smart and Ford 1981, P.L.Smart 1981, Stanton and Smart 1981).

Combined flood pulse and dye tracing provide additional information in the celerity between the impulse and the tracer (e.g.Atkinson et al.1973, P.L.Smart and Hodge 1980). Experiments on rivers (eg Buchanan 1966, Glover and Johnson 1974) have related dye cloud-flood pulse celerity to channel form.

Relatively little work has focussed on the nature of the breakthrough curve in water tracing (ie. the time-concentration or time-mass curves of tracer passage at a particular point). The understanding of this curve is the objective of dispersion theory in rivers, which has many practical applications (e.g.Belataos 1980, Brady and Johnson 1981, Fischer 1968, Taylor 1954). Recently the study of dispersion in groundwater has become important (e.g.Pickens and Lennox 1976, Meyer et al.1981, Smith and Schwartz 1980, Schwartz 1977), but applications in carbonate aquifers are rare (eg Grove and Betem 1971). Behrens et al.(1975) studied the dispersion of dye passing through the conduit system of a glacier, and interpreted the breakthrough curve as demonstrating open channel flow. The major constraint on this method has been technological. Until recently, sensitive, inexpensive, high frequency, real time techniques were not available for use in tracer investigations.

3.1.3. SUMMARY OF APPROACH

This highly condensed review of methodology in karst hydrology has demonstrated the effort which has gone into the understanding of the karst aquifer. However, two main points do arise: first, that the principal difficulty in karst hydrology is catchment definition, a trivial problem in more conventional studies; second, there is a danger of letting preconceived notions colour the interpretation placed on data. As a result, it is advisable that a foundation of empirical data is available before more general statements are made. The experimental design for the present work is correspondingly designed to construct a basis for the definition and description of what appeared to be a remarkably dynamic karst aquifer. To what extent this behaviour reflected the impact of glacierisation or glaciation was unknown, and constitutes a principal objective of the study. However, there seems little purpose in developing untestable hypotheses in the absence of a firm definition of the aquifer in question.

3.2. RESEARCH DESIGN AND IMPLEMENTATION

3.2.1. INTRODUCTION

The Cas[illegible]rd karst is the most extensively documented in C[illegible] yet previous knowledge of the active karst groundwater system was somewhat less than that presented in section 2.3. At the commencement of this study, the system was conceived as:

1. Castleguard III: the link between the upper Meadows and the Big Spring with a flow though time of about 3 days.

2. Castleguard II: the connection hypothesised between the Columbia Icefield and the Big Springs.

3. Sinking streams on the south Benches of unknown destination.

4. Floods of unknown provenance emerging from the Cave entrance for up to 19 days.

5. One small spring 30m below the Cave entrance, and another 500m away to the south east at the same level.

6. Various vadose inlets passing through Castleguard Cave, which was inferred to pass completely under the Icefield, because of the existence of the cave wind.

7. The Big Spring and the Cave were known to be dry during the late winter. The hydrochemistry was dominated by meltwaters.

The two problems of interest are:

1) to improve the definition of the karst catchment, and

2) to document aquifer behaviour, and to gain some understanding of the nature of the inaccessible aquifer, in particular with reference to the effect of glacier ice.

To this end field investigations took place in July 1978 (9days), July-October 1979 (94 days), and

June-September 1980 (107 days). In addition, a single dye trace was made in August 1981, and two April expeditions to Castleguard Cave took place in 1979 and 1980 for 10 and 20 days respectively.

3.2.2. THE CATCHMENT

Catchment definition was attempted by dye tracing accessible sinking streams on the Benches and in the upper Meadows, and in one case from a marginal moulin on the Saskatchewan Glacier. The absence of any discrete streams and the inhospitable nature of the Columbia Icefield (it is highly crevassed) prevented more extensive tracing. In addition, only a single assistant was available, and even the unglacierised catchment rather extensive, and logistically demanding.

The discharge per unit glacierised area (the "specific runoff") was estimated for several fine summer days for the Peyto Glacier, an I.H.D. Glacier 70 km south of Castleguard (Young 1977b). This was compared to results from other instrumented sites in the Rockies (Mathews 1964), the Purcell Mountains (Weirich, pers comm.), and the Coast Ranges (Mokievsky-Zubok 1973). These data were fairly consistent, and were compared to results from the East Castleguard Glacier, and the South Glacier. The total discharge from the karst catchment may then be crudely related to an equivalent catchment area.

Such a crude estimate may be evaluated in the context of regional geology and topography. Radio echo sounding data (Waddington and Jones 1977) were combined with the seismic results of Meier (1960) to construct a map of the subglacial bedrock topography. The surveyed position of the ice blockage in the Cave provided a further datum, and the entire headward complex a minimum elevation for the glacier sole. It was assumed that icefalls represented relative increases in slope, and that extraglacial contour lines could be reasonably extrapolated beneath the ice.

If water in different environments has distinct characteristics in its chemical composition and discharge variability, then these properties can be sought in the spring water, to identify contributions by waters of different provenance. Carbonate chemistry is not necessarily a good indicator, and more conservative characteristics such as the isotopic composition are more useful, although they cannot be determined in the field.

In 1978 deuterium samples were obtained during a brief reconnaisance. Samples of rainfall, snow, ice and meltwater were collected from around Castleguard Mountain and down the Saskatchewan Glacier. The Meadows Creek was sampled over several days near its source, and the cave springs were sampled during a cave flood.

The data demonstrated the expected distinction between meltwater and precipitation and a crudely diurnal behaviour of the melt stream. These results encouraged a major effort in 1979 producing more than 600 samples, many from regular sampling of the Big Spring. Unfortunately, facilities for analysing these samples were not available and they deteriorated in storage.

Samples for carbonate chemical analysis were taken in 1979, regularly from the Big Spring and irregularly elsewhere. Samples were analysed in the field for calcium, total hardness and occasionally for bicarbonate and pH. 1979 sampling showed little variation at the Big Spring, but considerable variation at the Cave Springs. The remote location of the latter springs (some two kilometres from base) meant that sampling frequency was inadequate. Continuous conductivity was attempted in 1980 to overcome this, but the data obtained was transitory and somewhat unreliable. The water chemistry was otherwise abandoned in 1980, because of lack of clear variability, and instability resulting from low partial pressures of carbon dioxide and very fine suspended sediment. The very low chemical load of many of the waters also gave the field analyses poor precision. Instead emphasis shifted to the hydrological and dye tracing program.

3.2.3. HYDROLOGY

The hydrological program demanded measurement of total springflow and inflow to the aquifer. Major inputs were inaccessible, therefore an attempt was made to define an analogue input to the aquifer. This was pursued at two levels: meteorological measurements (temperature and rainfall) were made at high altitude, and the discharge of the East and South Glacier melt streams was recorded. The meteorological stations were located on the South Benches and in the central Meadows in 1979. In 1980 the Meadows station was moved to the upper Meadows into the karst area to be more relevant to recharge and because 1979 data indicated less rainfall further up the Meadows. A rainfall record was also obtained from base camp at the Big Spring.

The gauging stations on the Castleguard River and the Meadows Creek were maintained in 1979 and 1980. Additional input information was obtained in 1980 by gauging sinking streams on the Benches and in the upper Meadows.

The output from the karst system was initially conceived as the Big Spring and the Cave Springs. The Big

Spring emerges at the head of a 40m cascade, and stage was recorded from the plunge pool in 1979 and 1980. Other springs nearby were included in the rating of this recorder, because of gauging difficulties, and it was implicitly assumed that a consistent relationship would hold between the Big Spring and the others. In August 1980, the Big Spring dried up and a recorder was installed downstream to measure flow from the remaining springs. This site would have been preferable at higher flows, but was hydraulically and geometrically unstable (Fig 3.1).

Exploration of the Castleguard Valley in 1979 revealed numerous other springs downvalley of the Big Spring and along the course of the tributary Meadows Creek. The highest elevation springs were apparently abandoned, followed by a level of intermittent springs, some lying on the divide between the Meadows Creek and the Castleguard River. The lowest level springs were often heavily alluviated and some marked by quicksands. The Artesian Spring rises in a spectacular boil besides the Castleguard River some 3.5 km downstream from the Big Spring. Water quality and temperature of these springs suggested a provenance similar to the Big Spring water. Gauging in 1979 showed springflow to contribute 3.5 cubic metres per second over a 1.5 km reach of the Meadows Creek, and in 1980 two recorders were employed spanning this reach to measure collective springflow (Fig.3.2).

The summer of 1980 was somewhat cold and wet (in marked contrast to 1979), and the small high altitude sinking streams dried up. As a consequence, an adjustment was made to the field program, in order to document the recession. Redundant high altitude recorders were moved to three valley locations, the Castleguard River Spring (#1), and the Artesian Spring and Creek. It was not possible to calibrate any of these recorders due to their brief operation and remote setting.

The Cave Springs were collectively gauged 1km downstream and 200m lower at a convenient location (Fig.3.3). The perennial Red Spring was overrun when the Cave flooded and could not be separately gauged. The apparently sequential flooding of the nearby Forest Spring and the cave was investigated by placing a stage recorder on the former spring in 1979. The sequence was confirmed and the recorder was unnecessary in 1980. The overall gauging of the Cave Springs was complicated by the two orders of magnitude difference between Red Spring discharge and the Cave Floods. The cave floods were considered of greater importance and as a result the resolution and error in the Red Spring discharge record are rather poor.

Total springflow could not be measured due to the number and distribution of springs, and the magnitude of total discharge. The downvalley extent of the aquifer was initially unclear, but the southerly structural dip brings the Stephen Formation into the valley floor some 5km downstream of the Big Spring, where it may well provide a confining bed. No obvious springs were found beyond the Artesian Spring.

Figures 3.4 and 3.5 show the distribution of instrumentation in 1979 and 1980 respectively. Tables 3.1-3.3 describe the locations of hydrometeorological instrumentation and the period of operation. Not all gauging sites were successfully rated, because of inaccessibility, unstable section, or brief tenure, nor was all the information furnished by an instrument necessarily digitised and calibrated; this depended on the importance and reliability of the data. Therefore the stated period of operation for the instrument does not necessarily imply a corresponding production of usable data.

3.2.4. DYE TRACING

Interrelations between hydrologic and tracer behaviour may contribute information on the nature of the aquifer. Dye tracing was one of the few aspects of the present work that was amenable to replication. Rather than relying on a single trace, multiple traces from a single point were made under different flow conditions. Furthermore, because spring waters were found to have low natural background, high resolution of the fluorescent dyes was possible. Rapid flushing of the dye allowed freqently repeated traces, and gave well defined breakthrough curves. Militating against this was the complexity of the system and the flow regime, the relative insignificance of known inputs, the logistical difficulty of working in an alpine environment, and the unpredictability of the weather. The inability to measure aggregate discharge prevented estimation of total tracer recovery, an important parameter in water tracing (Brown 1972, P.L.Smart and Smith 1976). Nevertheless, the data were gathered with the intention of investigating variations in the breakthrough curves obtained from various springs under different discharge condtions.

Six dye traces were made in 1979, eight in 1980 and one in 1981. In 1979 direct water sampling took place only at the Big Spring, activated charcoal detectors being used elsewhere. In 1980 controlled discrete water sampling took place at a variety of springs, depending upon operational equipment. The trace of 1981 was from the Saskatchewan Glacier. Detectors were placed at the glacier snout, the Red Spring, and the Big Spring, and occasional grab samples taken.

Nine traces were made from "Met Sink" on the south Benches, and two nearby sinks ("Midway" and "Main"). "Frost Pot", a sink fed from a perennial snowpack lying outside the neoglacial limits was also traced twice. These sinks are all in the Eldon-Pika Formation. Two traces were made from "Dye Sink" in the Col Karst. This is in the Cathedral Limestone, but is fed from springs perched on the Stephen Formation.

Sites of injection and sampling are listed in table 3.4 and illustrated in figures 3.4 and 3.5. The actual dye traces are summarised in table 3.5. The distribution of charcoal detectors is not described, because they were so numerous and changed frequently, but irregularly.

3.2.5. ADDITIONAL STUDIES

Less sustained studies of temperature and turbidity were also made at the Big Spring. Temperature was measured semi-continuously for a few days in 1979 before instrument failure, and over several days in 1980 during which time the Big Spring ceased to flow.

The turbidity of the Big Spring is remarkably low compared to the nearby Castleguard River. In the early summer waters were slightly turbid, as they were in September. However, during peak flow the water was very clear. To test these observations and to see if more subtle changes were also taking place, continuous flow nephelometry was attempted when power was available and the fluorometer not otherwise in use.

Appendix A summarises the techniques employed in the field, and the methods of installation, calibration and digitisation.

Reference	Description	Variable	On	Off	Digit	Grid
1979						
1. Benches:	300m from Glacier Bare Rock	Temp	20/7	20/9	Yes	831709
		Rain(c)	13/7	20/9	Yes	
		Rain(t)	13/7	20/9		
2. Meadows:	1km below E.Glac. Open parkland	Temp	20/7	20/9	No	856710
		Rain(c)	12/7	20/9	Yes	
		Rain(t)	12/7	20/9		
3. Upper Meadows:	3km NNE 2. Alpine tundra	Rain(t)	18/8	20/9		868738
1980						
1. Benches:	see 1979	Temp	13/7	24/9	Yes	831709
		Rain(c)	13/7	19/9	Some	
		Rain(t)	13/7	24/9		
2. Meadows:	see 1979	Rain(t)	12/7	19/9		856710
3. Upper Meadows:	see 1979	Temp	12/7	19/9	Some	868738
		Rain(c)	12/7	19/9	Yes	
		Rain(t)	12/7	19/9		
4. Big Spring:	100m SSW of spring, gravel flat	Rain(c)	11/7	30/9	Yes	852675
		Rain(t)	11/7	30/9		

Key: Temp : Temperature, Cassela Thermograph
Rain(c): AES tipping bucket raingauge
Rain(t): Standard 5" unshielded collecting raingauge
Digit : Data digitised
Grid : 6 figure grid reference, sheet 83C/3

Table 3.1. Meteorological Stations: 1979-1980.

Ref.	Name / Description	On	Off	Rated	Digit	Grid
2.	Meadows Creek(I): Small gorge 1km below E.Glacier	19/7	15/9	Yes	Yes	856710
4.	Forest Spring: Intermittent spring below Cave near Red Spring	29/7	20/9	Part	No	852688
5.	Cave Stream: Aggregate Cave Springs 1.5km below cave	19/7	21/9	Yes	Yes	851682
6.	Castleguard River: Near 7 at Canyon	18/7	29/9	Part	No	850675
7.	Big Spring: Plunge pool beneath spring	17/7	28/9	Yes	Yes	852676

Note: Grid references are to NTS sheet 83C/3

Table 3.2. Water level recorders 1979

Ref. Name	Description	On	Off	Rated	Digit	Grid
1. Polje Sink:	Stephen Spring flowing into flooded polje, upper Meadows	16/7	26/8	Part	Part	870737
2. Meadows Creek (I):	see 1979	13/7	29/8	Yes	Yes	856710
3. Benches Creek:	Small proglacial stream in open joint above sink	19/7	21/9	No	Part	831709
5. Cave Stream:	see 1979	17/6	30/8	Yes	Part	851682
6. Castleguard River:	see 1979	17/6	29/9	Yes	Yes	850675
7. Big Spring:	see 1979	17/6	29/9	Yes	Yes	852676
8. Lower Big Spring:	collective springs 50m above Castleguard R.	29/8	29/9	Yes	Yes	853673
9. Castleguard River Spring (1)	first major spring on left bank 1km below 7	9/9	23/9	No	No	858670
10. Meadows Ck. (II):	400 m below junction with Cave Stream, at log Bridge	20/7	22/9	Yes	Yes	859674
11. Meadows Ck. (III):	falls at junction with Castleguard River	21/7	22/9	Yes	Yes	868661
12. Artesian Spring:	Spring on trail at entry of small creek	31/8	16/9	No.	No	877654
13. Artesian Creek:	3.5 km below Big Spring small creek crossed on log	16/9	22/9	No	No	877655

Table 3.3. Water Level Recorders 1980.

Ref	Name	Description	Discharge	Geology	Grid
a.	Glacier Sink:	Marginal stream sink below Cliff upglacier of Meadows	0.01	Ice/Cath.	853754
b.	Dye Sink:	sink in choked shaft on broken rock shoulder, upper Meadows	0.001	Cath (U)	872738
c.	Midway Sink:	Choked shaft on high benches 150m from glacier	0-0.001	E/P	831711
d.	Met Sink:	Active proglacial sink in narrow karst fracture, 30m from glacier	0.01	E/P	831710
e.	Main Sink:	As (d), but 200m from glacier	0.01	E/P	831708
f.	Frost Pot:	Snowmelt sink in felsenmeer on ridge between benches-Meadows	0.03	E/P	836710

Key : Geology

E/P :Eldon Pika Formation
Cath(U):Upper Cathedral Formation
Cath(L):Lower Cathedral Formation
Cath(?):Presumably Cathedral (obscured)
Uncons.:Unconsolidated Rock

Grid: refers to NTS 83C/3

Table 3.4a. Dye Injection Sites

Ref	Name	Description	Discharge	Geology	Grid
g.	Cave Stream:	e of Cave Springs	0.03-6.0	Cath(U)	851682
h.	Big Spring:	Major spring	0.0-6.0	Cath(L)	852676
i.	Watchman Spring:	On right of Castleguard River opposite (h)	.01-0.1	Cath(L)	850674
j.	Gravel Spring:	in boulderlag below (h)	0.01	Uncons.	852675
k.	Castleguard Spring (I):	Left Bank Cg River 1 km below (h)	1.0	Cath(?)	858670
l.	Tangle Springs:	Meadows Creek at junction with Tangle Creek	0.1-1.0	Cath(L)	864671
m.	Artesian Spring:	Left bank Cg. River on trail by small creek	0.5-1.0	Cath(?)	877654
n.	Saskatchewan Glacier:	Proglacial river 3km below snout	3.0	Uncons.	933792

Key : Geology
E/P :Eldon Pika Formation
Cath(U):Upper Cathedral Formation
Cath(L):Lower Cathedral Formation
Cath(?):Presumably Cathedral (obscured)
Uncons.:Unconsolidated Rock

Grid: refers to NTS 83C/3

Table 3.4b. Dye Sampling Sites.

Ref	Injection	Date	Time	Dye	Mass(g)	Sampling
DS0CS	Dye Sink(b)	18/8/79	16.00	FLU	974	(Red Spring)
MS1BS	Met Sink(d)	24/8/79	14.00	RWT	1282	Big Spring
MS2BS	Met Sink(d)	5/9/79	15.30	Flu	1620	Big Spring
MS3BS	Met Sink(d)	9/9/79	16.00	RWT	752	Big Spring
MS4BS	Met Sink(d)	13/9/79	17.10	RWT	710	Big Spring
MS5BS	Met Sink(d)	21/9/79	18.00	RWT	893	Big Spring
DS6CS	Dye Sink(b)	20/7/80	20.30	RWT	1026	Cave Stream
DS6BS						Big Spring
MS7BS	Met Sink(d)	27/7/80	16.00	LFF	2511	Big Spring
MS7AS						Artesian Sp
FP8BS	Frost Pot(f)	27/7/80	19.15	RWT	1169	Big Spring
FP8AS						Artesian Sp
MS9BS	Met Sink(d)	2/8/80	15.20	RWT	1410	Big Spring
MS9CG						Cg (I) Sp
MS9TC						Tangle Sp
MS9AS						Artesian Sp
MS10BS	Met Sink(d)	15/8/80	15.30	RWT	908	Big Spring
MS10CG						Cg (I) Sp
MS10TC						Tangle Sp
MS10AS						Artesian Sp
FP11BS	Frost Pot(f)	22/8/80	16.05	LFF	4727	Big Spring
FP11WS						Watchman Sp
FP11TC						Tangle Sp
MD12BS	Midway Sink(c)	25/8/80	17.00	RWT	892	Big Spring
MD12TC						Tangle Sp
MD12AS						Artesian Sp
MA13BS	Main Sink(e)	10/9/80	16.10	RWT	898	Gravel Sp
MA13CG						Cg. (I) Sp
MA13TC						Tangle Sp
GS14	Glacier Sink(a)	3/8/81	13.00	RWT	2800	Cave St
						Big Spring
						Sask Gl

Dyes: RWT: Rhodamine WT
FLU: Fluorescein
LFF: Lissamine FF

Table 3.5 Dye traces - injection and sampling sites.

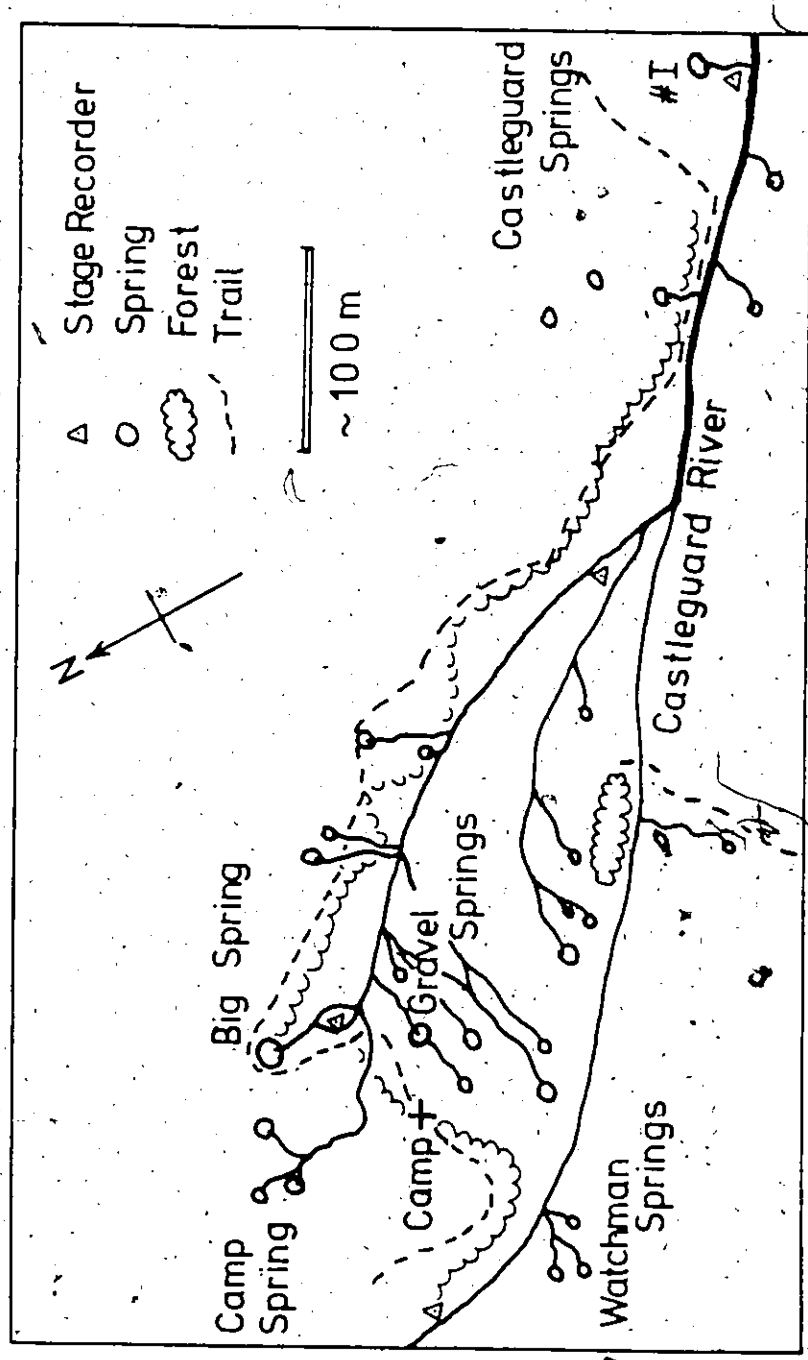


Figure 3.1. Sketch Map of the Big Springs Area

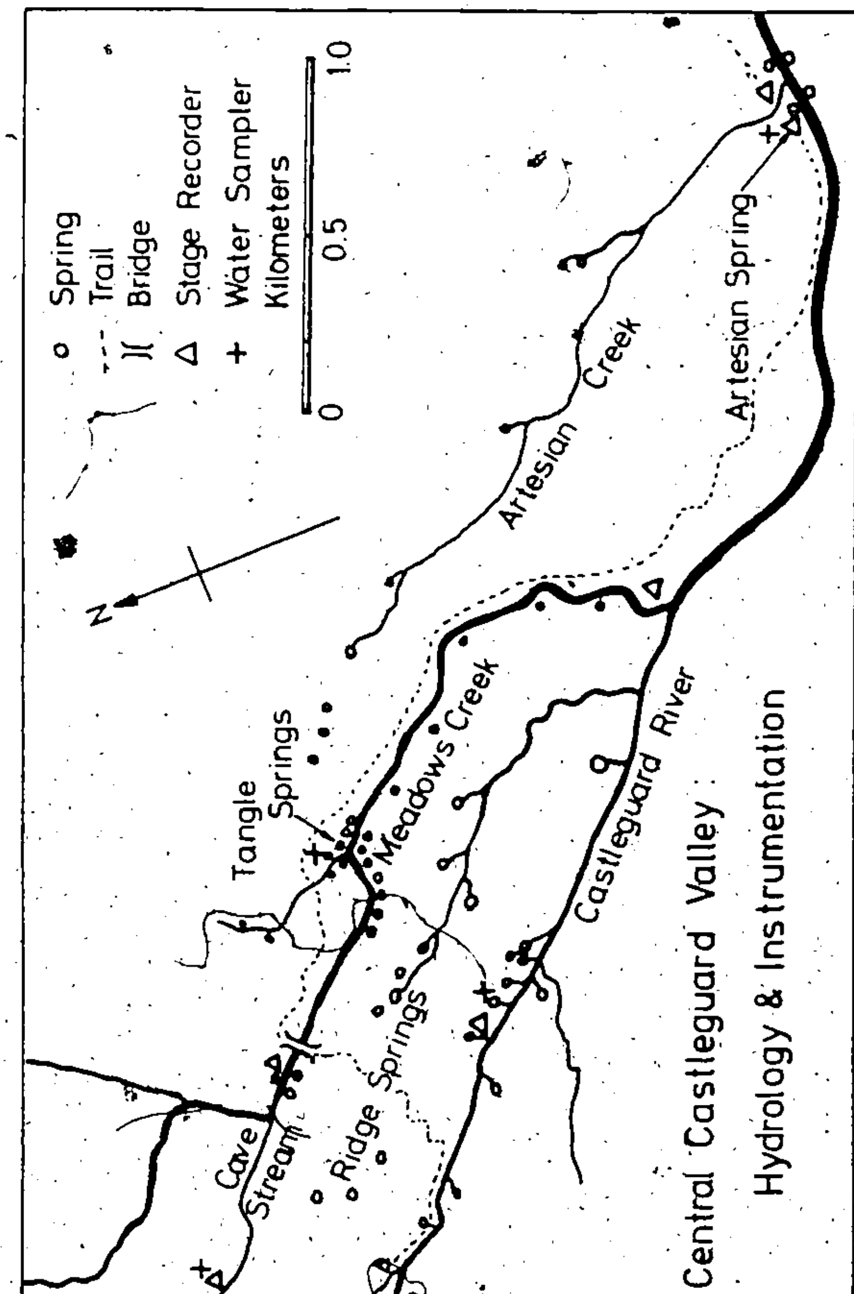


Figure 3.2. Sketch Map of the Central Castleguard Valley

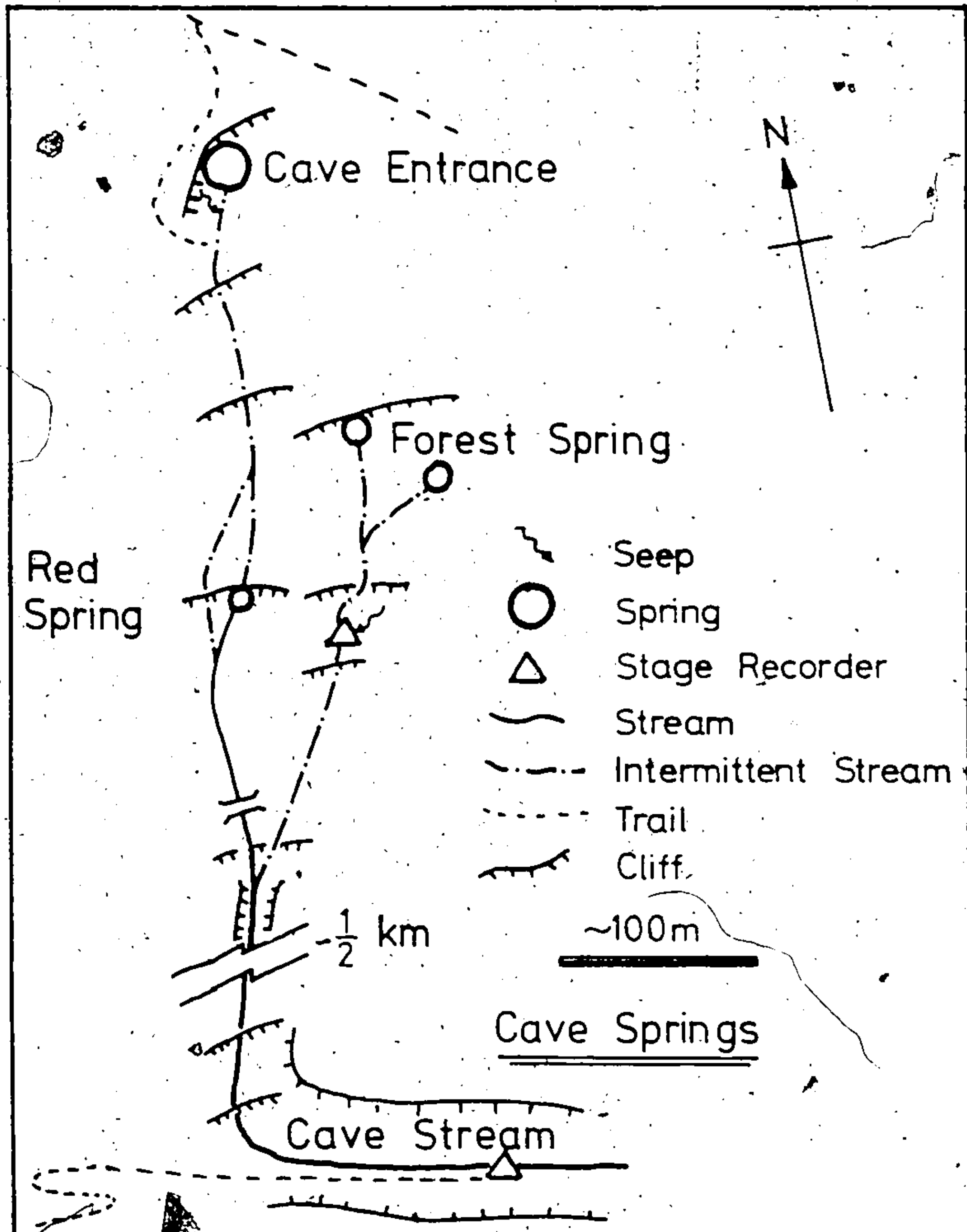


Figure 3.3. Sketch Map of the Headwaters of the Cave Stream

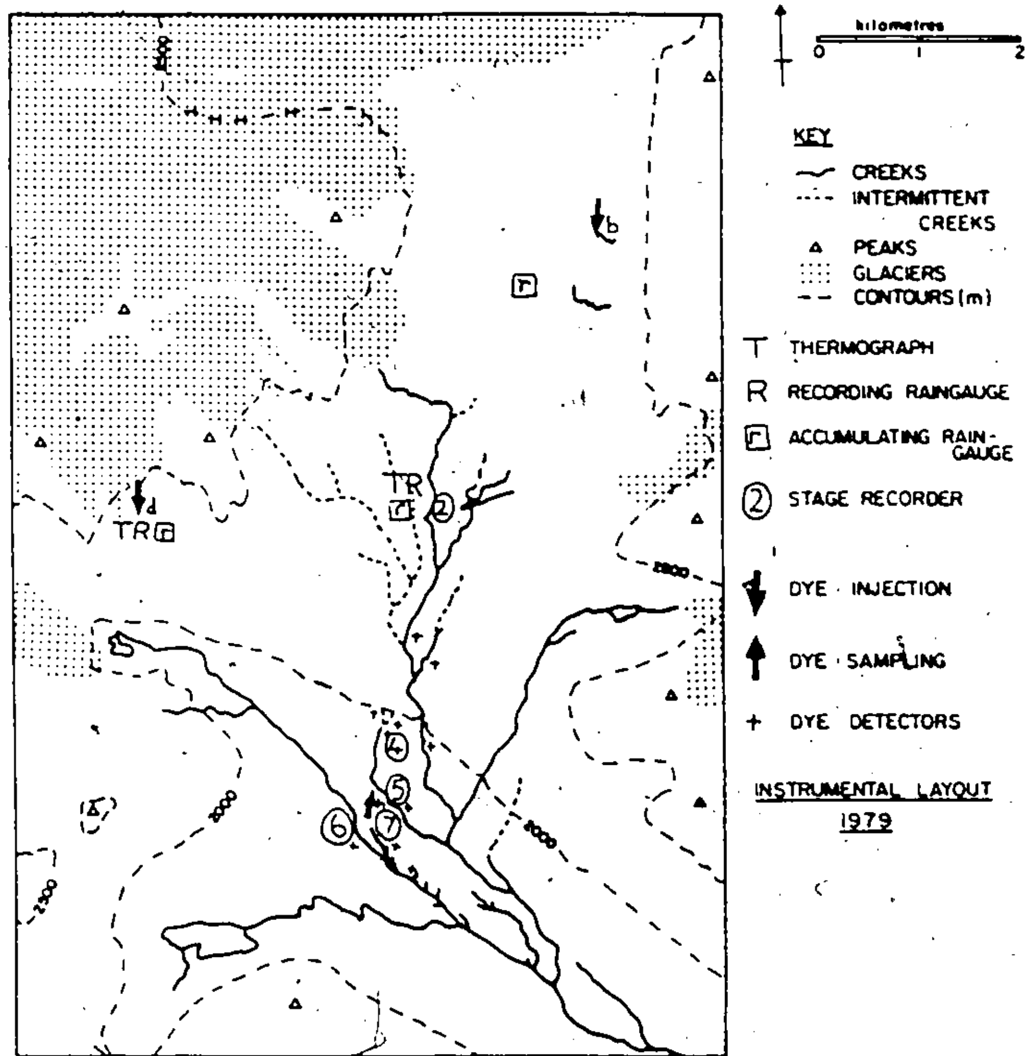


Figure 3.4. Distribution of instrumentation and dye tracing locations 1979

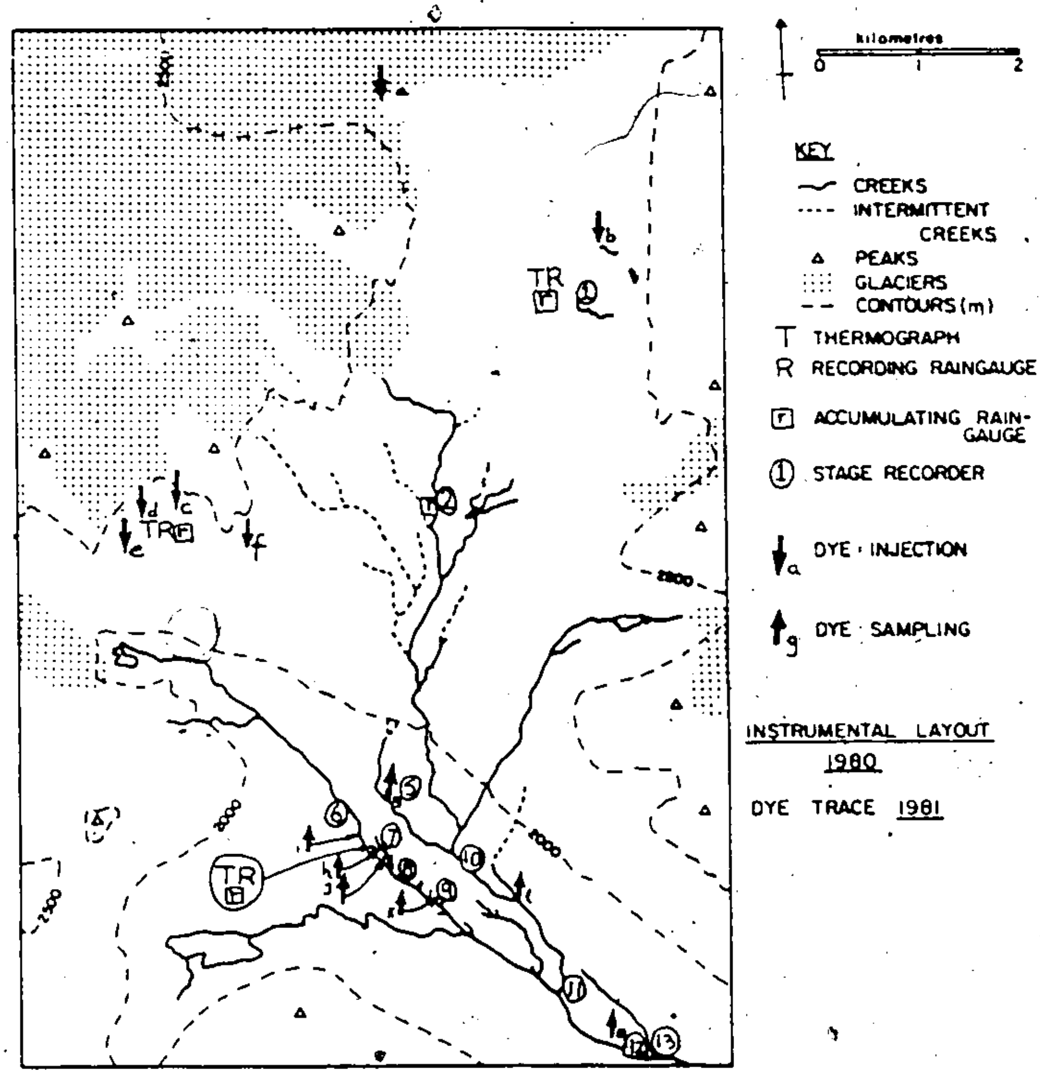


Figure 3.5. Distribution of instrumentation and dye tracing locations 1979

CHAPTER FOUR

HYDROLOGY

This chapter presents results from investigations described in Chapter Three, and is a preliminary attempt at understanding the nature of the karst aquifer. The Cave Springs are considered in full detail in this chapter, while the more abundant information concerning the Valley Springs is distributed over this and the next two chapters.

4.1 CATCHMENT DELINEATION

The dye tracing results are shown in Figure 4.1.. All dye traces were positive to every spring sampled in the floor of the Castleguard Valley, although the Cave Springs were only positively traced from the Upper Meadows. The Saskatchewan Glacier trace was detected only at the Big Spring by grab samples, and no dye was detected at the Cave Springs, nor at the glacier snout during the 3 days after injection. Figures 4.2 and 4.3 show the underground routes drawn with respect to known subsurface geology and morphology.

The relative strength of the traces varied in detail, discussion of which is deferred to Chapter 5. However, the traces from the upper Meadows are discussed below in Section 4.3.3.(3). The positive result from the Meadows to the Big Springs confirms that reported by Ford (1971a). No dye was detected in cave floods per se.

Specific runoff calculations are shown in Table 4.1. The Castleguard Glaciers are obviously releasing relatively little water to stream flow. Although it is tempting to attribute the loss to groundwater flow, these figures may merely reflect characteristics of local mass balance. They are similar to the results from the nearby Athabasca Glacier.

The total springflow in the Castleguard Valley may reach an estimated 20 cubic metres per second, comprising a measured peakflow of 14 cubic metres per second, plus a conservative estimate of 6 cubic metres per second ungauged springflow. Extreme specific runoff values of 57.6mm/d (Peyto Glacier) and 13.5mm/d (East Glacier) may be used to determine the equivalent catchment area for the springs, giving corresponding estimates of 30 sq.km. and 128 sq.km. The former estimate alone would encompass all the glaciers within a 3 km radius of Castleguard Mountain. The topographically defined catchments of the Saskatchewan Glacier and upper South Glaciers extend 7 km north west of

Castleguard Mountain totalling an area of approximately 50 sq.km. This constitutes 17% of the total area of the Columbia Icefield, while the estimate of 128 sq.km. constitutes 42%.

The low values of runoff estimated for the Castleguard Glaciers imply an unreasonably large catchment area for the spring. This is particularly so when it is recognised that much of the catchment is in the accumulation zone where relatively low ablation rates are expected. It is therefore implied that the Castleguard Glaciers are indeed losing a part of their melt to groundwater.

During minimum flow periods, the Castleguard River showed diurnal oscillations while the springs did not. This suggests that at this time aquifer recharge areas lay above the altitude of continuous freezing and that it is only from the lower altitude ablation zone of the South Glacier that these pulses were derived.

The bedrock topography of the Icefield is shown in Figure 4.4. The landforms of alpine glaciation are retained beneath the broad cover of glacial ice. Two closed depressions are present, one described by Meier (1960) from the Saskatchewan Glacier and the other inferred to occupy the cirque head of the South Glacier. Such features are common in glaciated terrain in the Canadian Rockies (Ford 1979), and may be drained underground. If the upper catchment of the South Glacier is assumed to be captured by the closed depression, a reduced catchment area of 7.5 sqare kilometres for the Castleguard River yields a 37.8 mm/d specific runoff; a more reasonable figure, although still low for what is now dominantly an ablation zone.

If the bedrock depressions are assumed to drain their bedrock topographic catchments, a drainage basin extending well into the Columbia Icefield is implied, corresponding closely to the 50 square kilometre surface topographic catchment described above. If this is so then the meltwater discharge of the Saskatchewan Glacier should be considerably underfit. No figures are available to test this hypothesis.

The possibilites discussed above rest on geologic concordance with the putative catchment. Direct extrapolation of the geologic structure demonstrates the closed depressions to be in the karstic Cathedral Formation (Fig.2.2). The anticlinal axis along the Saskatchewan Glacier trough has not been considered in the extrapolation, but may have provided a fracture pattern conducive the development of karst ponors. Much of the remainder of the catchment lies above the inferred level of the Stephen Formation. If this is impermeable, then subglacial conduits will respect the bedrock topography, draining meltwaters to the closed depressions. The Castleguard Valley lies down-dip from both depressions, creating conditions suitable for karst drainage.

The catchment proposed by Ford (1971a) is supported by these findings, although considerably greater detail has been added. All meltwaters produced in the bedrock catchments of the subglacial depressions may be contributing to groundwater. To this may be added the contribution of proglacial streams around Castleguard Mountain, and snowmelt and precipitation occurring in subaerial karst areas.

4.2. INFLOW TO THE AQUIFER

4.2.1. METEOROLOGY

Hourly air temperature records for 1979 and 1980 at the south Benches (2500m asl.) are presented in Figure 4.5. The records are markedly diurnal, the normal daily cycle of temperature being moderated by katabatic winds at the proglacial site. The two series show a strong contrast between the 1979 and 1980 seasons. In part this is because of the different screens used in the two seasons. However, Table 4.2 shows that the variation in mean monthly temperatures between years during the period of measurement at Castleguard is supported by the monthly data from the nearest established stations (Atmospheric Environment Service), although their distance prevents a direct comparison.

A similar contrast applies to the precipitation data for 1979 and 1980 (Tables 4.3, 4.4a and Figure 4.6). Comparison of mean daily rainfall between sites at Castleguard produces consistent results, with the heaviest rainfall at the Benches, followed by the lower Meadows, followed by the upper Meadows. Furthermore, the ratios between comparable sites for 1979-1980 are consistent. This suggests a relative stability in the heterogeneous distribution of rainfall. However, the time distribution of individual events depends on storm proyenance, and occasionally falls are limited to a single raingauge.

The character of individual storms varied from intense falls of over 6mm/h to prolonged light drizzle. The temperature response to rainstorms varied from dramatic falls of up to 6 degrees Celsius, to rises of a degree or so during cold conditions. The precipitation snowline was correspondingly unstable. The broader time distribution of late season storms in part results from the delayed melt of snow trapped in the raingauge.

4.2.2 RUNOFF

Both rainfall and ablation contribute to runoff, therefore the quantity and distribution of precipitation and surface snow and ice are fundamental controls on runoff. Unfortunatley, precipitation falling as snow does not contribute directly to runoff, and the exact proportion of a particular storm falling as snow is unknown.

Ablation generated runoff is difficult to model, because of ripening processes in the snow pack, the variable contribution from ephemeral snow, and the difference in the ablation response of snow and ice bodies. Although Power and Young (1979) have shown that detailed measurements are not essential in modelling runoff from glacierised basins, the crudely defined catchment at Castleguard makes any such attempt unrealistic. Runoff is more strongly correlated to temperature than to radiation (Jensen and Lang 1973, Lang 1972, Ostrem 1972), which is rather surprising, because physically the sensible heat transfer has litte influence on ablation. However, Braithwaite (1981) has demonstrated that this correlation results from the similarity in variability of temperature and runoff compared to net radiation. A clear example of the effect of sensible heat was shown by the Castleguard River from the 5-7/9 1980 when, responding to rising air temperatures, discharge increased through the early hours without solar radiation or rainfall.

Figure 4.7 illustrates runoff generation from the East Glacier and the larger and lower South Glacier (note the different discharge scales). The Castleguard River which drains the South Glacier is less flashy, presumably because the larger catchment has a broader response time. Flow is seen to be decreased drastically by temperatures falling below zero degrees C, especially in Meadows Creek. This in part results from the greater altitude of the latter (some 340m above the South Glacier snout), as is seen when diurnal events are present in the Castleguard River, but not in Meadows Creek.

Spring rain-on-snow events (Meier and Tangborn 1961) were not covered by the observation season. The characteristic flashy summer response to rainfall is seen for much of July and August, and a gradual dampening caused by snowfall appears in late August to September. The hourly rainfall and discharge data give a zero rainfall response time for Meadows Creek and a delay of approximately one hour for the Castleguard River. This is in marked contrast to the results for Peyto Glacier (Goodison 1972), where characteristic response times are in days, depending on the condition of the glacier. Unfortunately, Goodison, like most other workers has considered only daily data, precluding higher frequency responses. Gudmundsson (1970), Lang (1973) and Östrem (1972) all define rainfall responses in terms of days, again using daily data. Lang describes both positive and negative responses to precipitation for particular glaciers. The seasonal variability in behaviour described above was handled by subdividing the year into "seasons" of relatively homogeneous behaviour. Power and Young (1979) route melt and rainfall through unit hydrographs assigned to altitudinal zones with distinct snowpack characteristics. Once more response is in the order of days, except for rain falling on the snow-free ablation zone.

A possible explanation of the rapid rainfall response at Castleguard lies in the anomalous nature of the presumed catchments. Krimmel et al.(1972) have used tracers to demonstrate that meltwater velocities from a melting firn surface are from 6-27 m/h compared to 266-2450 m/h for meltwater streams on bare ice. This difference is of the same order as that described above. The low specific runoff figures obtained from the South and East Glaciers (Table 4.1) led to the hypothesis that runoff from the upper portion of these glaciers was captured underground. The remainder of the glacial catchment is largely bare ice during the summer and much of the extraglacial area is unvegetated. Such a catchment would have a rapid response to rainfall, compared to an entire glacier catchment. A similar argument may also be applied to the East Glacier. This provides further evidence for the nature of catchments hypothesised in Section 4.1.

No attempt was made to estimate contribution from the annual snow pack. In July 1979 and 1980, seasonal snow was present in the upper Meadows where it was concentrated in the karst depressions. This snow disappeared relatively suddenly during warm weather.

4.2.3 SINKING STREAMS

The plan to approximate inflow to the aquifer by measurement of Meadows Creek and the Castleguard River is somewhat undermined by the above discussion. Most of the contributing area of the aquifer is inferred to lie in the glacier accumulation zone, while the studied rivers appear to be dominated by runoff from the ablation zone. Only runoff from the exposed ice of the Saskatchewan Glacier component of the catchment might be approximated by these data.

Stage measurements of two sinking streams were made in 1980; a proglacial stream on the South Benches and a stream in the upper Meadows fed by springs at the Eldon-Stephen contact and draining into a karst pond. Neither location was entirely successful in yielding sustained data and neither was fully gauged.

Peak daily discharge for the proglacial stream was crudely approximated by considering an equivalent rectangular weir (Gregory and Walling 1973 p139). Overnight flows always fell to zero, except when rainstorms occurred. This sharp diurnal impulse is again probably uncharacteristic of much of the aquifer catchment, but provides a measurement relevant to tracer injections into similar streams in the vicinity. The overall pattern of proglacial recharge (Figure 4.8) correlates closely to temperature. However, the total cessation at night suggests either that overnight temperatures always reached zero degrees at the ice surface, or else that some part of the flow was being lost to groundwater upstream. The former hypothesis is contrary to the temperature data from the station (Figure 4.5), and the well-developed karst of the South Benches makes the latter hypothesis more acceptable.

The spring-fed stream flowing into "Polje Sink" showed strong diurnal variations peaking from 23.00 to 4.00h. This particular stream entered a small lake which would damp out such fluctuations, although a few other such streams entered the aquifer directly. The volume of flow involved is trivial, but the behaviour of this stream is discussed below under spring flow (Section 4.3.2. Figure

4.17).

4.3 SPRINGFLOW

Here three spring groups are considered: (i) those in the floor of the Castleguard Valley, (ii) springs rising at the contact between the Eldon and Stephen Formations, draining the aquifer beneath the Terrace Mountain Range, and (iii) the Cave Springs which are at an elevation intermediate between the two former groups.

There is a large quantity of information on the Valley Springs, so that discussion is extended over the following two chapters. The more limited data concerning the Cave Springs is, however, synthesised in this Chapter.

4.3.1 THE VALLEY SPRINGS

This section first presents information on the curious flow behaviour of the Big springs, for which a possible explanation is then developed. The springs are considered to be elements of a vertical hierarchy in which springflow discharge behaviour is diagnostic of spring status. The valley springs are inspected for this behaviour, and subsequently some aspects of the valley aquifer are presented. Finally, a few days of continuous turbidity data from the Big Spring are inspected and interpreted.

1) Spring Behaviour

The springs in the floor of the Castleguard Valley (Figures 3.1 and 3.2) lie along the axes of the two rivers: Meadows Creek and the Castleguard River. Total hardness of the Valley Springs was steady at around 25mg/l throughout the summer for all samples. Changes in sample turbidity, treatment and delay in analysis appeared to be the dominant cause of chemical variability (see Appendix A.1.3.). Dye tracing confirmed that the springs were all interconnected in a single system. The largest spring (the "Big Spring") furnished a remarkable hydrograph (Figure 4.9), in which the amplitude of diurnal oscillations reaches a maximum at intermediate flows. At peak discharge flow is characteristically steady over the day, but shows high frequency fluctuations of 2-5 minute period. At the hourly sampling interval employed for digitising these oscillations are strongly aliased and constitute "noise". At low flows, the Big Spring shows very rapid recession and an almost instantaneous rising stage. The seasonal behaviour of the spring relative to the meteorological variables and the Castleguard River is shown in Figure 4.10. The daily average flow reaches a plateau from which it falls off relatively rapidly compared to the Castleguard River. Precipitation response varies with the proportion of snowfall in a storm, the extent of snowcover, and thus with season.

2) The Overflow-Underflow Spring Hierarchy

The Castleguard data are remarkable because of the intensity and regularity of the hydrologic signal produced by snow and ice ablation, which demonstrate aquifer behaviour more clearly than do conventional rain-induced hydrographs. As a whole the Big Spring discharge shows no consistent relation to any supposed input signal or forcing function, and is highly non-linear in its behaviour. Whether this non-linearity is a product of the superimposition of glacier ice on the karst, or of some karst aquifer characteristic is unknown, but must be understood before any synthesis is attempted.

Similar behaviour has been reported from the Maligne Basin, some 150km north of Castleguard (Kruse 1981, Luckman 1980), where a karst lake is assumed to be controlling aquifer behaviour. There is no such lake at Castleguard, but is this a possible effect of the Icefield? Other information suggests not; the Penyghent karst resurgences in Yorkshire, United Kingdom, have similar peculiarities (Proudlove pers. comm.), and inspection of other published data (e.g.Bahrens et al.1981, Mangin 1975) suggests that the phenomenon may be fairly common, although less obvious on conventional hydrographs than is the case under the present diurnal regime, and usually obscured by the collective measurement of spring discharge. Aley (1963) has identified plateau-type hydrographs as "typical of a stream with impeded flow", an assertion which can be tested by studying tracer behaviour (Chapter 5). The remainder of this section attempts to establish an hypothesis to account for this curious and virtually unreported phenomenon.

The behaviour of an individual spring within an intimately interconnected group of outlets will depend upon its position within the vertical hierarchy, the size of its attendant conduit and the water levels within the aquifer. There is thus a division into the topmost active spring, the "overflow" spring ("trop plein" of Mangin 1975), and those springs at lower levels called "underflows". This distinction has long been implicitly recognised in karst hydrology. However, the identification has been based on observation of the sequence of flooding within a group of springs (e.g.Talour 1978, Tratman 1968), or by noting discontinuities in the daily discharge ratios between springs (Mangin 1975). In this case, the primary diagnostic criterion is the characteristic behaviour of any single spring. Furthermore, no inferences concerning the nature of the aquifer have been drawn from this recognition.

The Big Spring shows a hydrograph characteristic of both overflow and underflow states. When recharge of an aquifer exceeds the outflow rate, storage occurs and energy gradients change, increasing the output. In a conduit aquifer, the discharge in a single phreatic conduit increases with head. In a non-captive aquifer, the rising water levels will often encounter an alternative flow route which can function as an overflow. Up to this point the lower spring would increase in discharge according to the square root of head. Activation of the overflow changes the collective discharge capacity of the aquifer and greater inflow will increase overflow discharge at a greater marginal rate resulting in a relatively small increase of head in the aquifer. As a result, discharge of underflow springs is less responsive to changes in recharge. As discharge falls, springs will be progressively abandoned, and the topmost currently active spring will exhibit the variability characteristic of an overflow.

The behaviour of springs in such a hierarchy depends intimately on the internal configuration of the aquifer: conduit size, network topology and storage volume. The response to unsteady flow is frequency dependent. High frequency inputs will characteristically be lost if the flood is small in total volume compared to the volume of the conduit space which needs to be filled before changing the effective configuration of the network.

A simple approach to modelling a karst aquifer can demonstrate the principle of the spring hierarchy outlined above. This is attempted in Chapter 6, along with a crude analysis of the response to varying flows.

Anticipating somewhat the model results, it is possible to summarise an overflow spring hydrograph as containing a high proportion of input variance, with a steep rising limb, and without an exponential recession. Once an overflow is active, the net flow resistance decreases so that a given increase in total discharge will induce a smaller rise in head. For this reason, the underflow springs will exhibit relatively little variance in discharge while an overflow is active, compared to their behaviour without the overflow.

3) Identification of Springs in the Hierarchy

In support of the above hypothesis, overflow and underflow spring behaviour may be seen in Figures 4.11 and 4.12. In 1979, warm weather brought the aquifer water level up to the point where the Big Spring became an underflow, with little variation in discharge, although very strong diurnal variations are still seen in the

Meadows Creek record of glacial melt. In 1980 the Big Spring did not become a "pure" underflow, although diurnal variation was maximised at intermediate flows between 1.5 and 3 cubic metres per second.

The Meadows Creek Springs record (Figure 4.12) is a composite record of net contributions from numerous springs over a 1.5km reach of the Meadows Creek (including a small surface stream) and as such might not be expected to have a well defined character. The early peak in the record corresponds to the cave flood. It represents either gauging error ,or direct, in phase, contribution from the same source as the cave flood. The brief duration of this high magnitude flow means that no rating data could be gathered and discharge is therfore calculated from an extrapolated rating curve. The flood is characteristic of an overflow event and matches the cave flood in form. It is unlikely that the Meadows Creek springs would contribute a signal similar to the cave floods emerging over 300m above, or that the small surface stream could contribute a matching flood. Therefore gauging error is accepted as the more likely explanation, although the Cave Stream does lose water in its bed and the recorded cave flood maximum is larger than the total flow downstream. This again probably represents gauging error at high flows as is reviewed in Appendix A.

Following this brief impulse, the Meadows Creek Springs show a very flat hydrograph characteristic of an underflow. High frequency oscillations represent contributions from the small surface stream and errors due to assuming instantaneous transmission over, and no storage within the gauged reach. Springs were observed discharging in Meadows Creek upstream of the upper gauging station at this time, so that it is not necessarily the Big Spring which is the controlling overflow in this case. Figure 4.13 shows this period. Note that Meadows Creek (#2) (the upper station) shows additional early morning pulses not seen in the Big Spring (nor upstream near the source of the stream). This is inferred to be spring water arriving from a source different to that of the Big Spring waters.

As total discharge falls, so the Meadows Springs commence oscillation and the Big Spring reaches its maximum diurnal variability (Figure 4.12). The recession for Meadows Creek is gradual, while the Big Spring dried up rapidly leaving only relatively minor springs in the group active. These springs in turn ceased to flow in strict order of vertical position. At this time gauging commenced on the Castleguard River Spring (1) 1 km downstream of Big Spring, and the Artesian Spring, the lowest spring in the Valley. These records were also characteristic of underflow springs, maintaining relatively steady flow until higher springs ceased to function.

As flow is restored in early September, the Meadows Creek recovers relatively rapidly. The main outlet to the Big Spring Group only starts to function when water levels in the acquifer have built up. The five sharp pulses in discharge at this time (Figure 4.14) all result from activation of the Big Spring outlet, which becomes a prevailing overflow.

In the specification of the spring hierarchy, there remains the high discharge overflow to be identified. Flood pulses at the Cave Springs some 300m above Big Spring are a possibility. They have a hydrograph characteristic of an overflow and there is a correlation between August (1979) cave floods and underflow behaviour in the Big Spring (Figure 4.11). However, close inspection shows this relationship to hold for neither the large, early 1979 floods, nor for the three days of flooding in 1980. Furthermore, the 300m head elevation would have a marked influence on discharge at the Big Spring. The hydraulic gradient implied is about 1:4.4 which is excessive in terms of the spring behaviour outlined above, and indeed for any known aquifer. However, the notion of a continuous, smoothly varying water table results from a static concept of water in the aquifer, which is inappropriate in the present dynamic system. The routing and exchange of water within a conduit network is influenced not only by relative elevations, but by the energy loss at junctions, blockages, and other discontinuities, so that overflows can be also be induced by "hydraulic damming" (e.g.Mohring et al.1983).

The rating of the Big Spring stage recorders presented major difficulties which are reviewed in Appendix A.3. The final rating curve contains a discontinuity resulting from initiation of flow in "Camp Stream", a small creek draining a series of springs 50m northwest of, and a few metres higher than the Big Springs (Figs. 3.1, A.2). Although no continuous record exists from this stream, general observation suggests that this is the main overflow to the Big Spring and to the Castleguard River Springs.

It is not known whether there is an overflow spring on Meadows Creek, as the most likely location lies in the floor of an inaccessible canyon. However, the total discharge leaving Castleguard Meadows in Meadows Creek is somewhat greater than that recorded 1.5 km downstream at the Meadows Creek (#2) gauging station. The springs observed in the bed of the creek may be members of a group of potential overflow springs, obvious only when actively discharging water. At other times, depending on conduit configuration, stream water may be able to enter the aquifer through these openings. Such reversing features are termed "estavelles" (e.g.Stanton 1982), and their presence further complicates aquifer behaviour. Insufficient hydrologic data is available to document their influence, although the dye tracing results may reflect their impact.

4) The Valley Aquifer

None of the Valley Springs is enterable so that direct inspection of the active aquifer is impossible. Only relatively simple models are therefore justified, given the available data. Such models are developed in the following chapters, based on interpretation of the hydrograph and tracer information, and observation of the relict aquifer in Castleguard Cave.

A crude estimate of the aquifer reserves supplying the springs can be made using recession analysis (e.g.Milanovich 1981). Taking the total springflow (Meadows Creek Springs and Big Springs) a brief "recession" occurred 30/8/80 to 4/9/80 (Figure 4.15). This was a period when diurnal fluctuation were least, and recharge can be assumed to be a minimum. No correction is made for ungauged springs. A calculated recession coeffecient of 0.086/d gives storage at the onset of the "recession" as 1.68 million cubic metres. The rapid drawdown rate is demonstrated by the "half-life" for the aquifer of about 8 days.

These figures, however, are only crude approximations. Furthermore, discussion in Chapter 6 will question the general theory of recession analysis of karst springs, and for this reason the estimation remains approximate.

5) Big Spring Turbidity

The water emerging from the Big Spring was always slightly turbid, but became clearer during peak flow in summer and showed obvious correlation to rain pulses during times of low flow. A variable source area concept with supraglacial melt as the clear component, and proglacial and extraglacial as the turbid may be used to explain this.

The turbidity record, obtained through nephelometry (Figure 4.16) is in abritrary units, because of the complexity of light scattering theory (e.g.Kerker 1969). Reflectance is strongly dependant on the size, shape and colour of particles, the refractive indices of particle and fluid, and the wavelength of light used. Thus reflected light is only directly proportional to concentration, while proportional to the cube of diameter and inversely proportional to the fourth power of wavelength. The illumination in this case was a standard "cool white" light, and so the data may be interpreted in terms of particle concentration or size.

There is an obvious correspondence of diurnal pulses in discharge and turbidity. The high turbidity of the 22-25/7 corresponds to the brief cave floods of 1980, but spread over a longer period. Response to rainfall is less clear, but a response with a lag of 6-8 hours might be present.

The raw nephelometric data also contained noise which was always positive, occurring as definte, but irregular pulses. The noise frequency appeared to be similar to that of the "surging" of the Big Spring. This observation suggests that there is an internal sediment reservoir which is a source and sink, highly responsive to discharge variations. When draining at low flow, the marked increases in spring turbidity appear to be unrelated to discharge or rainfall and may represent collapse and mobilisation of this local sediment store.

The delayed response to rainfall suggests turbid water arriving from a greater distance. The cave floods of 1980 were turbid events and their association with Big Spring turbidity is the only evidence for a physical link between Big Spring and the Cave Springs underflow. The broader base of the turbidity event is an underflow property. The non-specific nature of such tracers prevents any firm conclusion concerning provenance. However, it is suggested below that the early cave floods result from snowmelt in the upper Meadows. Cave floods thus have an association with events at the Big Spring. The only evidence for a similar link to the Meadows Creek is the out of phase pulses described earlier (Figure 4.13).

Although the data are somewhat crude, they may well be an indication of the kind of conditions which produced the extensive silt-clay laminates in the explored cave. Samples of suspended sediment collected on Millipore filters from the Big Spring bear at least a superficial resemblance to these deposits.

4.3.2 THE ELDON SPRINGS

Marking the top of the Stephen Formation along the eastern margin of Castleguard Meadows is a prominent spring line (Figure 4.1) which rises to the north with the Stephen Formation aquiclude. A few matching springs on the opposing flank of the Meadows are relatively small and flow only briefly. The southernmost springs coalesce into a channel which flows into the Meadows Creek. The central springs flow out onto the Meadows floor where a complex of channels exist. A portion of the flow in these channels may soak into groundwater. The northernmost springs rise some 50m above the Meadows floor, coalesce into streams and sink into the Upper Cathedral Formation.

One spring-fed stream in the upper Meadows feeds a large pond of about 4000 square metres. After being full for much of the summer, this drained under dry conditions to reveal a constricted soakaway and a seal of beige laminates overlain by less differentiated, ochorous mud. The stream feeding this lake was gauged in 1980, but the section was not fully rated, and only stage data are presented (Figure 4.17). Measured discharges were of a few litres per second; although marked flow peaks occurred around 24.00h and were never actually witnessed. There is clearly a considerable delay between melt in the Terrace Mountain Range and the spring response, which may result from a vadose flow system, or delay by replenishment of aquifer storage. The peaked response suggests a discrete rather than a diffuse recharge process.

Inspection of Figure 4.17 suggests that this is an overflow spring, as defined in the preceding section. The absence of other data makes it impossible to identify the corresponding underflow. Alternatives are any of the springs along the Meadows, or unknown springs flowing into Terrace Creek to the east of the range. The southern Meadows springs showed sustained flow, which is characteristic of underflows.

The total hardness of spring flow was between 50 and 60mg/l, at the northern springs (9 samples) and 68 and 79mg/l for the southern springs (2 samples), which suggests water approximately equilibrated to a standard atmosphere. These values are quite distinct from those obtained from the Valley Springs, and probably represent flow from an independent aquifer, with differences in residence time accounting for the slight difference in values from the two sites. While the aquifer can not be defined from the above data, it is clear that a conduit network exists within the Eldon Formation of the Terrace Mountain Range, and is presumably perched on the Stephen Formation.

4.3.3 THE CAVE SPRINGS

Emerging 300m above the floor of the Castleguard Valley, the Cave Springs include the Red Spring, the Forest Springs and Castleguard Cave entrance (Figure 3.3). In addition, the Cascade Spring discharges some 20 l/s from a 0.5m diameter conduit, and emerges with associated seeps on an unstable ledge in the canyon of Meadows Creek, some 500m downstream from the edge of the Meadows. Although part of the present group, its inaccessibility meant that it was little studied.

The behaviour of these springs is complex, being dominated by the cave floods which are periodic overflow events. The low flow behaviour is of an entirely different magnitude, but is of interest, as is the interaction between these two modes of flow.

1) Hydrology of the Red Spring

The Red Spring is perennial and responds to both rain and melt events. Figure 4.18 shows that this behaviour is extremely subdued. In fact, digitising to the nearest 0.005 inches was necessary in order to generate a realistically smooth series, and the data is consequently sensitive to the slightest recorder error. The rapid response to rain is probably derived from channel precipitation, while the subsequent broad pulse shows a peak delay of five to ten hours. Crude baseflow separation for the storms of 27/7 and 28/7 gives estimates of the total discharge due to rapid and to delayed response. Using the upper Meadows precipitation data, and dividing it into the total flood volume, the effective catchment areas can be calculated (Table 4.5). The rapid runoff area seems a reasonable estimate of channel and bare rock area upstream of the gauge, confirming channel precipitation to be its origin. The delayed response represents contribution from a very small catchment equivalent to runoff from a circular area 112m in diameter. The runoff volumes were crudely estimated from the discharge data by rectangular integration. Unfortunately, the recorder resolution was designed to pick up the 5 cubic metre per second cave floods, and was insufficient to justify more precise analysis of Red Spring discharge.

Beyond their basically diurnal similarity, there is only a weak association between Red Spring discharge and air temperature. Diurnal variations of discharge continue even when the small sinking streams on the upper Meadows are dry. The low amplitude of the diurnal variation (plus or minus 2% of mean daily flow) and its consistent magnitude (2-3 l/s) suggests that the diurnal component is supplementary to a relatively independant baseflow.

The continued flow of the Red Spring through the winter into late April suggests a reservoir of considerable size. Data on the winter recession are not available. However, a discharge of 40 l/s in late September, and an estimated 1 l/s in April may be taken and an exponential recession assumed for a reservoir without recharge. Reserves of .20 million cubic metres are estimated for the end of September and the reservoir has a half life of 40 days. The latter figure is in marked contrast to the 8.1 day half life estimated for the Valley aquifer and suggests a radically different supporting aquifer.

2) Cave Floods

At a larger scale, the cave floods are of interest because of both the threat they pose to speleologists in suddenly filling parts of the cave, and the anomally of such discharges emerging 300m above local base level. The floods represent an overflow (Figure 4.11) with an underflow probably routed to the valley springs. In this situation a rising water table is unlikely to be the cause of the overflow events. More likely is hydraulic damming caused by a constricted link to lower outlets. A conduit morphology in which this might occur is seen in the First Fissure of Castleguard Cave, where constricted shafts drain a larger, graded vadose canyon. Although the shafts are able to absorb moderate discharges, the large headloss at such junctions during floods limits flow into the shaft, and the canyon is re-activated as a flood overflow route. This emphasises that adherence to the water table model in conduit aquifers is likely to result in serious misinterpretation of aquifer behaviour. This type of flooding is unlikely to demonstrate the rigid hierarchic sequence characteristic of floods in a free-flow aquifer.

The Forest Spring is an intermittent spring lying some 30m below the cave entrance, and is the lowest outlet point for the cave floods. The discharge capacity of the spring is limited, however, and larger floods build up inside Castleguard Cave, eventually spilling from the cave mouth an hour or so after the Forest Spring commences flowing. When the cave floods, a marked rise is seen in the total flood hydrograph, but the Forest Spring maintains an extemely steady discharge of some 0.9 cubic metres per second. Therefore the Cave Springs comprise an overflow-underflow group in their own right. Throughout flood events the discharge of the Red Spring varies rather little, suggesting that it is fed by an independant underflow conduit. However, in the entrance complex of Castleguard Cave a small stream flows in April, presumably feeding the nearby Red Spring. This stream is adjacent to a flooded shaft which is the inferred source of the floods within the cave. Therefore the Cave Springs occupy a common conduit although their behaviour is so different, and the ultimate provenance of the Red Spring and flood waters are unknown.

When the cave was not flooding, the response of the Red Spring to rainfall was shown above to be relatively feeble. However, there is also a massive response only observed during periods of cave floods (Figure 4.19: see 25/8/79), suggesting a considerably enlarged catchment area at this time. Considering the latter example, subtraction of "expected" discharge based on neighbouring dry days (and relatively low flows) gave a total rain induced discharge of 42,000 cubic metres. 5.1mm of rain fell on the upper Meadows at this time, giving a catchment area of 8.3 sq.km., assuming 100% runoff. This area corresponds to the centripetally drained portion of the upper Meadows. However, the underflow conduit discharge has also to be maintained. Perhaps the runoff from glacier ice within the catchment balances this loss.

The first cave floods of 1979 and 1980 occurred before the summer runoff regime was established in the valley springs. In 1979, the floods were observed to be coincident with the disappearance of snow in the depressions of the upper Meadows, and were very turbid when compared to later floods. The relative turbidity of these floods suggests that at least a component was derived from extraglacial areas. Total discharge from these floods (minus Red Spring Discharge) was 777,000 cubic metres, equivalent to a snow pack .32m in depth, assuming a snow density of 300kg/cubic metre within the putative catchment area of 8.3 sq.km. This is presumably an underestimate, because the cave floods are an overflow and so some critical underflow discharge must be satisfied prior to their occurrence. The failure of heavy rains alone to induce cave floods during non-flood periods suggest that the underflow loss must be fairly substantial. Some indications of this contribution to the Valley Springs were noted above.

3) Tracing

Traces made from the Benches during cave floods did not appear at the cave, nor did the Saskatchewan Glacier trace which was made at low flow. However, a small sinking stream on the upper Meadows was traced twice to the Red and Cascade Springs, with a weak trace also to the Big Spring (Figure 4.20). This was during low flow conditions.

The average flow-through time for the trace was 30h, with a recovery of 0.01% at the Red Spring. There was a double pulse; the earlier and smaller arriving at rising stage and the later one at peak discharge. Charcoal detectors suggest that a similar quantity of dye passed out of the Cascade Spring. Some data were lost from the Big Spring, but detectors suggest that little dye was missed through this omission. In this case the Big Spring breakthrough curve is an attenuated mimic of that from the Red Spring, but with a two hour lead. Detectors on other valley springs suggest a similar recovery to that of the Big Spring.

The major problem with the Meadows-Red Spring trace is the loss of at least 90% of the dye. The loss can be to an unknown outlet or into storage, although there is clearly a relatively efficient conduit to both the Red and Big Springs. The residence time of water should be reflected by the water chemistry and these data are now considered.

4) Carbonate Water Chemistry

The carbonate chemistry of the Cave Springs shows substantial variation (Figure 4.21), in contrast to the steady values from the Valley Springs. Cave floods dilute the spring waters to a level comparable with fresh meltwater. After cessation of floods there is a period of recovery lasting through to April when total hardnesses in excess of 120mg/l have been measured (T.C.Atkinson pers.comm., the exact figures are in doubt). The detailed behaviour is more complex, however: the Red Spring continues to discharge hard water for some hours after initiation of floods. Similarly, the first floodwaters of the Forest Spring flood pulse are hard. The waters also take some time to recover following cessation of floods. Unfortunately, the remote location of the springs prevented more thorough sampling, but the general relationship to floods is seen in Figure 4.21.

The need for higher frequency sampling led to the continuous conductivity study of 1980. The relatively dilute character of the water gave a poor resolution on available equipment and an operational amplifier was necessary to increase sensitivity. Unfortunately, this introduced a non-linear response, which in combination with general drift means that the data can only be presented in arbitrary units (Figure 4.22), and are only ordinal.

The cave flood is seen to be the most significant dilution event. The lag time between opposing peaks and troughs of discharge and conductivity during cave floods is 1.5h. The conductivity recovery time from the cessation of the cave flood is 9h. The contrast reflects the differences in active discharge in relation to the volume of water stored in the conduit. Although it is tempting to estimate reservoir volume from these figures, the result will be strongly discharge dependent, and will include channel storage in the kilometre or so of channel between the springs and the recorder. The gradual evolution of water chemistry in response to floods shows that mixing is occurring for which a piston model is inadequate. Furthermore, the conductance signature of neither flood nor base flow water is established and linear mixing cannot be assumed for these data. An attempt is made below to model the system from isotopic data gathered from individual springs during a flood event.

During low flow when only the Red Spring is active, conductivity is less varied. Rainstorms induce dilution, both from channel precipitation and from the subsequent minor flood pulse. Diurnal discharge pulses are associated with lower conductivity water.

The evidence suggests a rapid passage of rainwater compared to the dye transit time. Rainfall is clearly routed through a more efficient system than that the dye followed. Similarly, the diurnal pulses are apparently synchronous with the dilution of the spring. The celerity is remarkable, suggesting negligible channel storage and a short channnel length. Unfortunately, the diurnal signature of the Red Spring was too poorly defined to justify cross-correlation with other time series such as temperature, to estimate pulse lag times.

The chemical data suggest that there is a storage volume between the springs and the locus of mixing of harder (baseflow) and softer (flood) waters. This pre-existing volume of harder water is rapidly flushed by flood events. In contrast, the relatively slow recovery of hardness after floods probably results from the slow replacement of this storage volume by baseflow. The longer term recovery in September probably has little to do with floods, but is a reflection of decreased recharge and longer residence time of the water.

The finite storage inferred by lagged hydrochemical response in floods is contradicted by the apparently 180 degree phase relation between discharge and conductivity series at low flow. Assuming these data to be approximately correct, the dilution processes in flood and at low flow apparently involve an independant conduit system.

5) Isotope Hydrology

The only absolute data available with a reasonable sampling frequency during a cave flood event are isotopic. Deuterium variations in the Red Spring and in a cave flood are shown in Figure 4.23. At left are samples from 20/7/78. The cave flood started at 17.35 and rapidly assumed a distinctively light isotopic composition (approximately -174 per mil), while the Red Spring is more variable. Unfortunately, continuous discharge data are not available from this time.

Although the Red Spring and flood waters are isotopically distinct, there is some mixing of the waters. This is seen in the slight dilution of the Red Spring during the flood and the initially heavy composition of the flood waters. The residual pool water (-174 per mil) and sedimentary evidence suggest that the cave had also flooded the previous day (22/7/78). The evolution of Red Spring water from its initially light composition early in the day may represent a recovery from dilution by this flood to more "normal" values of around -166 per mil.

The parallel behaviour of Red and Forest Springs is unexpected. Although distinctly different in character, they show a similar "resistance" to dilution. The samples from 20/7/78 support the close relationship. The Forest spring started to flow at approximately 16.00h, as the Red Spring became more dilute.

If the transition from preflood water to flood waters was delayed, but rapid, then a "Piston" model could be applied to the data, in which case a flooded discrete conduit or a well defined open channel might be inferred to be the conduit feeding the spring. The volume of water initially in this conduit could be estimated from the quantity of pre-flood water ejected.

However, the gradual transition suggests that progressive mixing is taking place before outflow. Although numerous conduit configurations can be invoked to account for the mixing, one of the simplest is a continous mixing model in which a finite reservoir of preflood waters is progressively mixed with and replaced by flood waters. This is an exponential mixing model in which the proportion of preflood water in spring discharge decreases exponentially with time, assuming steady discharge of flood water and no recharge of preflood water.

Insufficient data exist to test the validity of the model. However, exploration of the flood zone of the cave shows the entrance complex to be characterised by broad low bedding plane passages occupied by extensive breakdown and pools of water. Flooding of such passages would result in a progressive mixing. The assumed absence of preflood water recharge of the flood conduits during floods is supported by the relatively long term dilution of the Red Spring waters after cessation of flooding. However, at this time the Red Spring is often the only active spring and flood water residue from other conduits may be discharging this way. The model is only applied to rising stage and its results are only of comparative value.

The Red Spring is assumed to be initially at -166 per mil and mixing with -174 per mil water with a constant discharge of 50 l/s. The resulting hypothetical reservoir has an initial storage of 1,500 cubic metres and a half life of 5.5h. The apparently slow dilution of the Forest spring water suggests a yet larger reservoir. Assuming a steady 900 l/s discharge the two(!) data points suggest a reservoir of 21,500 cubic metres with a half life of 4.5h.

The cave flood rapidly rose to a steady discharge estimated at 0.5 cubic metres per second. Taking the first five flood deuterium values (except number two), a similar exponential model was fitted, assuming an ultimate deuterium concentration of -174 per mil. The heavy water reservoir in this case was estimated at 600 cubic metres with a half life of 14 minutes. The rapid transition between water types therefore suggests that a piston flow model is more appropriate in this case. The known conduit is indeed more open and of larger size than other passages, so that the dilution may be attributed to dispersive rather than physical mixing.

The data base for the above models is extremely sparse and the model is based on gross assumptions. However, it is the relative response and resistance of various conduits which is of concern, and in this respect the type of mixing model used is more significant than the parameters calculated.

The conduits behind the Red and Forest Springs are believed to join in the cave in the region of Boon's Blunder. In summer a cool draught issues from the Forest Spring when it is not in flood and is presumably an air flow from Castleguard Cave. The slow dilution (and recovery) of these springs in flood suggests that this spring is served by the sort of passage explored in the entrance complex of the cave: a wet, low bedding plane conduit, much obstructed by breakdown, and through which there are several flow routes. C.Smart (1981b) suggests that such passages may develop where almost saturated waters are involved.

6) Isotope Geography

The deuterium sampling program of 1979 established the basic isotope hydrogeography of the area around Castleguard Mountain (Figure 4.24). The results are complex. Winter snow produces very light meltwater. The altitude effect on Castleguard Mountain is the inverse of conventional results (e.g.Moser et al.1972), probably the result of relatively heavy spring and summer snowfalls. The most distinctive water was rainfall.

Glacier ice meltwaters appear to fall in the -160 to -170 per mil range. However, a few samples of locally derived meltwaters on the Saskatthewan Glacier have a progressive lightening with distance from the snout. These data show a linear relationship between distance and deuterium concentration, with a correlation coefficient of 0.984. Previous work on ice from this and other glaciers showed the inverse of this relationship and considerable scatter (Epstein and Sharp 1959; Hambrey 1974, Krouse 1970). Perhaps meltwater samples provide more representative data than the ice samples used by previous workers. Further data are necesary to test this finding, however.

Meltwater from the East Castleguard Glacier in Meadows Creek showed crudely diurnal fluctuations (Figure 4.25), much as have been reported elsewhere (e.g.Ambach et al.1976, Behrens et al.1971). Sampling on the source glacier suggested, however, that a variable source area-routing model might explain these variations just as well as the conventional and rather speculative ablation-base flow mixing model.

The provenance of the cave flood waters is clearly not from new snow or rainfall in this case, and they do not show the cyclicity of proglacial melt. In fact, few waters were found with a similar composition, the most similar waters being mature glacier ice or high altitude winter snow. Extrapolation of the Saskatchewan Glacier data to a value of -174 per mil gives a distance 8km from the snout as a possible source, a location corresponding to a subglacial depression (Figure 4.4). Hydrogeologically, the area north east of Castleguard Mountain, including the upper Saskatchewan Glacier, is the most likely origin. The trace from the moulin on the Saskatchewan Glacier shows waters to drain in this direction, but to the Valley Springs. This route is perhaps the underflow to the cave flood system. A replicate trace made while the system is flooding might test this hypothesis.

7) Conclusions: the Meadows Karst

The Meadows karst has been shown to be non-linear in its hydrologic behaviour, because the configuration of the active system changes over time. Three elements may be identified: the low flow conduit system, flood flow conduit system and the large, slow draining aquifer which maintains perennial flow.

The low flow conduit system is one of very small magnitude, nourished by rain and melt events superimposed on "base flow" from the Meadows reservoir. The rain response and diurnal variations indicate a very small catchment. The negligible lag between discharge and hydrochemical response suggests that there is little storage in the conduit. Therefore water does not mix with Red Spring conduit waters as happens during cave flood events. The dye trace from the upper Meadows to the Red Spring at low flow followed a different conduit to that taken by rain-fed runoff, and behaved quite differently. The residence time was 30h compared to 5 to 10h for rain events. Only a small portion of the injected dye was recovered, either at the Cave Springs or in the Valley. The flow route to the Big Springs was marginally more efficient and represents an underflow loss. The 90% of the injected dye lost may tentatively be assigned to reservoir recharge, assuming no undetected outlet. During the summer, the chemistry of Red Spring baseflow is similar to that of the sinking streams in the upper Meadows.

During low flow conditions the link between the Meadows and the Valley Springs was able to handle runoff from the most intensive storms. In warm weather, however, the capacity of this link is exceeded and overflow floods from both melt and rainfall occur at the Cave Springs. The rainfall floods suggest that much of the upper Meadows becomes an active catchment for the Cave Springs. This catchment also appears to produce floods during spring snowmelt, while in summer, meltwaters from the upper Saskatchewan Glacier and vicinity are their probable origin. The trace from the latter area in low flow demonstrated the underflow link to the Valley Springs. Unfortunately, this trace was not quantitative and no breakthrough curve is available.

A relationship between the Meadows System and the Valley Springs is known to exist. The dye traces confirmed this link, and the turbidity record shows that cave flood events have an impact on the Valley Springs (Figure 4.16). Although a simple overflow-underflow relationship has been rejected, the Meadows appear to constitue an important recharge area for the Valley Springs. The early morning pulses recorded in the Meadows Creek Springs (Figure 4.13) and the turbid response of the Big Spring to cave floods (Fig 4.16) may represent this recharge taking place.

The Meadows Reservoir is the body of water sustaining the base flow of the Red Spring. This aquifer shows only subdued response to recharge events, and a gradual increase in hardness as residence time increases.

Mangin (1975) would refer to such a body as a "systeme annexe". The nature of the storage medium which constitutes this feature is unknown, but in the present context two possibilities may be envisaged. The perched Eldon-aquifer inferred to exist beneath the Terrace Range may be leaking into the upper Cathedral. Or, some pre-existing karst may lie beneath the Meadows and be partially infilled by unconsolidated deposits. Figure 4.26 shows a simple model of the Meadows karst System based on the discussion above.

GLACIER	ICE AREA (Sq.Km)	RUNOFF (1000's m**3/day)	SPECIFIC RUNOFF (mm/day)
PEYTO (Young)	13.9	400(min)	28.8
		800(norm)	57.6
		1400(max)	100.7
SENTINEL (Reid and Cherbonneau)	2.0	104	52.0
ATHABASCA (Mathews)	18.4	432	23.5
		604	32.8
UNNAMED (Weirich)	2.0	120	60.0
E.CASTLEGUARD	2.6	35(norm)	13.5
		40(max)	15.4
S.CASTLEGUARD	12.2(full)	284	23.3
	7.5(min)	284	37.8
CASTLEGUARD SPRINGS	30.	1728	57.6
	128.	1728	13.5

Table 4.1. Comparative runoff estimates from various glaciers and the estimated catchment of the Castleguard Springs

1979

MONTH:	JULY	AUGUST	SEPTEMBER
Castleguard:	8.9/11	7.5/31	6.0/21
Banff:	15.6(1.1)	15.2(1.8)	11.5(2.4)
Lake Louise:	12.4(0.0)	12.0(0.6)	7.3(0.0)
Jasper:	16.3(1.1)	15.7(1.6)	11.7(1.8)
Jasper(W.Gate):	13.8	13.6	9.8
Sunwapta:			7.0

1980

MONTH:	JULY	AUGUST	SEPTEMBER
Castleguard:	8.0/6	3.5/31	2.3/21
Banff:	14.7(0.2)	11.2(-2.2)	9.2(0.1)
Lake Louise:	10.9(-1.5)	7.9(-3.5)	5.9(-1.4)
Jasper:	14.7(-0.5)	12.0(-2.1)	9.6(-0.3)
Jasper(W.Gate):	13.4	11.2	
Sunwapta:			5.6
Parker Ridge:			3.1

Notes: 1. Castleguard figures are given as:
Mean Monthly Temperature / number of days of data
2. Other data is given as:
Mean Monthly Temperature (departure from average)
3. All figures in degrees Celsius

Table 4.2. Temperature Statistics for Castleguard and Adjacent Stations

1979

MONTH:	JULY	AUGUST	SEPTEMBER
Banff:	12.6(26)	42.7(88)	30.0(94)
Lake Louise:	29.4(50)	36.9(63)	44.7(99)
Jasper:	29.2(61)	60.4(126)	19.4(57)
Jasper(W.Gate):	43.4	47.0	37.3

1980

MONTH:	JULY	AUGUST	SEPTEMBER
Banff:	30.7(64)	63.6(132)	63.6(199)
Lake Louise:	39.1(66)	60.2(103)	89.0(198)
Jasper:	21.4(45)	64.7(135)	36.4(107)
Jasper(W.Gate):	53.3	91.9	
Sunwapta:			57.7
Parker Ridge:			57.2

Notes: 1. Data are presented as:
Monthly Rainfall(percent of normal)
2. Figures are in mm.

Table 4.3. Precipitation data from nearby stations.

LOCATION	**1979**		**1980**
Big Spring:			182/71
Upper Meadows:	64/40		141/65
Lower Meadows:	77/40	104/62	187/70
Benches:		144/69	219/65

Notes: 1. Data are presented as:
Rainfall in mm / days of record
2. 1979 data from two periods
for comparative purposes.

Table 4.4a. Total rainfall from the Castleguard Area

LOCATION	**1979**			**1980**
Big Spring:				2.56
Upper Meadows:	1.60			2.17
	(1.21)			(1.23)
Lower Meadows:	1.93	1.68	(1.59)	2.67
		(1.24)		(1.26)
Benches:		2.09	(1.61)	3.37

Notes: 1. Unbracketed figures are average daily rain.
2. Figures in parentheses are ratios between comparable figures by location and year. (notice the consistency despite the contrast in total rainfall, Table 4.4a)

Table 4.4b. The ratios between mean daily rainfall from different sites and years for comparable periods.

	STORM 1	STORM 2
Total Rainfall(mm):	4.826	2.540
Total Rapid Runoff(cu.m):	11.106	6.876
Total delayed Runoff:	45.468	27.720
Rapid Catchment(sq.m):	2,300	2,700
Delayed Catchment:	9,400	10,900

Table 4.5. Catchment areas estimated from the response of the Red Spring to Rainfall.

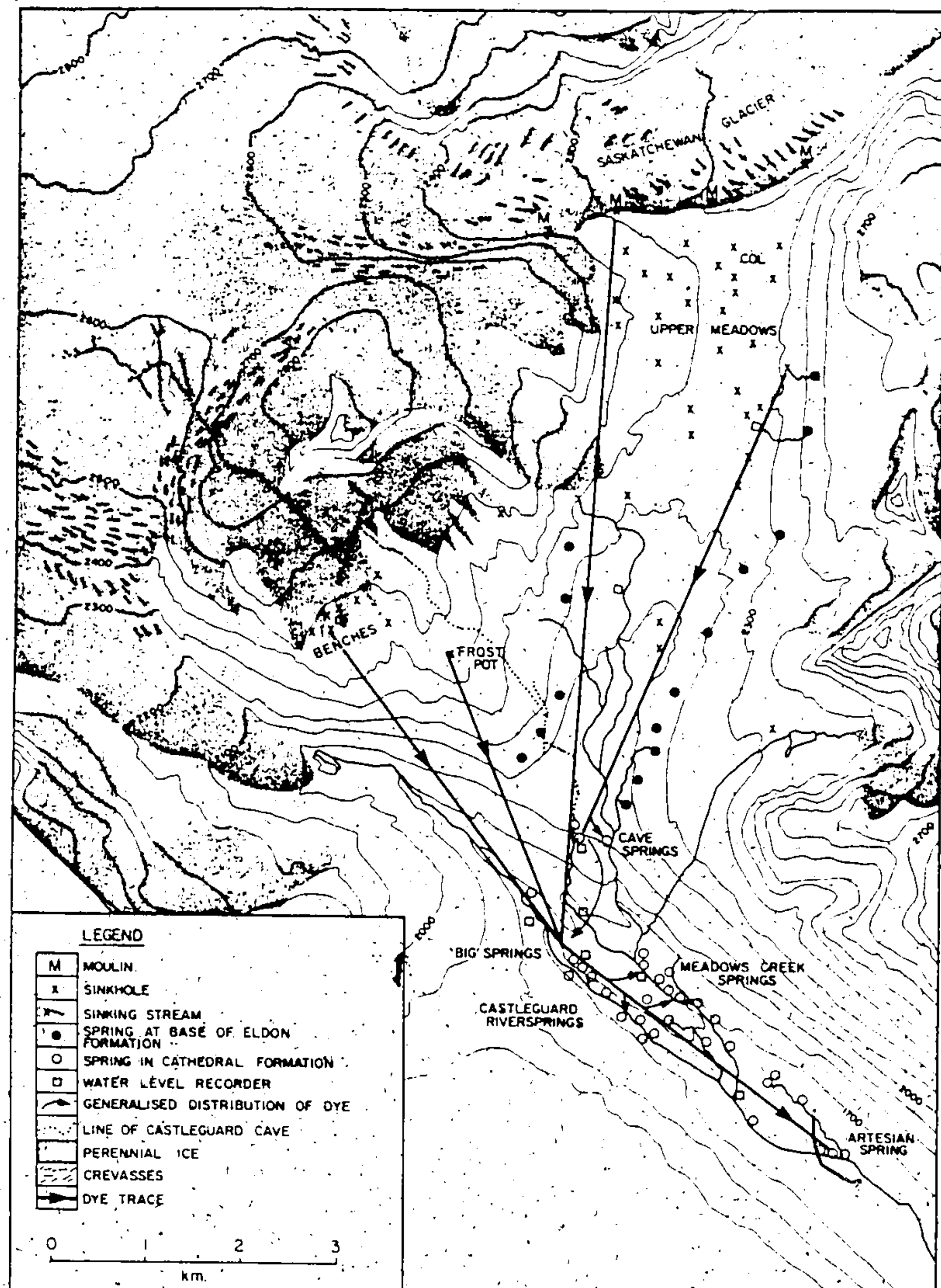


Figure 4.1. Groundwater hydrology and tracer results

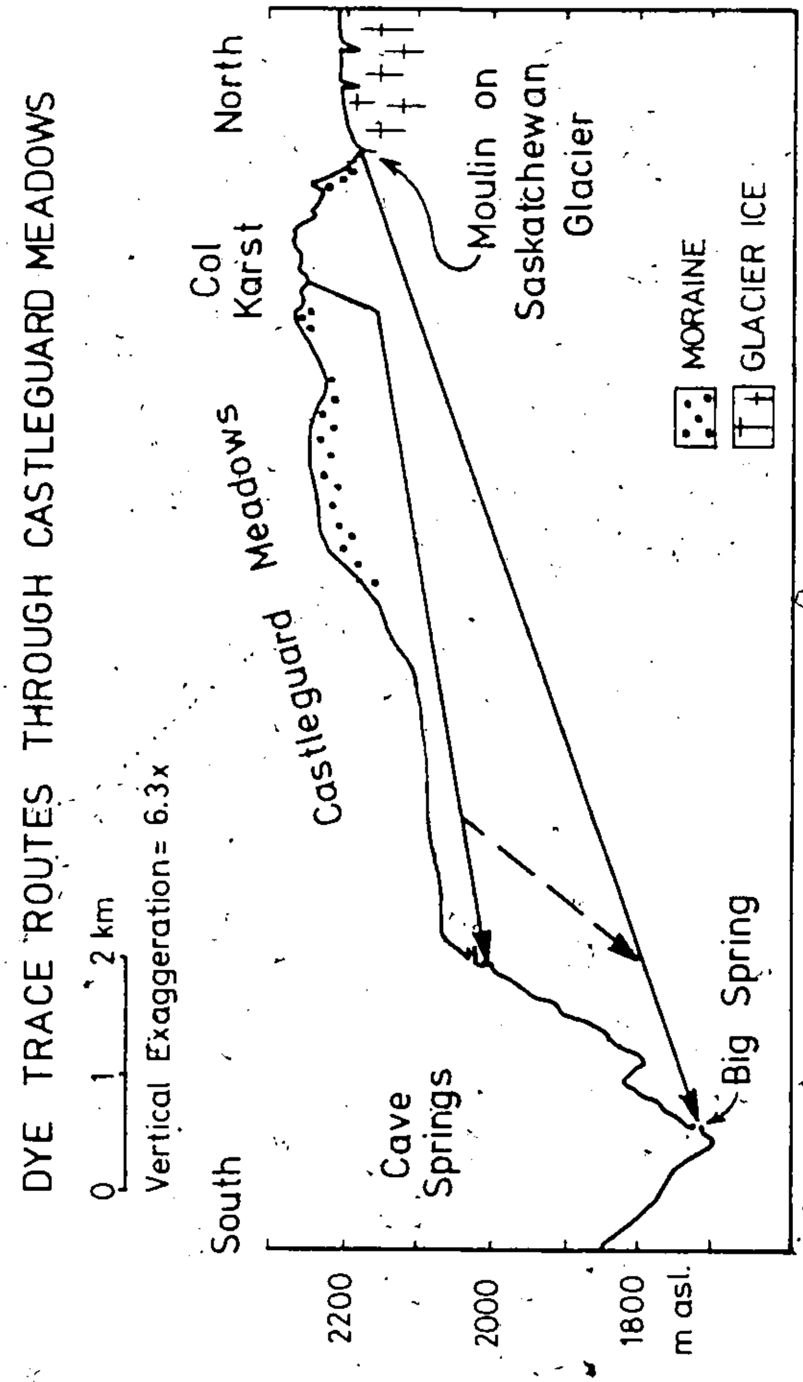


Figure 4.2. Cross-section of Castleguard Meadows, showing tracer trajectory

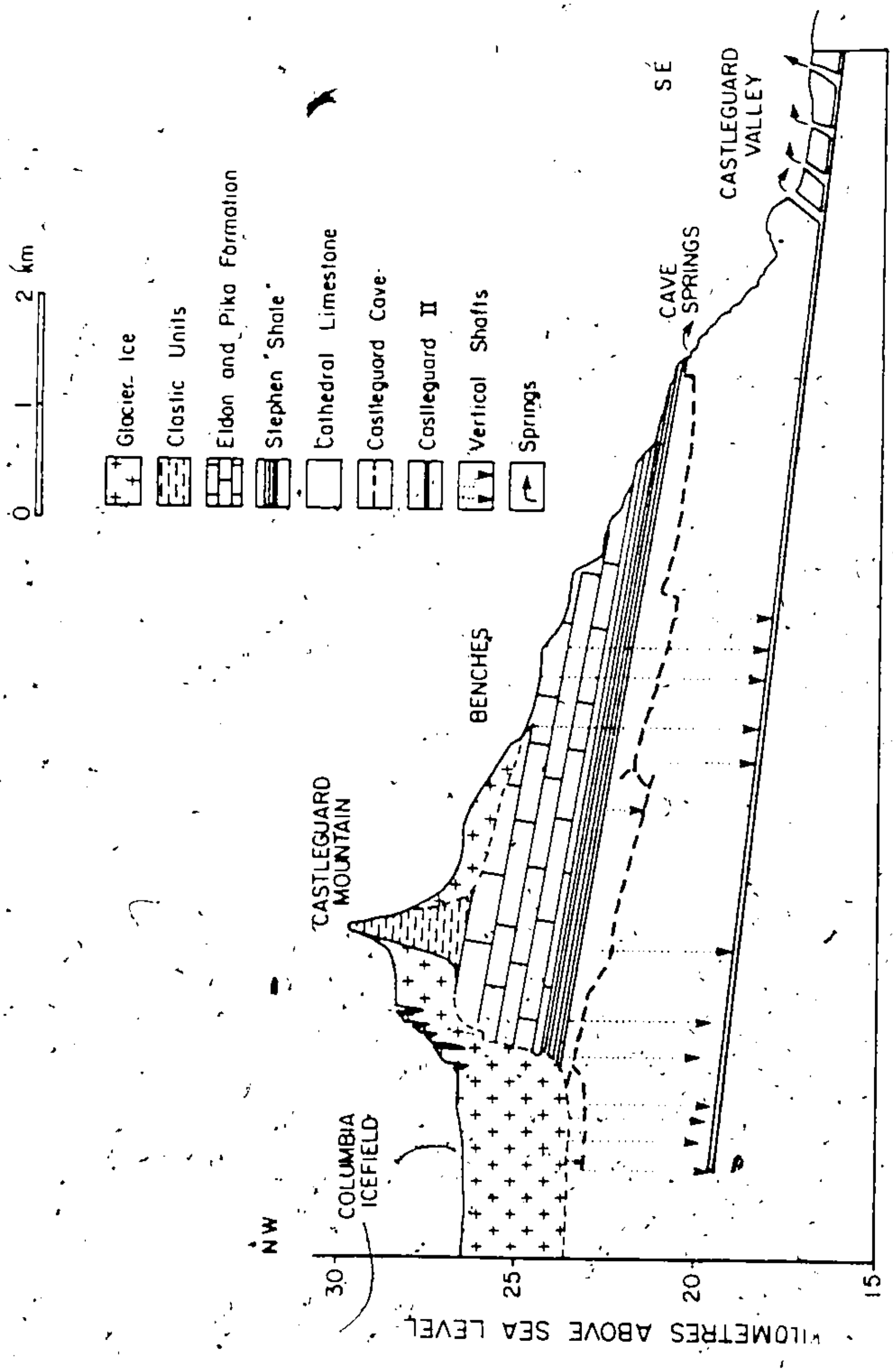


Figure 4.3. Cross-section of Castleguard Mountain showing tracer trajectory and inferred flow routes

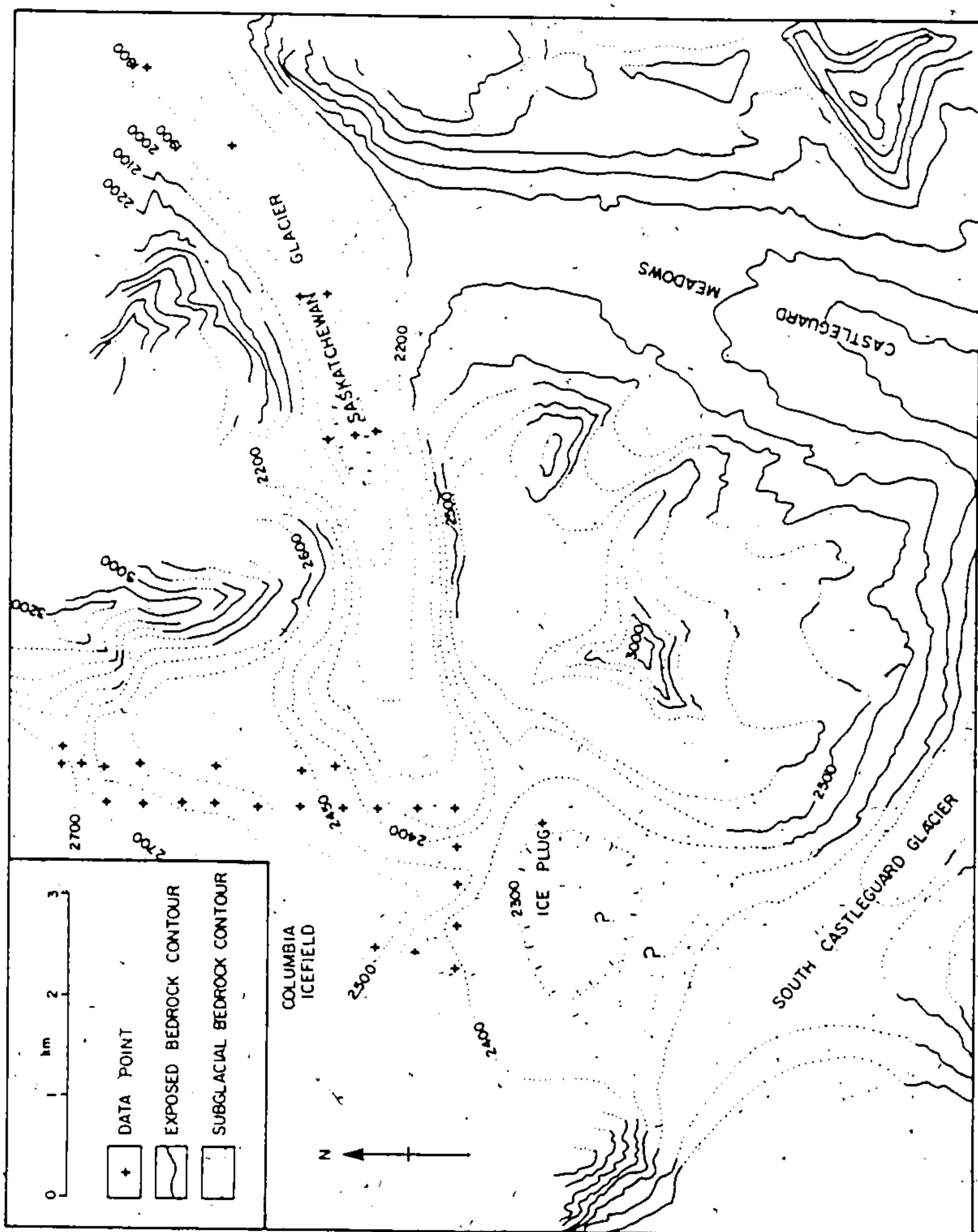


Figure 4.4. Bedrock topography of the southern Columbia Icefield

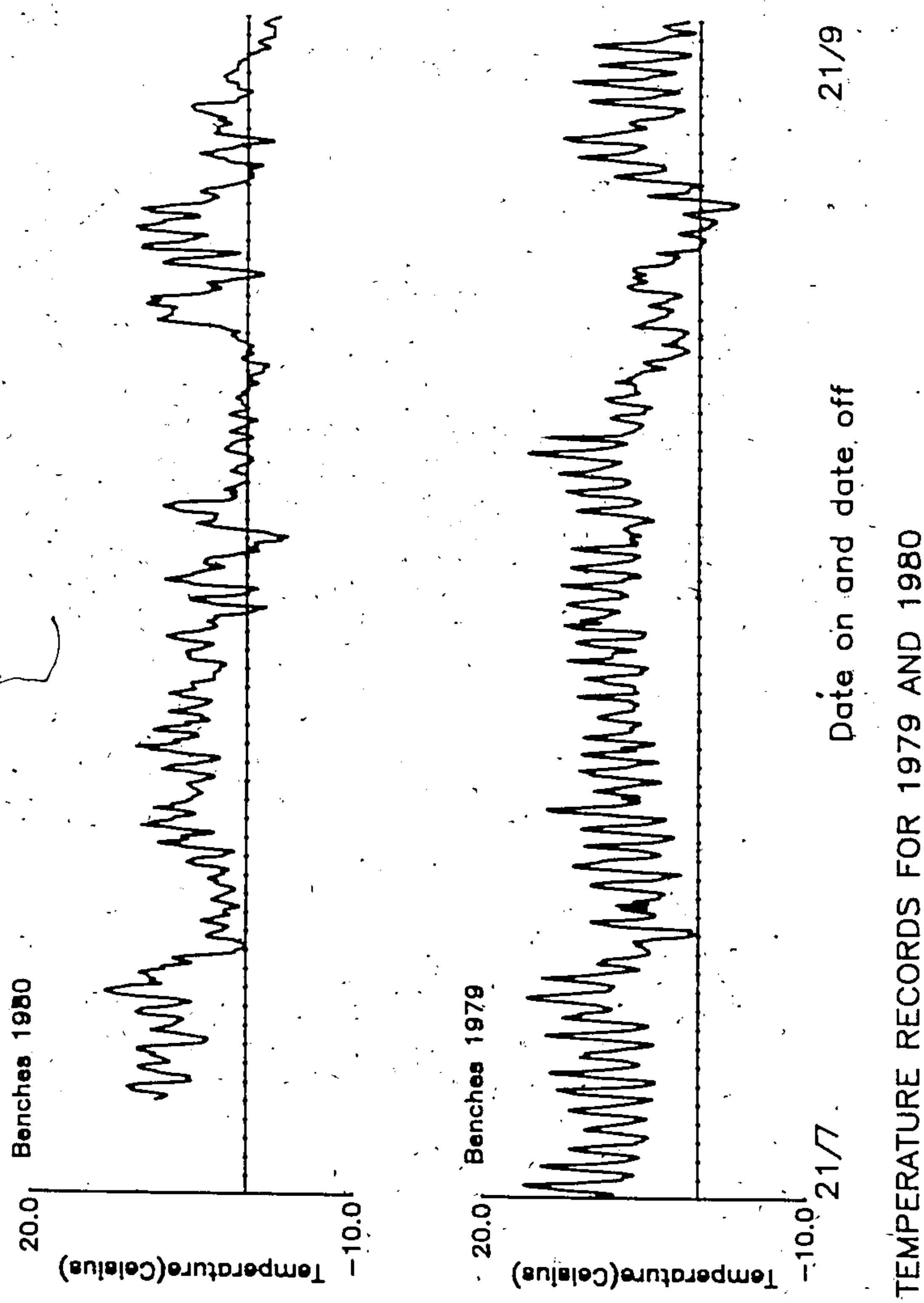


Figure 4.5. Temperature records from the south Benches of Castleguard Mountain: 1979 and 1980

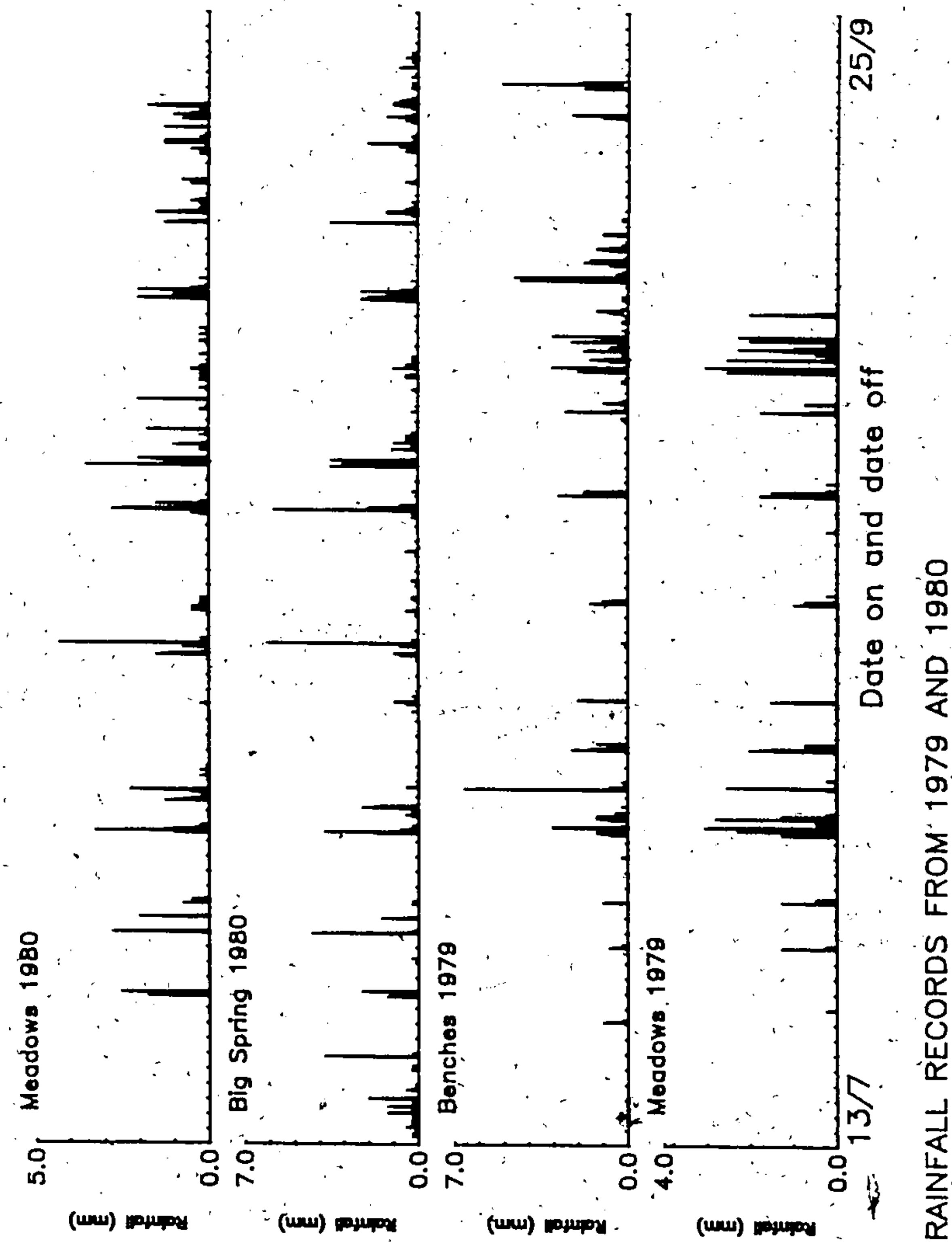


Figure 4.6. Rainfall records from 1979 and 1980

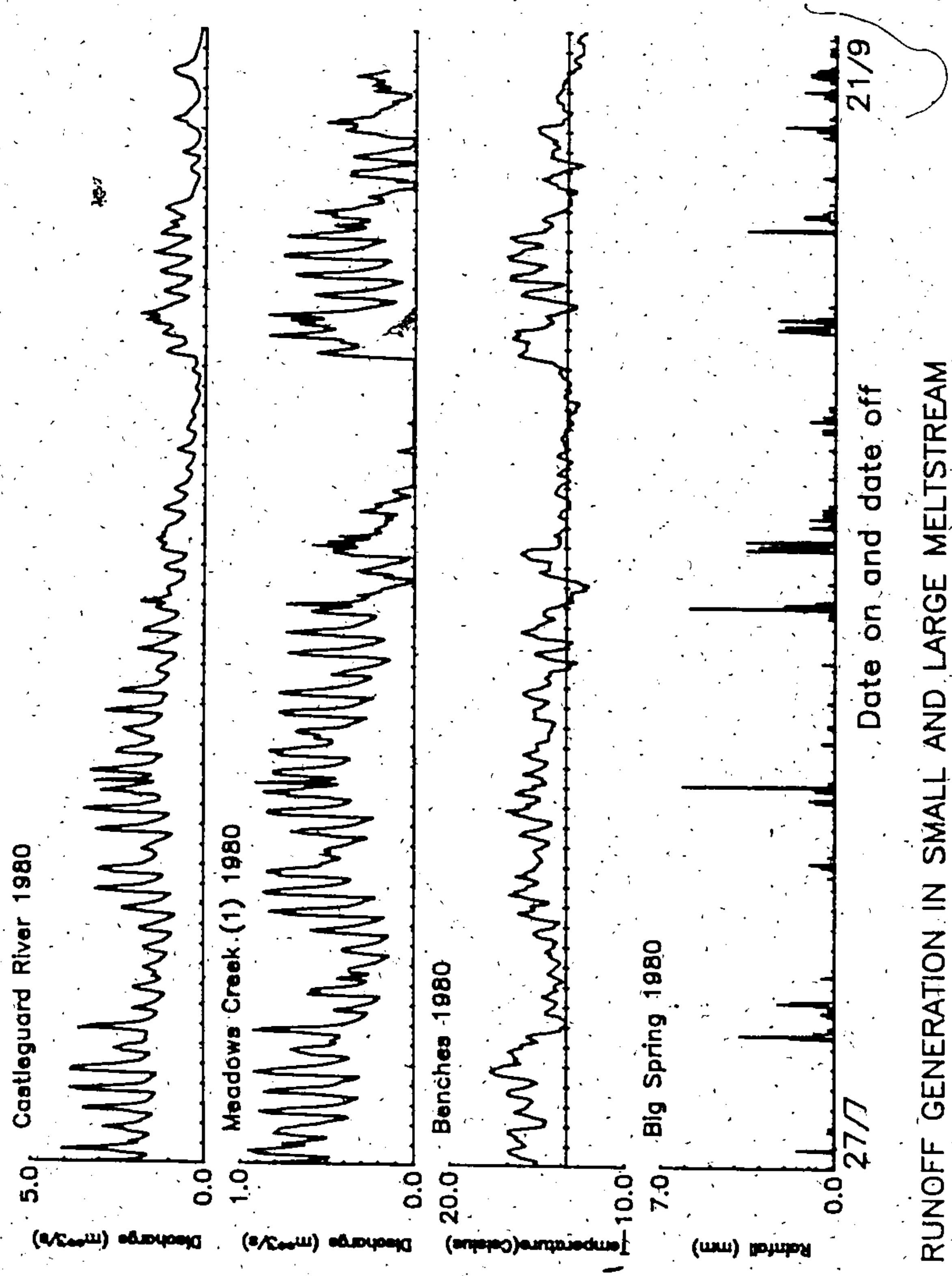


Figure 4.7. Runoff response of two glacial melt streams in relation to temperature and rainfall

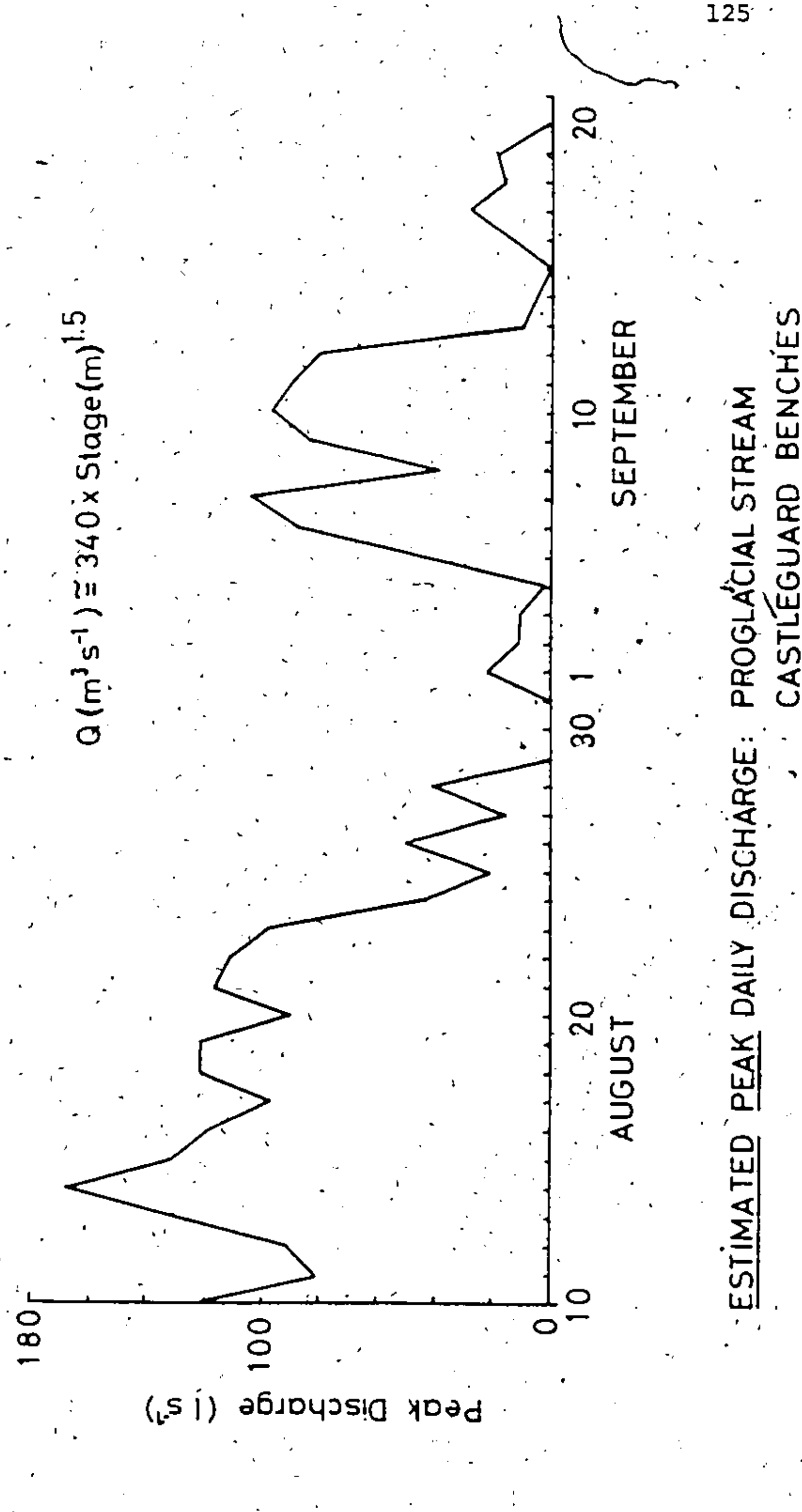


Figure 4.8. Daily maximum flow in a small proglacial stream south Benches, Castleguard Mountain

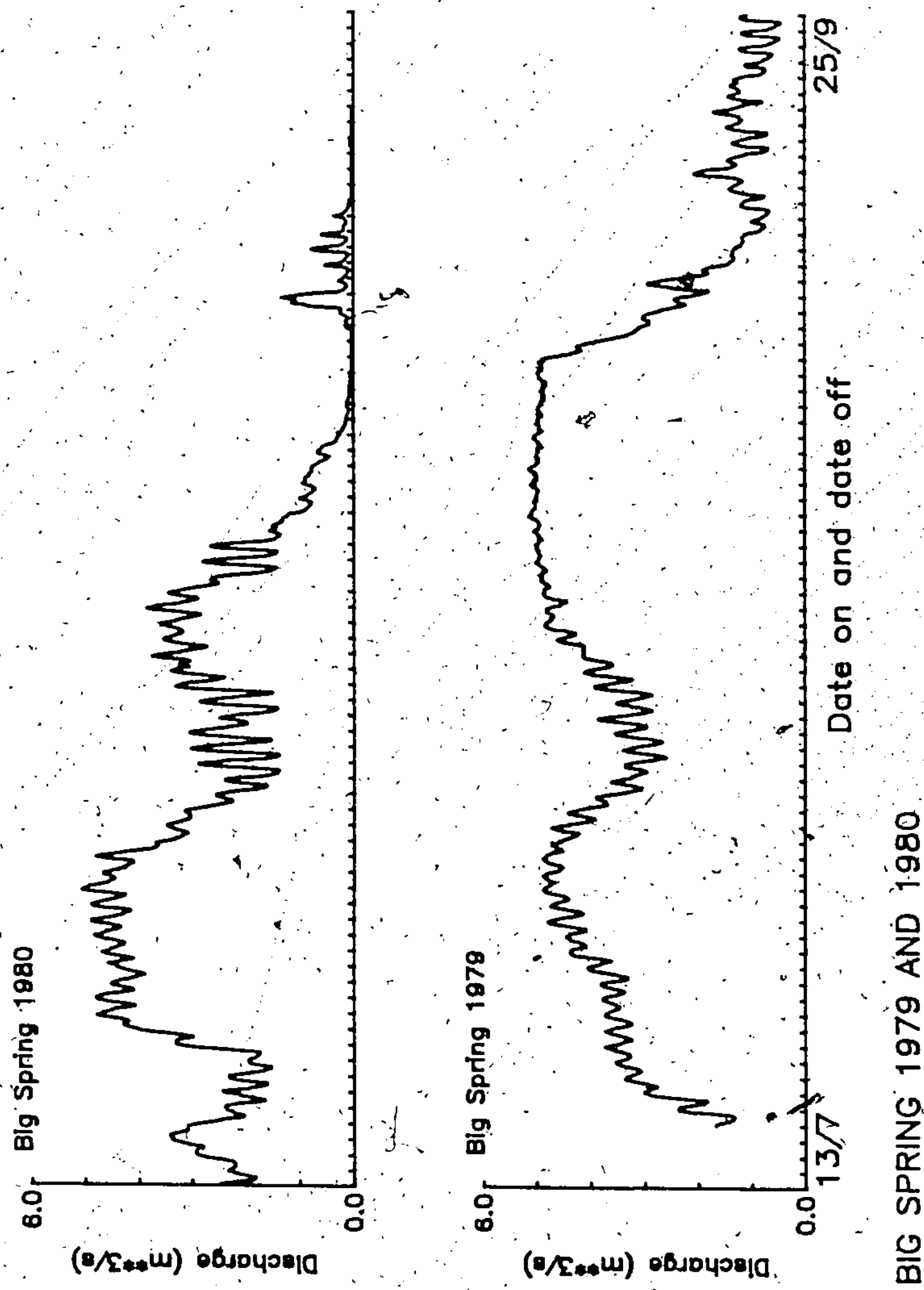


Figure 4.9. Discharge of the Big Spring: 1979 and 1980

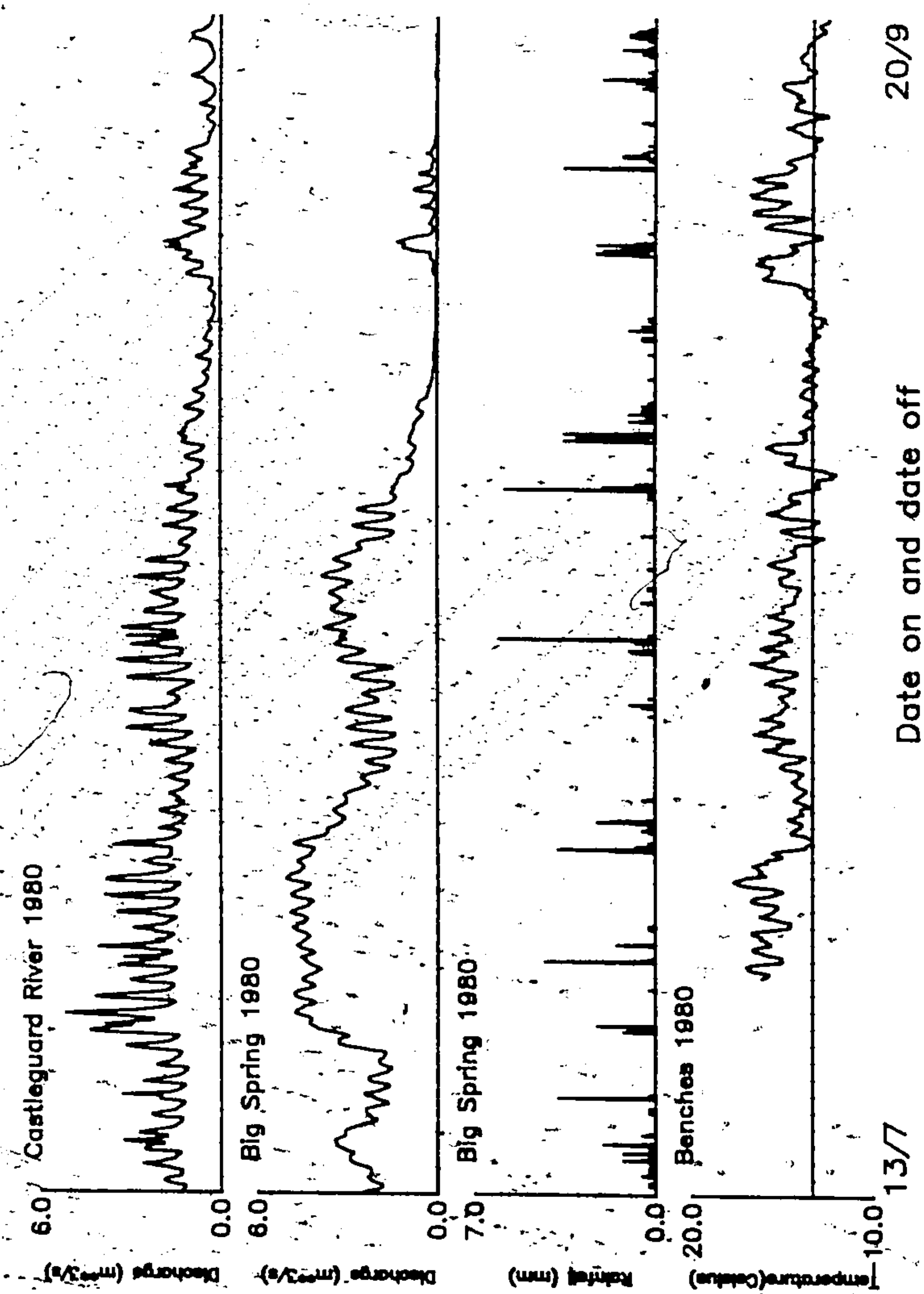


Figure 4.10. Discharge of the Big Spring (1980), in comparison to temperature, rainfall and glacial melt

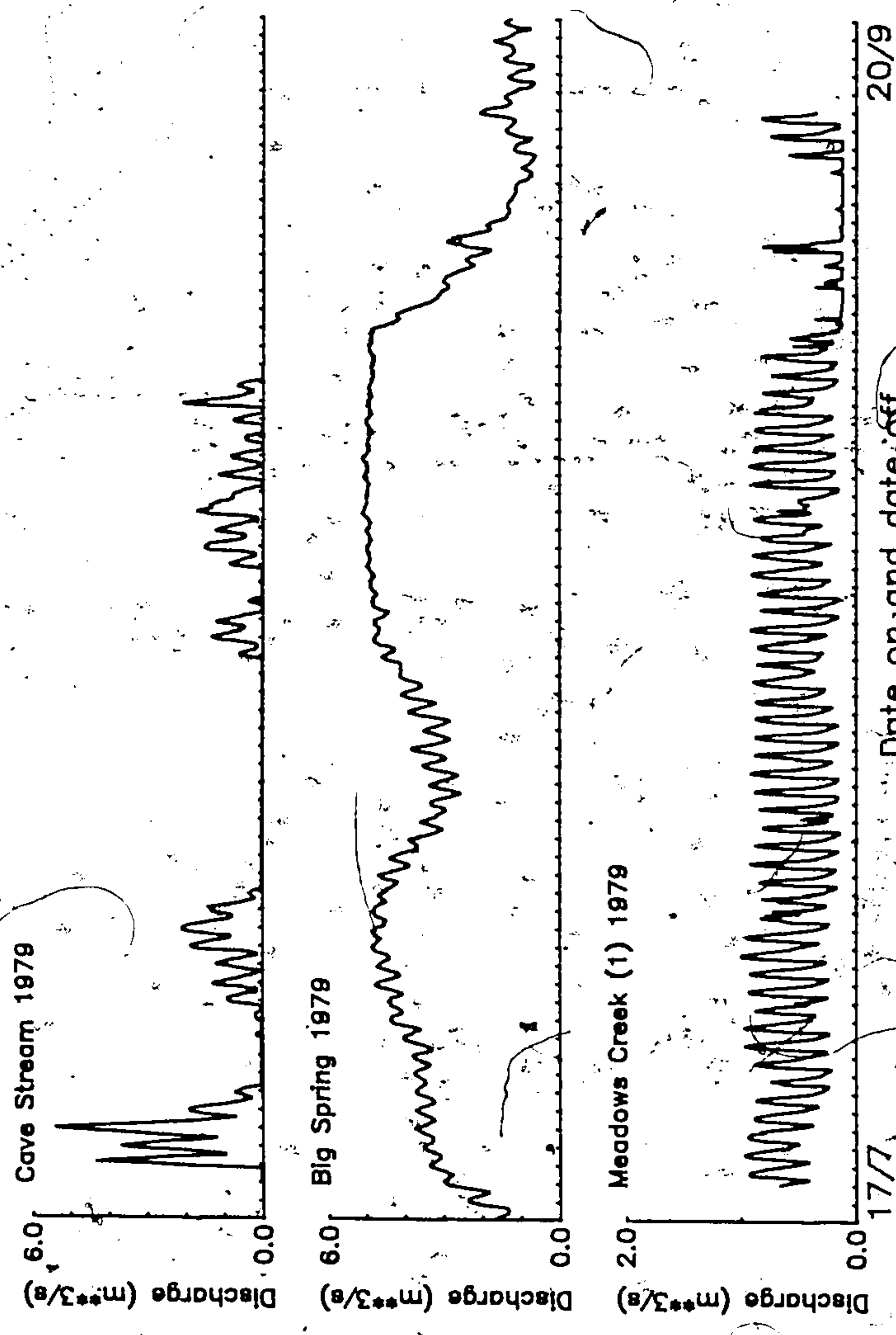


Figure 4.11. Discharge of the Big Spring (1979) in comparison to the Cave Stream and Glacial melt

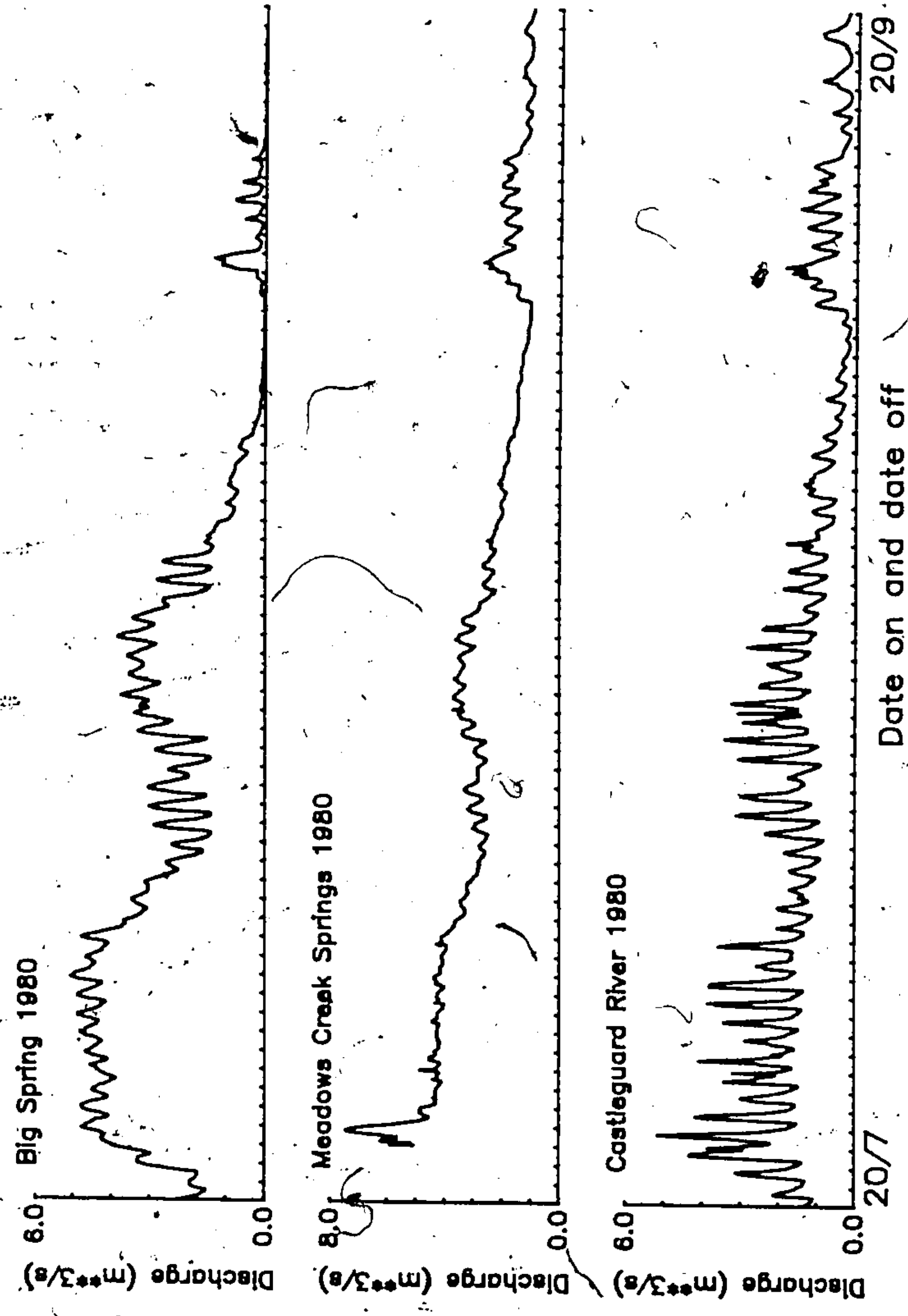


Figure 4.12. Discharge of the Big Spring, Meadows Creek Springs, and glacial melt, 1980

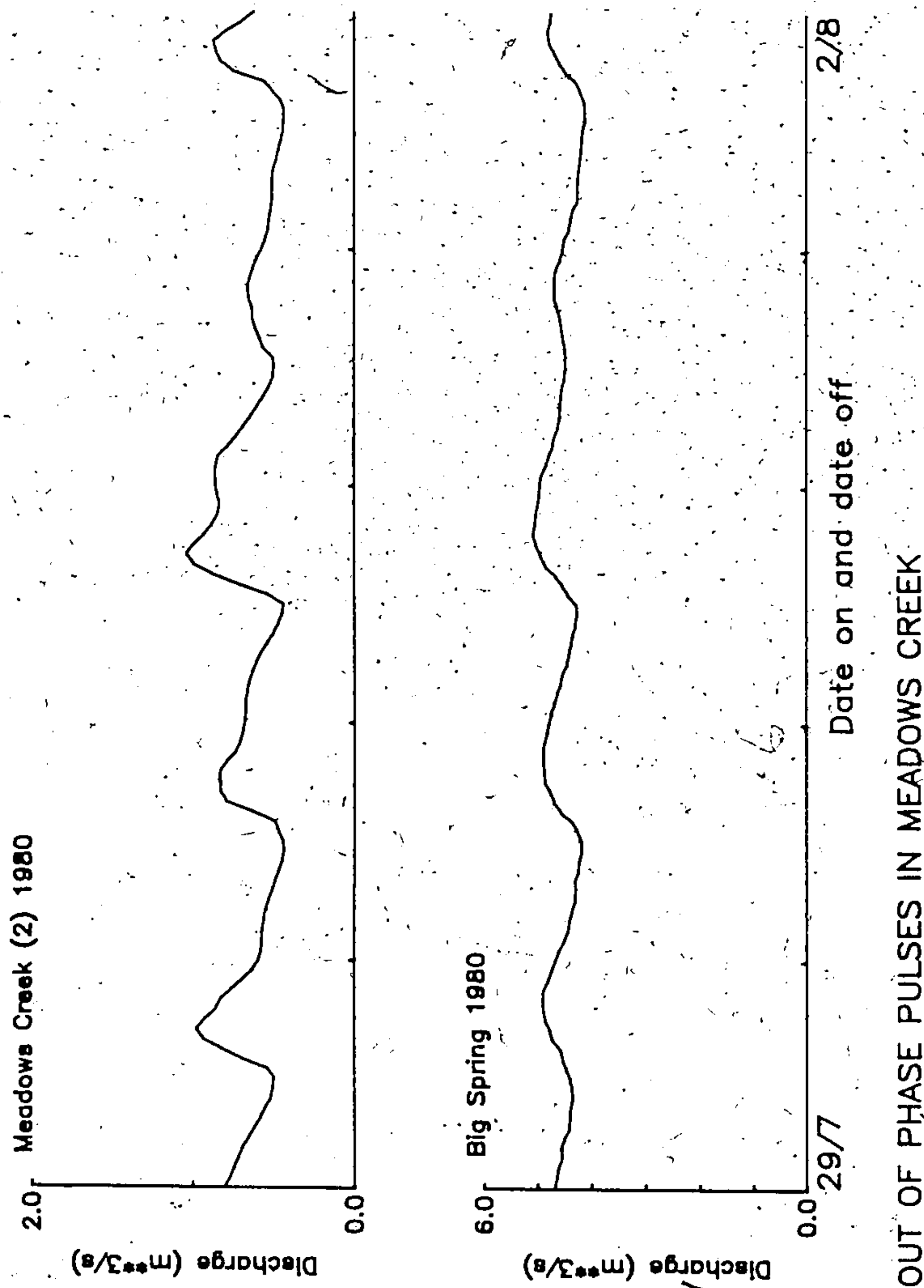


Figure 4.13. Early morning discharge pulses in the Meadows Creek, but absent from the Big Spring

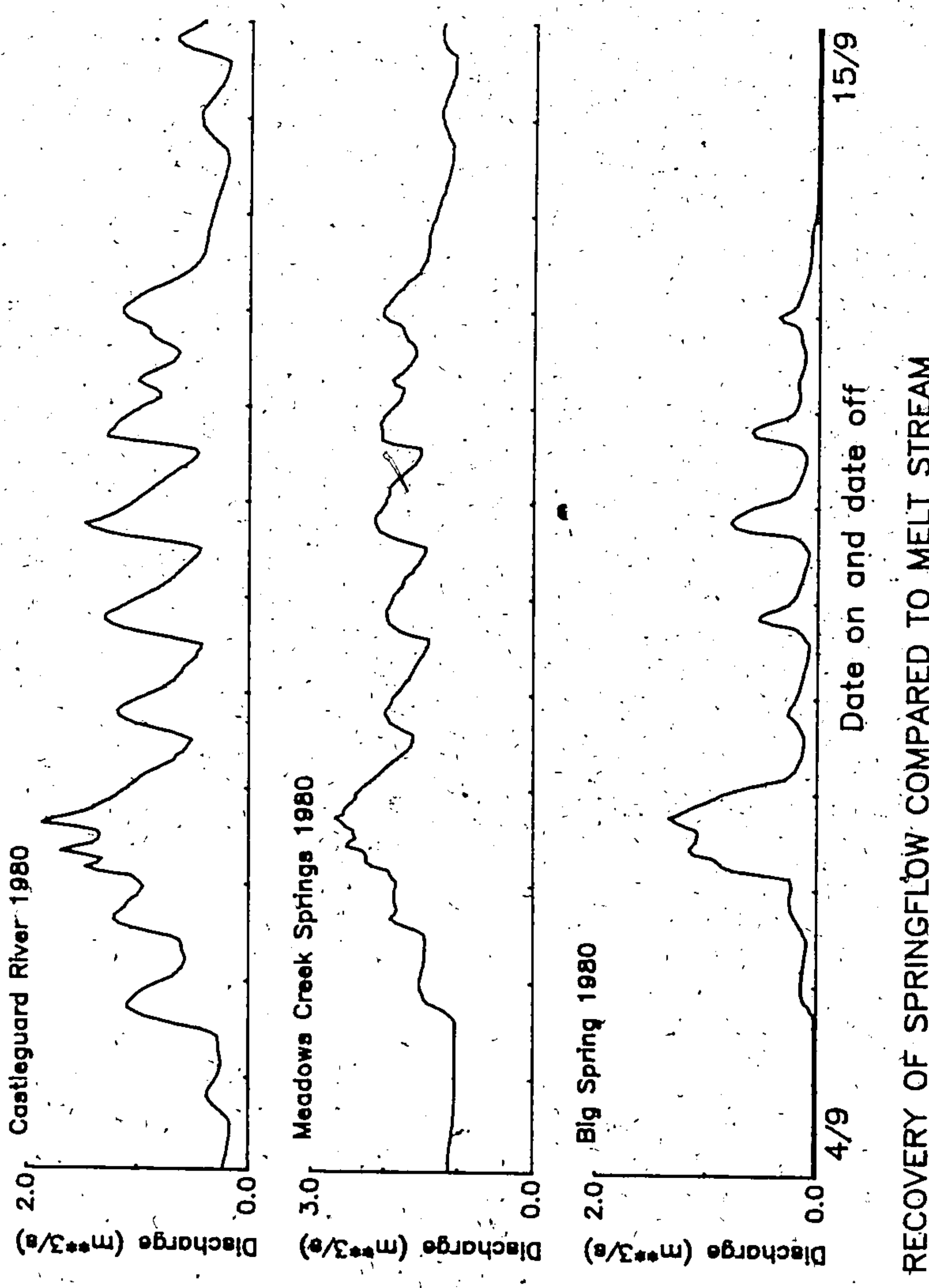


Figure 4.14. Recovery of discharge in September 1980: Big Spring, Meadows Creek Springs, and the Castleguard River

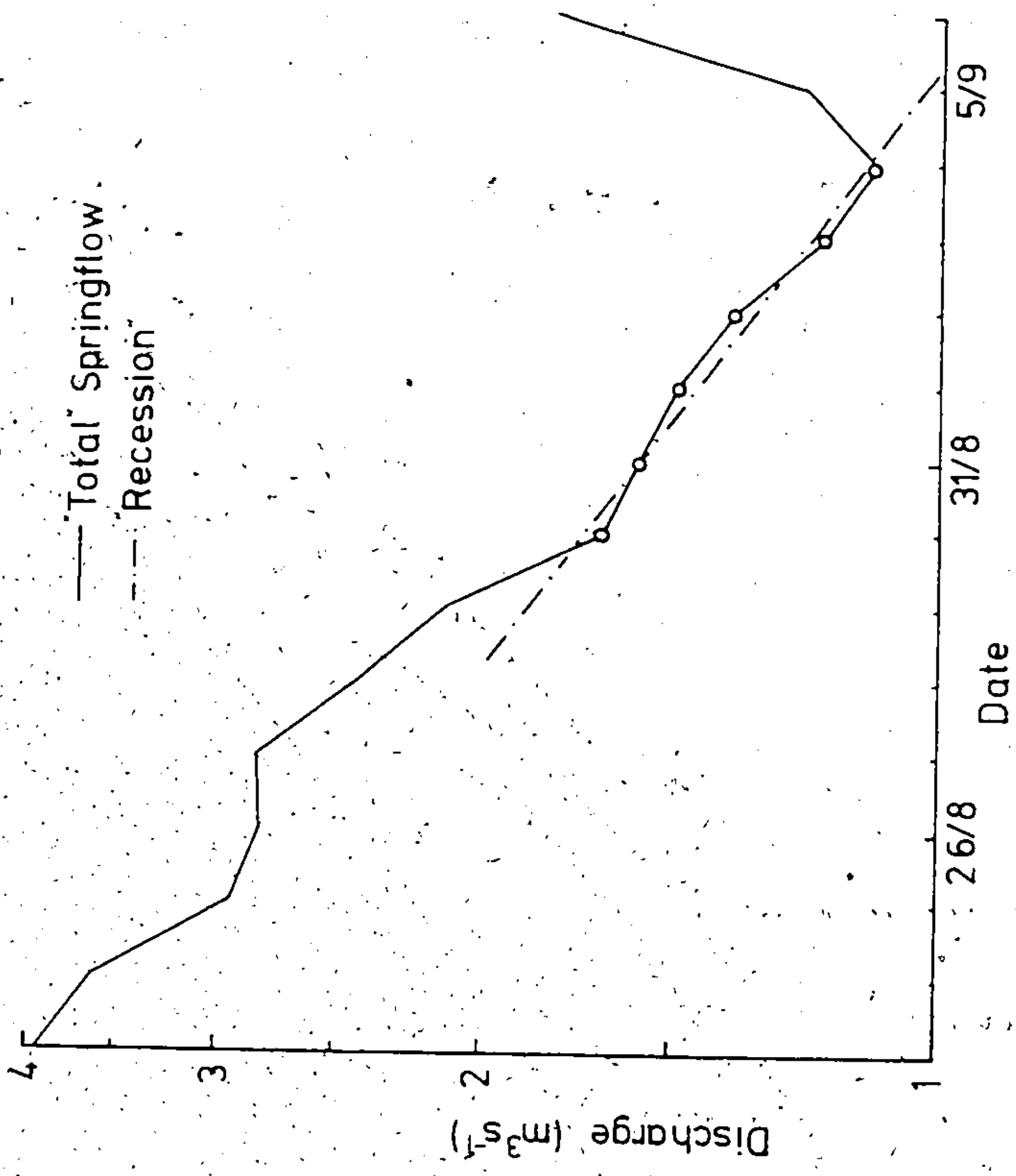


Figure 4.15. Total measured spring flow (Big and Meadows Creek Springs) during a recession in 1980 and best fit exponential decay

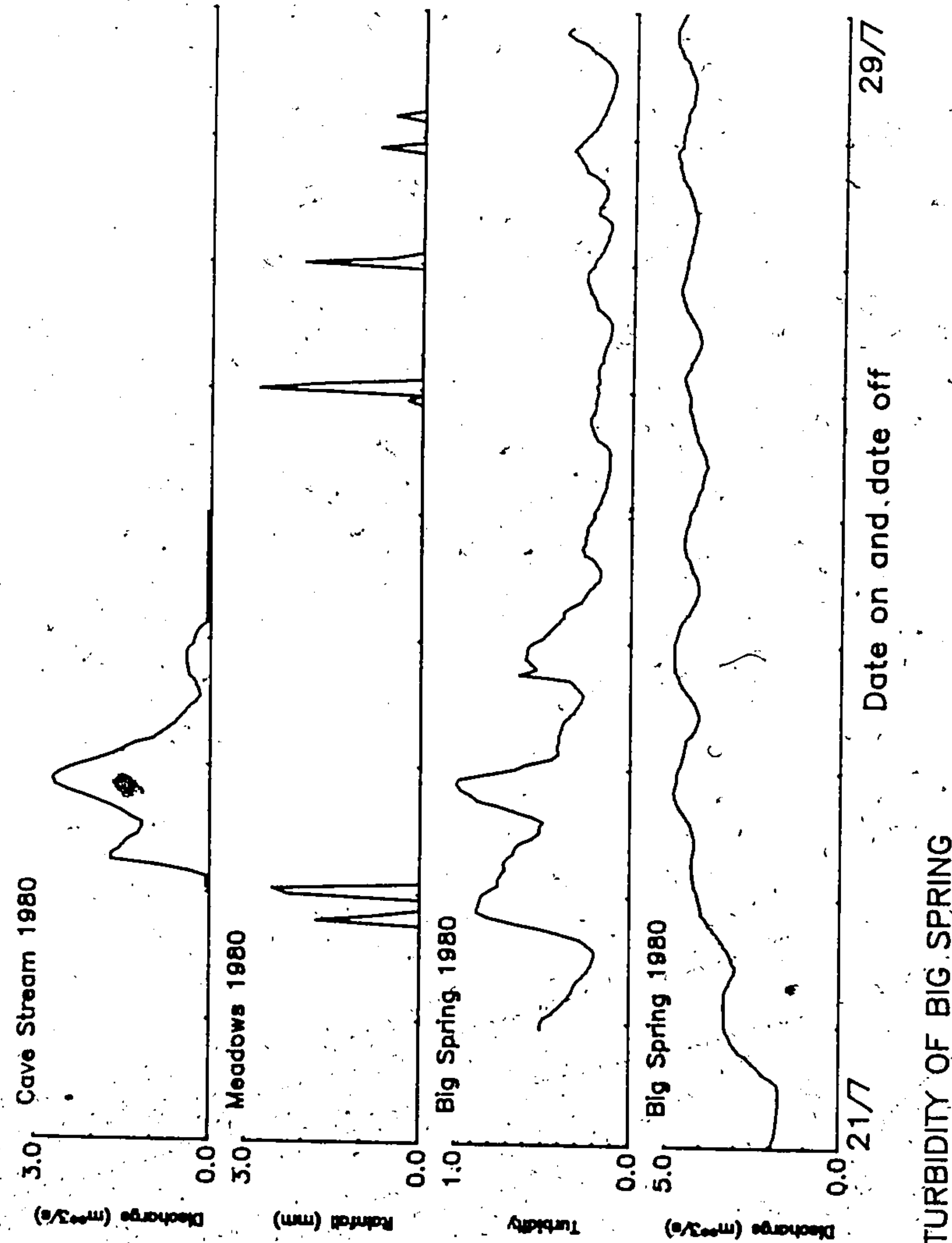


Figure 4.16. Turbidity (arbitrary units) compared to possible influences: Big Spring, rainfall and Cave floods

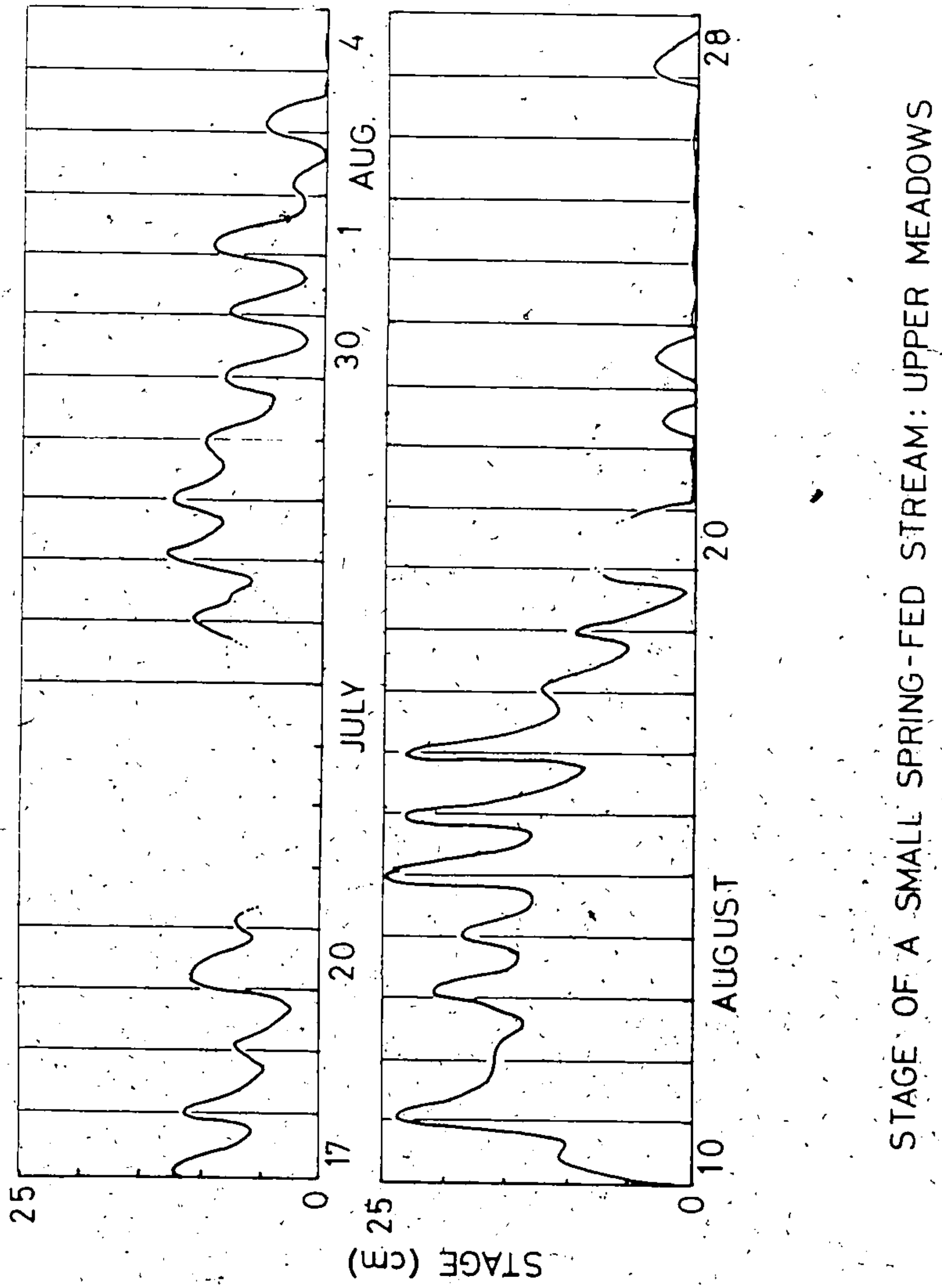


Figure 4.17. Water level in a small stream fed by an Eldon Formation Spring in the upper Meadows

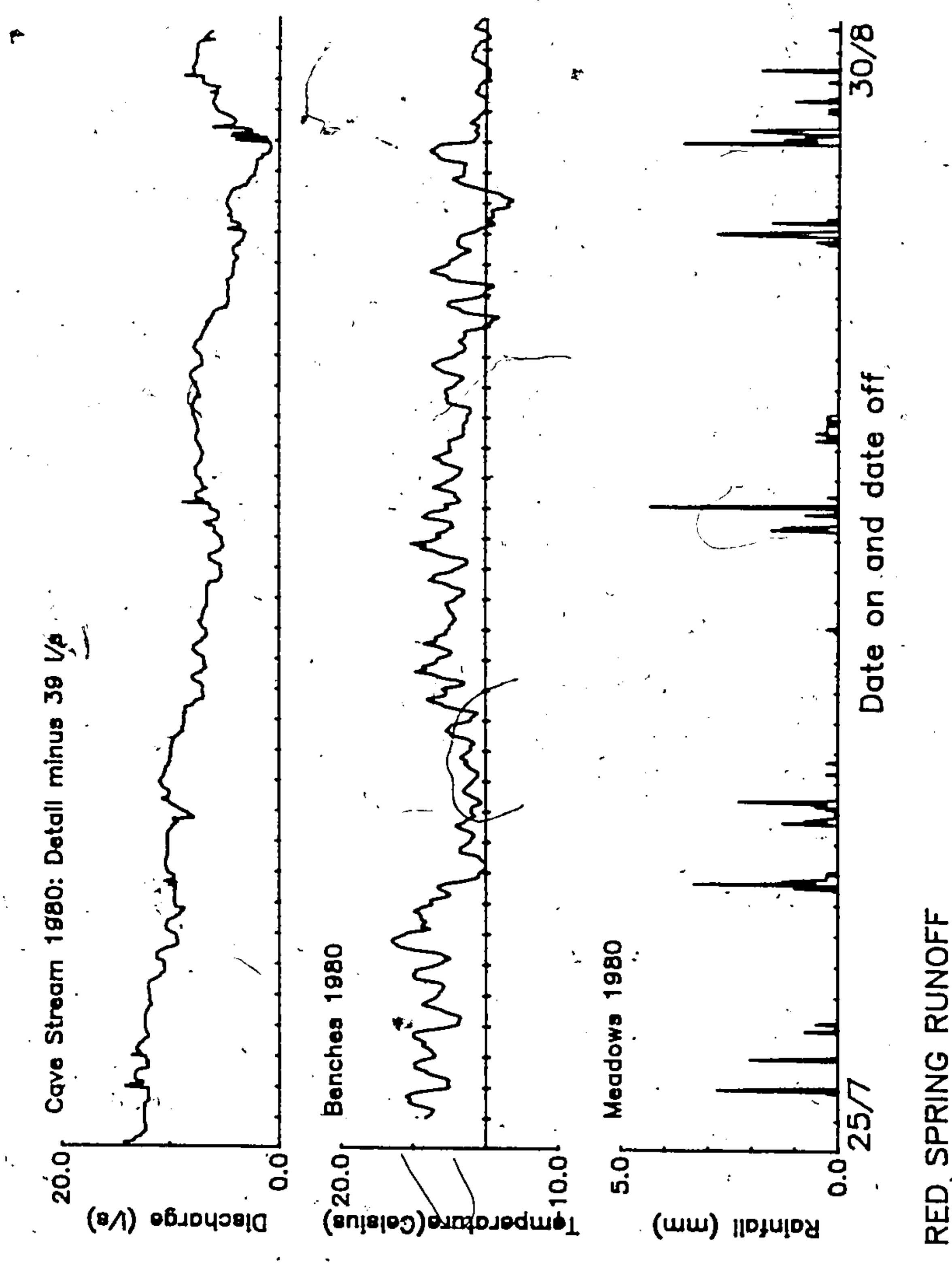


Figure 4.18. Low flow behaviour of the Red Spring (1980), compared to rainfall and temperature.

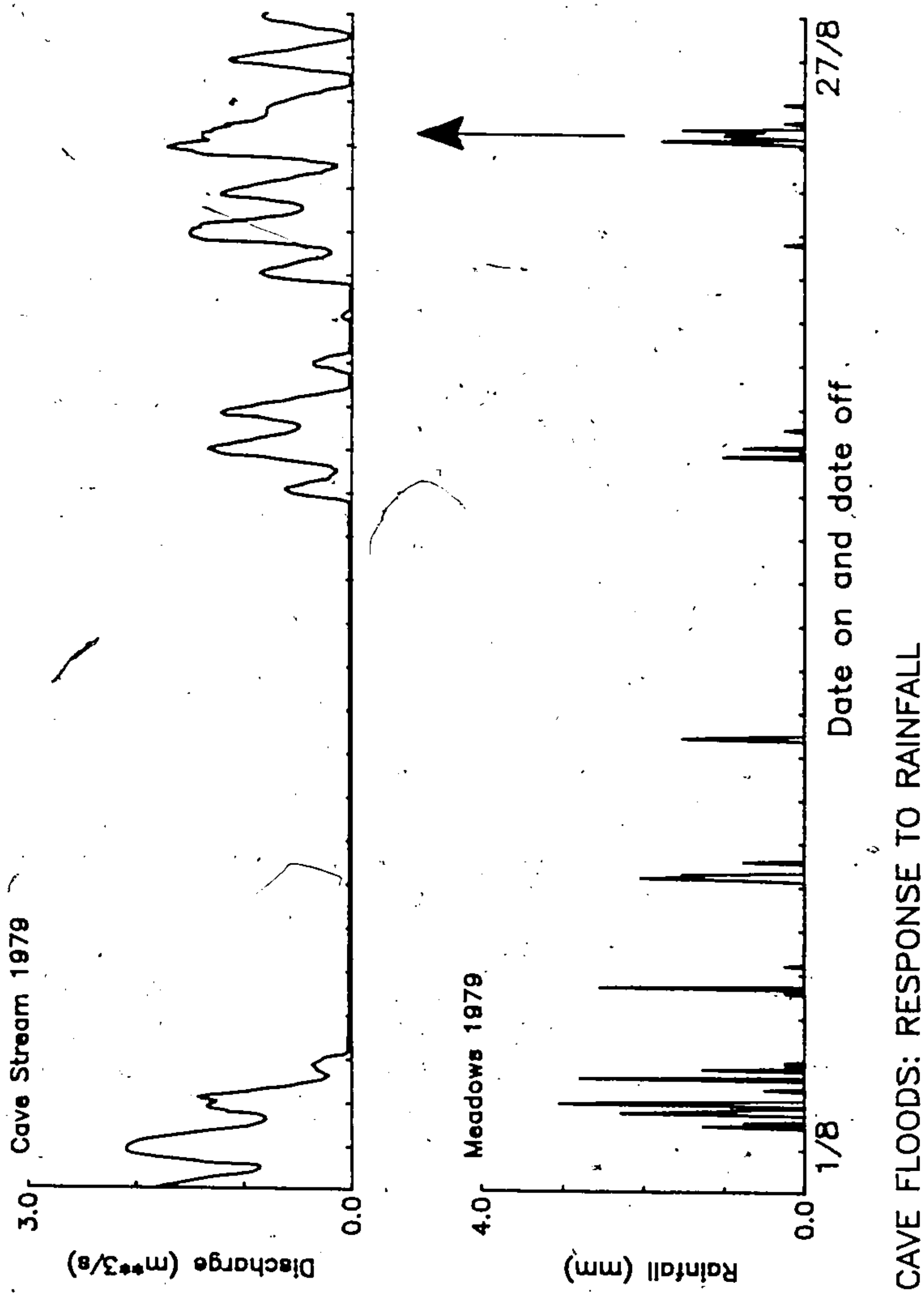


Figure 4.19. Sustained Cave floods caused by heavy rain

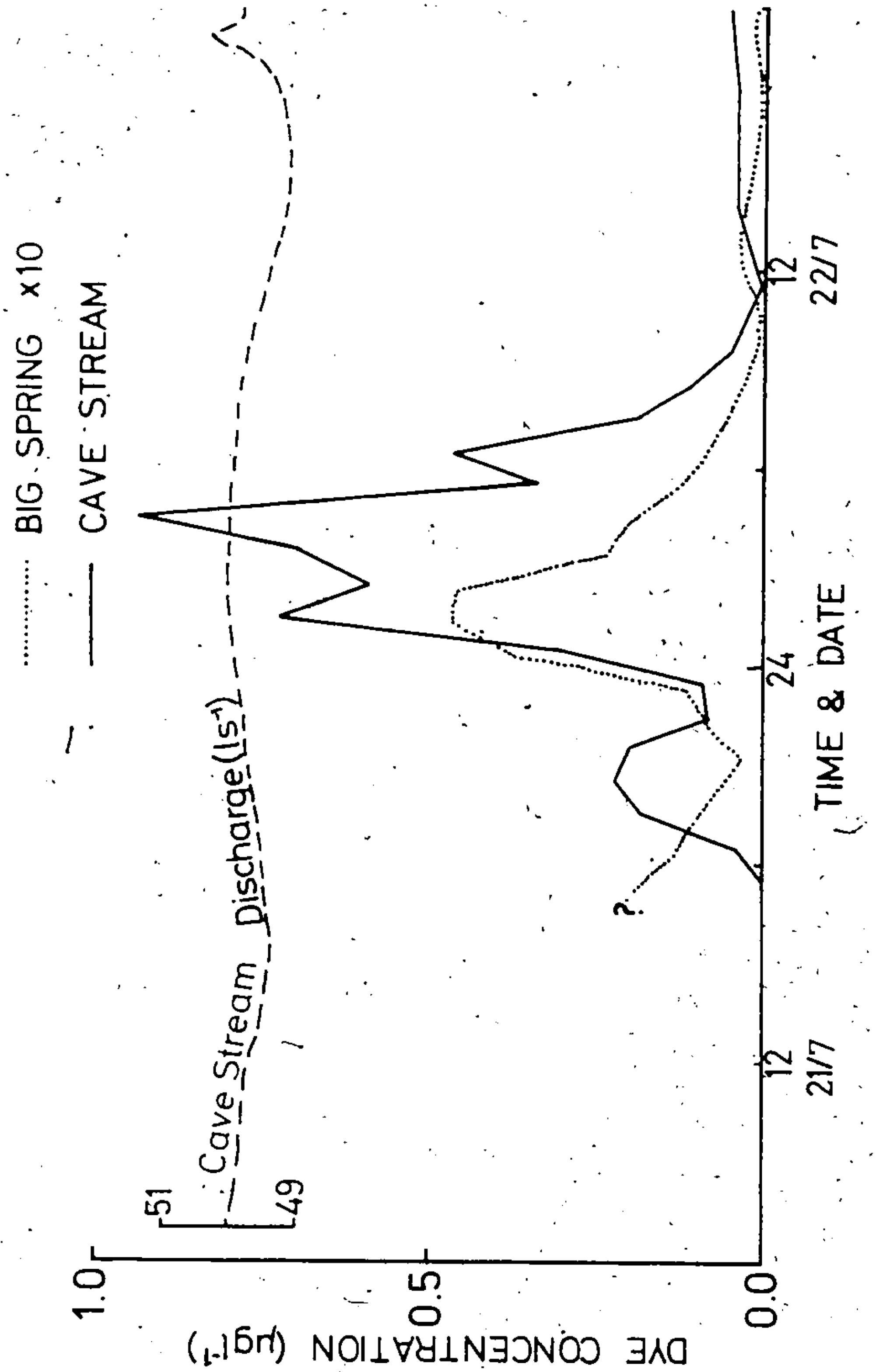


Figure 4.20. Dye breakthrough curves at the Red Spring and Big Spring from an injection on the upper Meadows at 20.30h, 20/7/80

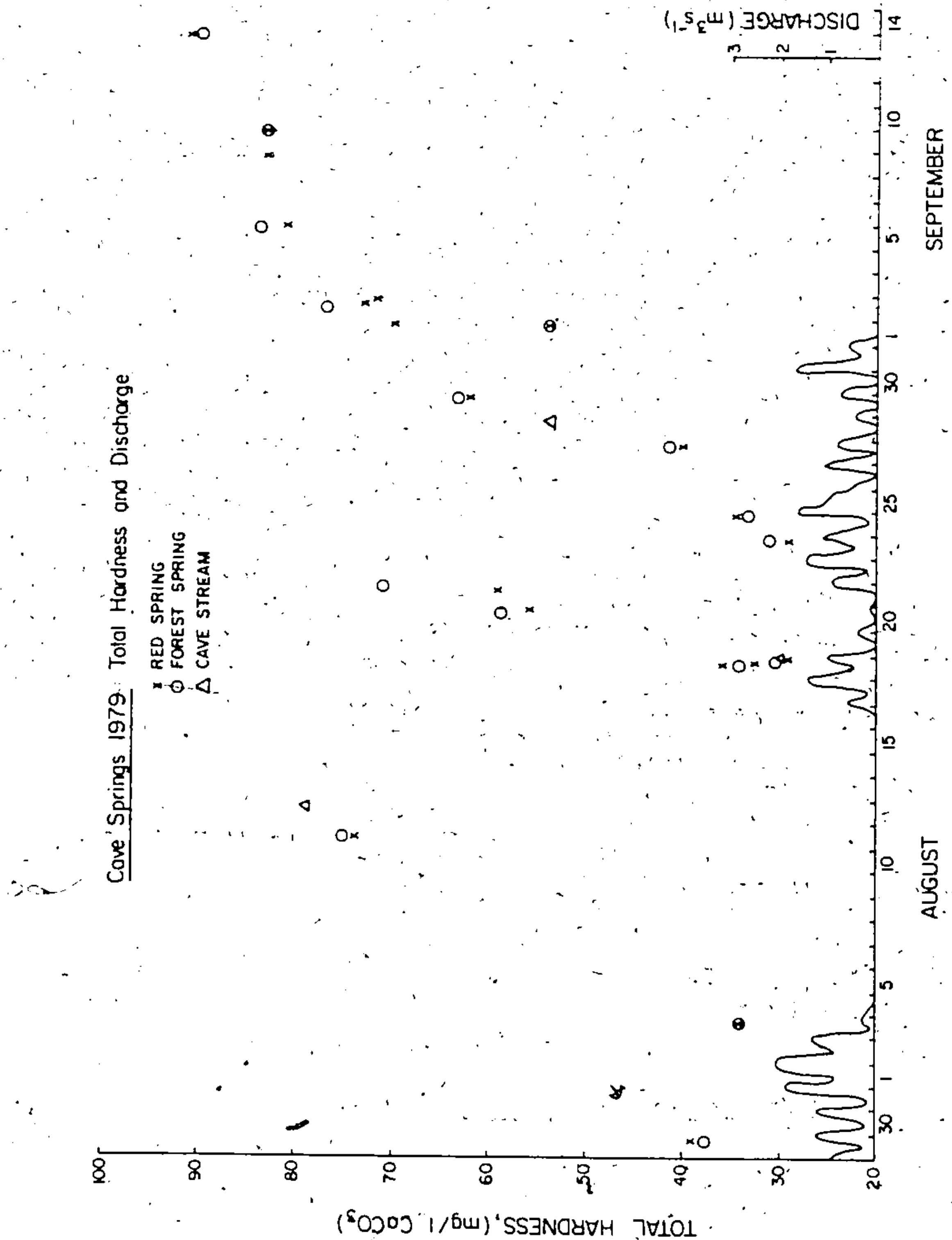


Figure 4.21. Total hardness of the Cave Springs (1979), and discharge of the Cave Stream

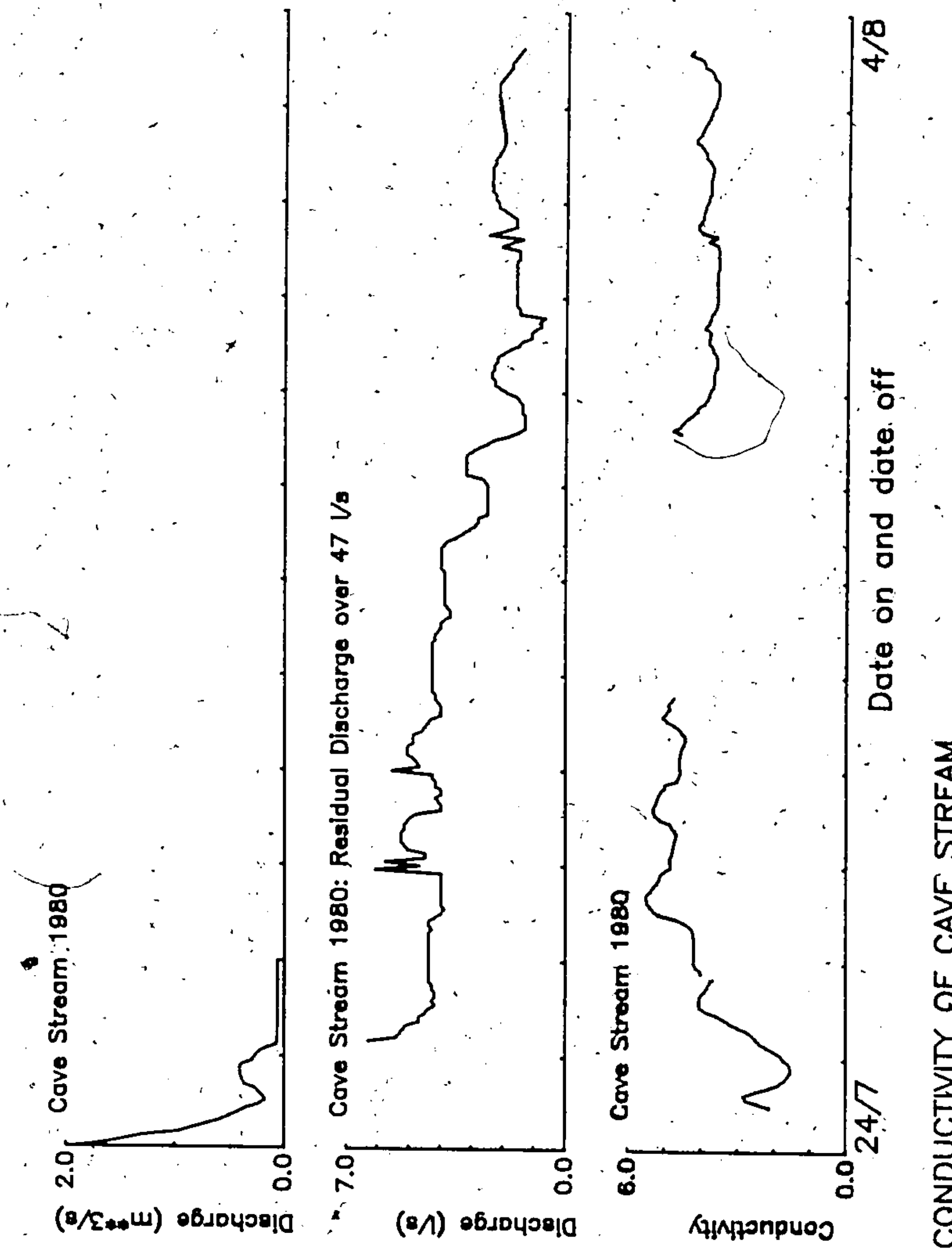


Figure 4.22. Continuous conductivity (arbitrary units) of the Cave Stream (1980), showing response to discharge at two different scales

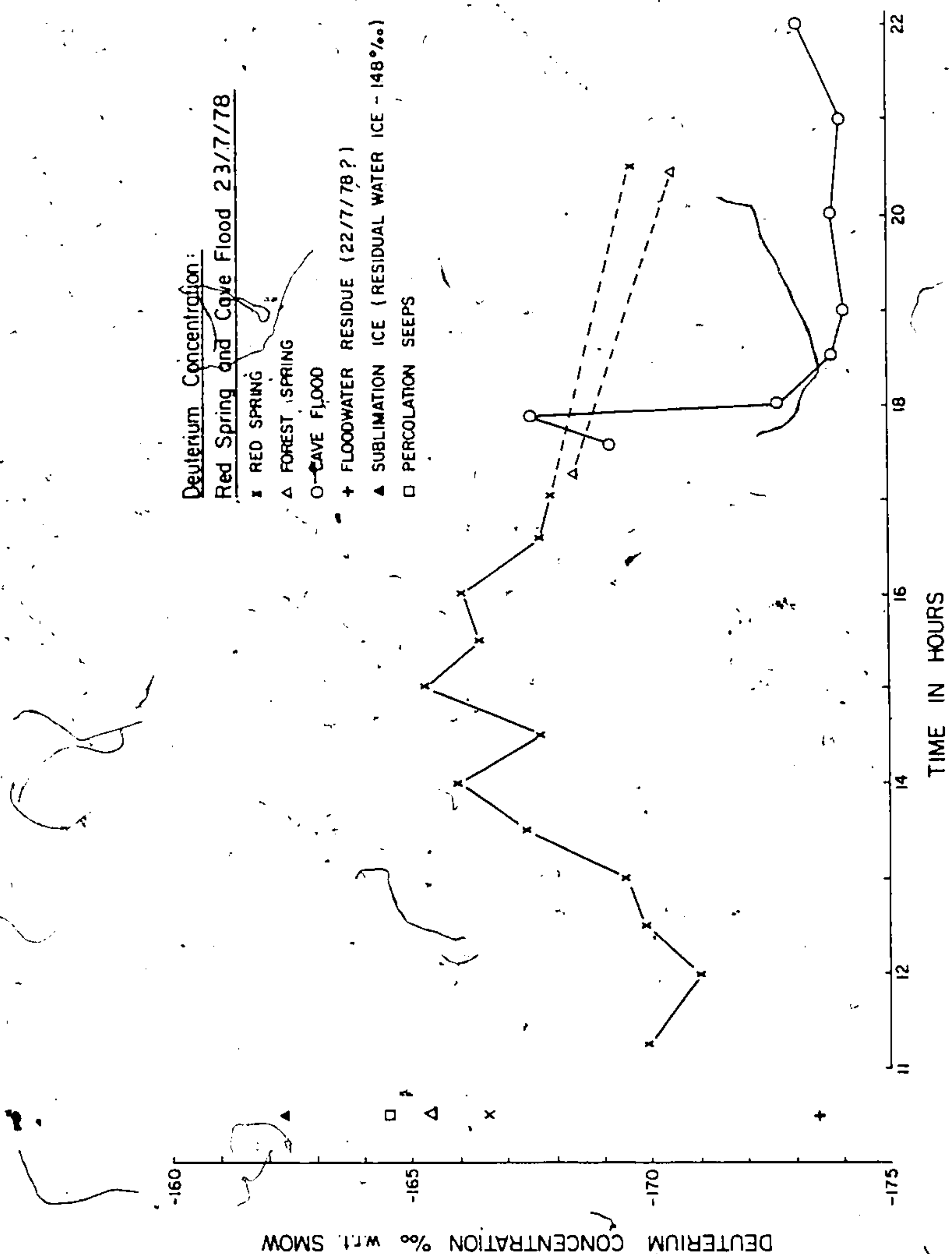


Figure 4.23. Deuterium concentration in the Cave Springs during a flood event (1978)

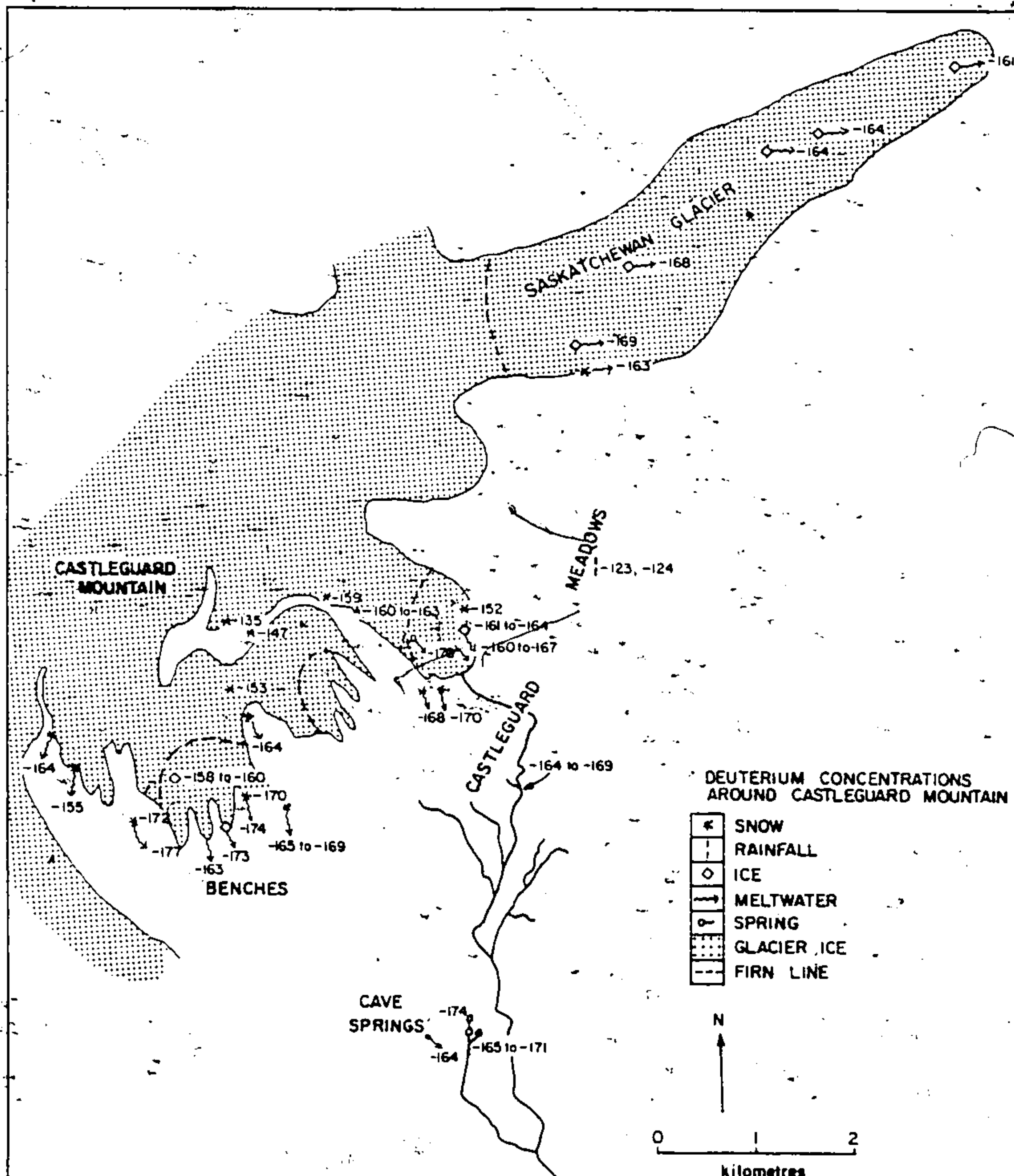


Figure 4.24. Deuterium hydrogeography of the Castleguard area

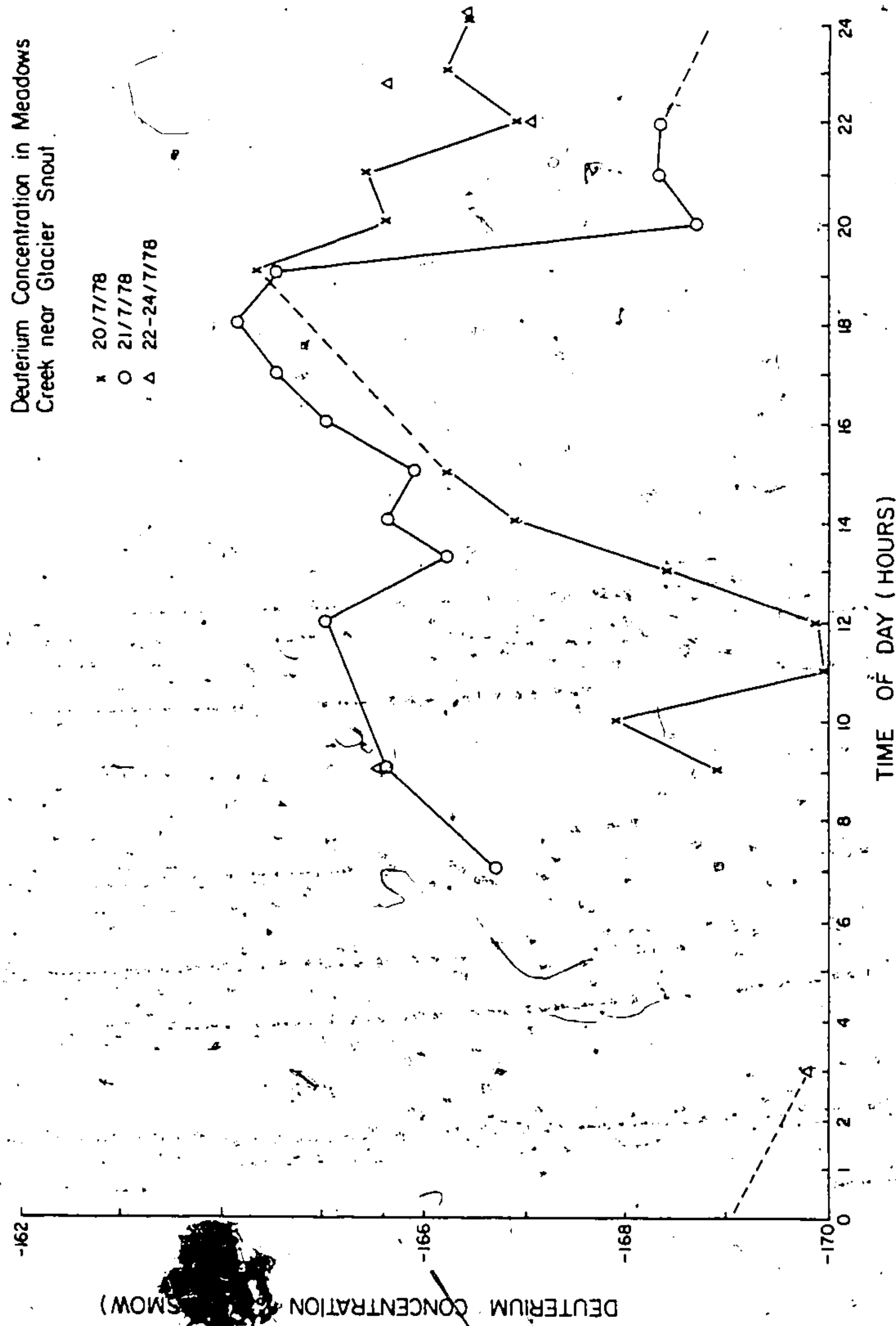


Figure 4.25. Deuterium concentration in a small glacial melt stream (1978), showing superimposed diurnal cycles

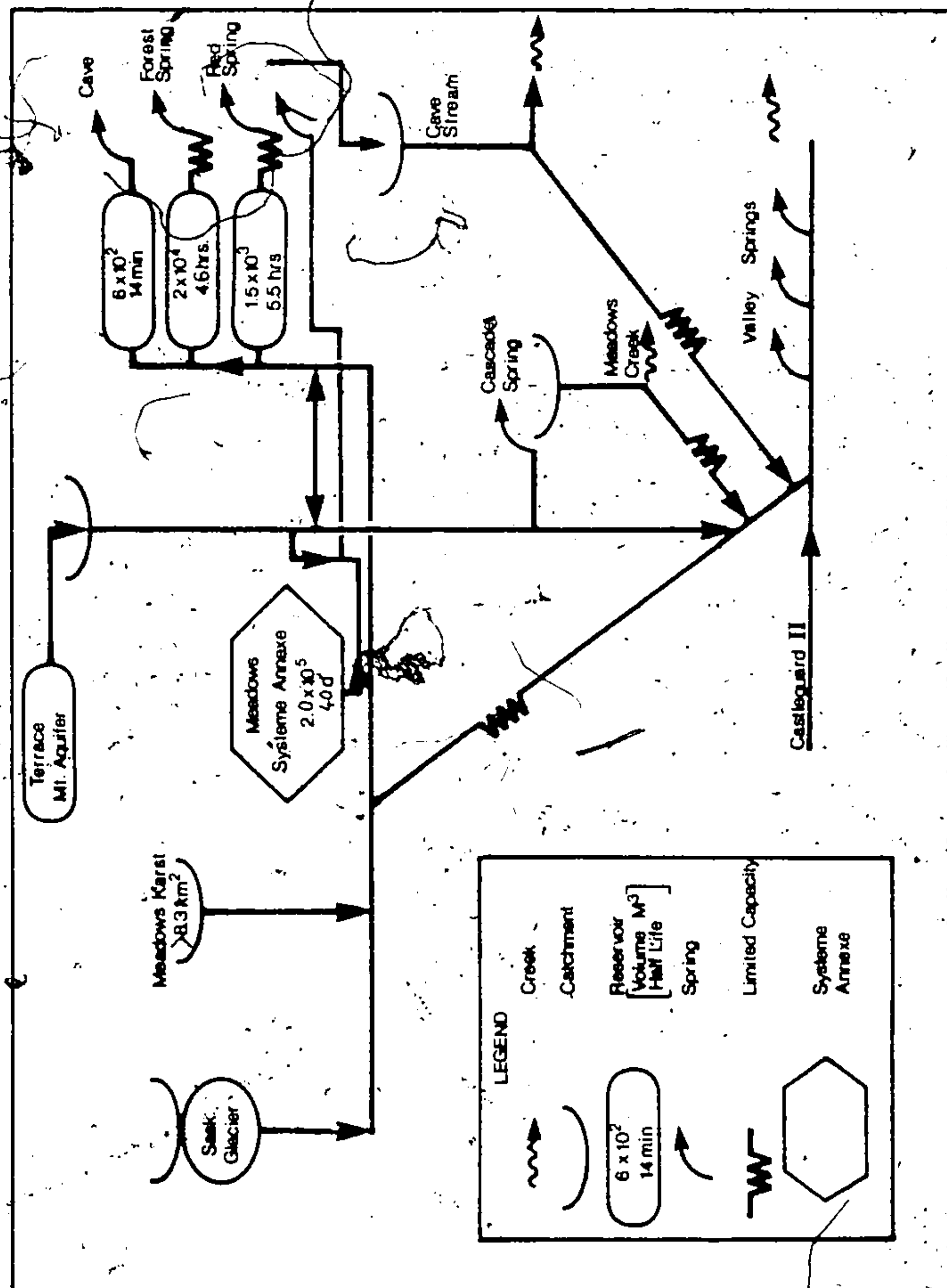


Figure 4.26. Tentative model of the structure of the Meadows Karst and its relation to Valley Springs and Creeks

CHAPTER FIVE

QUANTITATIVE FLUOROMETRIC TRACING

5.1. INTRODUCTION

There is an extensive literature on the preferred properties of hydrologic tracers (e.g. Atkinson and Smart 1980, Back and Zotl 1974, Brown 1972, Buchtela et al. 1968, Davis et al. 1980). Beyond the essential requirements of safety and detectability, choice rests largely upon the needs of the particular application and the personal bias of the reviewer.

The tracer of preference in karst hydrology is usually a fluorescent dye, because of availability, field detectability at low concentrations, and cost (e.g. P.L.Smart and Smith 1976). The red fluorescent dye, Rhodamine WT, is especially suited to quantitative applications, because it is highly detectable, has low natural background, and relatively low tendency to adsorptive loss (P.L.Smart and Laidlaw 1977). The major field problem with Rhodamine W.T. is the dependence of fluorescence on temperature (see Appendix A.4). In addition, technological advances in fluorometry mean that measurable concentrations are now so low that even small background variations or the slightest contamination can be mistaken for tracer material, unless special procedures are followed.

Field experiments are often designed to evaluate both tracers and aquifer characteristics, (e.g. Atkinson et al. 1973, Bauer et al. 1980, 1981, Behrens et al. 1981, Gospodaric and Habic 1976). The present work is concerned only with the latter problem, in particular the use of tracers to investigate the nature of the karst aquifer.

There exist a variety of investigative techniques involving combined hydrologic and tracer studies. For simple sink-rising conduit systems Atkinson et al. (1973) summarise methods of conduit volume estimation, and Brown (1973) has developed a technique of deducing simple conduit topology for sink-rising pairs based on the input and output mass balance of dye and water. The assumptions in such analyses (such as steady piston flow, constant phreatic and vadose volume during the test, and simple geometry) are probably never met in real situations, although the methods provide crude estimates of aquifer form.

The natural flood pulse methods advocated by Ashton (1966) violate these assumptions by definition, and make further assumptions concerning the identity of flood and preflood water. Atkinson (1977) and P.L.Smart and Hodge (1980) used natural and artificial flood pulses, and tracers repectively, in the Cheddar catchment, UK. They obtained similar values for the volume of phreatic storage, demonstrating the relative compatability of the methods. In contrast, Smart and Hodge also provide an example of complete attenuation of an artificial flood pulse, while Brown (1972) discovered output discharge series to be apparently leading the input!

An alternative procedure is to treat pulses as signals transmitted through a ~~filter~~ (e.g.Williams 1977), whose behaviour may be quite non-linear. The analogue input provided by measurement of swallet discharge or rainfall is usually only part of a complex recharge process and is most unlikely to represent a pulse at the upper end of a single phreatic conduit, except in trivial cases. This is reflected in the difficulty experienced in developing statistical models of karst aquifer response to rainfall (eg.Dreiss 1982, Knisel 1972). In fact, the definitive characteristic of karst floods (according to Bakalowicz and Mangin 1980) is the complexity of their composition at the rising.

Artificial tracers are modified in similar ways to natural signatures, but retain their distinctive identity. Thus, tracer recovery and distribution over time and between springs provide information specific to the route tested. However, the interpretation placed on such data appears to depend on the authors concerned as well as on the tracer data itself. In order to interpret tracer breakthrough curves, assumptions must be made concerning the reliability of the data (the significance of variations) and the nature of the intervening aquifer.

As an example of these assertions, both Atkinson et al.(1973) and Simeoni (1973, in Bogli 1980) describe polymodal breakthrough curves, which they take to be diagnostic of multiple flow paths. Initially, an inspection of the data of Simeoni shows that the multiple peaks are based on only nine data points, leaving doubts as to the significance of the peaks. In both cases green fluorescent dyes were employed, for which background may be a problem. In addition, the fluorescence of Pyranine (the tracer used by Atkinson et al.) has a strong pH dependence, so that natural pH variations may cause apparent fluctuations in tracer concentration. Furthermore, the interpretation of polymodal breakthrough curves varies: Simeoni held that branching conduits explained the dye behaviour, while Atkinson et al.suggested that a portion of the dye was stored in fissures in the limestone, before subsequently reappearing during flow recession. Mangin (1975) reported a complex breakthrough curve from a trace made during baseflow, in which a weak pulse appeared 22h after injection and lasted several days, only to be followed by a much more concentrated, but discrete pulse 6 weeks later at the onset of a major flood, presumably caused by the flushing of a "systeme annexe".

The two objectives in reviewing the above examples are (i) to emphasise the necessity of understanding specific tracer properties, and (ii) the influence of preconceived notions in interpreting tracer information There is seldom any explicit reference to this model, and an attempt is made here to develop a more explicit methodology for the interpretation of quantitative tracer information based on two complementary procedures: (i) the identification of a structural model defining likely form of the aquifer, and (ii) the consideration of a finite suite of processes acting on the dye cloud as it passes through the aquifer.

Tracer data from both quantitative and point-to-point traces are often of considerable significance, both economically and scientifically. However, there appears to be a reluctance to evaluate these traces by replication. Drew et al.(1970) based conclusions on the rather weak evidence of just two lycopodium spores recovered from a spring under investigation. Quinlan (1981) showed that catchment boundaries in the Kentucky karst may be discharge-dependent and that in order to develop reliable pollution contingency plans, repeated traces are usually necessary.

Brown (1972) attempted multiple traces of a single system producing inconsistent flow-through times. He concluded (p64) that "especially in cases involving the decomposition or dilution of a pollutant, many tests at differing discharges are necessary." Subsequent work by Kruse (1980) on the same system showed a reciprocal relationship between flow-through time and the level of the lake above the sinks. This was inferred to be the effect of progressive filling of the cave system at higher discharges, although more satisfactory interpretations are possible.

The ability to trace several flow routes simultaneously provides an important control on aquifer evaluations. Up to three fluorescent tracers may be simply and quantitatively identified in a single sample, using a filter fluorometer. Crabtree (1979) used three dye tracers simultaneously in three separate tests on a carbonate aquifer in the Mendip Hills. The resulting data allowed the definition of a conduit system feeding two springs whose anomalous behaviour had previously been inexplicable (Newson 1972, Smith and Drew 1975, Tratman 1963). This was largely possible because improved technology allowed simultaneous quantitative tracing.

Stanton and Smart (1981) ran numerous replicate dye traces at different discharges on three karst systems. The average flow-through time was inversely proportional to discharge, while dispersion increased with discharge. Dye recovery ranged from 100% at high flow to 12% at low flow. These data were inferred to demonstrate a simple phreatic system with insignificant change in conduit storage with spring discharge. P.L.Smart (1981) studied time of travel of multiple dye peaks in relation to both swallet and resurgence discharge. Low discharge breakthrough curves were single-peaked, while up to three peaks appeared at higher discharges. This was interpreted as activation of abandoned, higher level conduits by rising water levels. Significantly, the fastest flow through times suggest activation of a high-level vadose conduit, and velocities in the other two routes are constant at this time, probably because of the steady head maintained by the overflow.

The conceptual model evolving from these tracer results has much in common with that proposed in Chapter Four, in which multiple conduits operate in a systematic hierarchy.

5.2 STRUCTURE AND PROCESS IN KARST WATER TRACING

The tracer breakthrough curve is interpreted in the context of processes acting on the injected dye. The suite of possible processes may be evaluated using knowledge of the underground flow route, and the geography of groundwater recharge and discharge points.

5.2.1. STRUCTURE

Porous, fracture, and karst aquifers each have a distinctive style of flow and characteristic tracer behaviour. In the karst aquifer, exploration, knowledge of geological structure and lithology, combined with dye tracer results, provide the evidence from which the probable structure of the flow route may be inferred.

At Castleguard, surface and subsurface exploration and the hydrologic data provide contradictory evidence. The known cave is a singularly massive and discrete conduit, although its contemporary inlets have been destroyed. In contrast, the observed surface sinkpoints are very immature constricted shafts with small permanent catchments. The abundant springs and seeps in the Castleguard Valley suggest a disorganised system unable to handle its potential discharge. Aley (1963,p27) described "plateau" hydrographs such as that of the Big Spring to be "typical ... for small springs or those in immature karst areas." With an aggregate discharge of 20 cubic metres per second, and apparently free flow between springs, the present karst system appears to be neither. The rapid recession also suggests limited storage in fractured or porous media within the aquifer.

The dye traces, however, showed remarkably rapid groundwater flow, with a maximum mean velocity of 870 m/h, placing the velocity in the top two percent in the compilation of Milanovic (1976). There was also little dispersion and short-lived tracer recession; it is essentially piston flow (Calmels et al.1976). This suggests a very well-developed and integrated karst network similar to the known cave.

As a result of these observations, the structural framework of the tracer investigation consists of a system of vertical vadose shafts leading down to a well developed conduit draining from the Columbia Icefield and discharging though numerous, relatively small spring orifices. The vertical range from the Benches to the Big Spring is about 750 m, the horizontal distance 4.25 km, and it is an additional 3.5 km and 100 m fall to the Artesian Spring. Furthermore, it is known that water from the Meadows Creek (and possibly the Castleguard River) may enter the aquifer, as does the underflow component of the cave floods (see Chapter 4). The rapid flushing of the dye, and the rapid recession suggest that porous medium storage and adsorption of dye is probably negligible.

5.2.2. PROCESSES

Five classes of "process" have to be considered in quantitive karst water tracing with fluorescent dyes; those depending upon properties of the tracer, the sampling scheme, network effects, hydraulic effects, and storage (a combination of the preceeding two effects). Each class is examined here with particular reference to tracing at Castleguard.

1) Tracer Effects:

Diminution in tracer concentration which is relatively independent of the aquifer in question (such as loss by oxidation, photodecomposition, or radioactive decay) has to be considered. Although radioactive decay can be effectively modelled, in general these processes are difficult to detect and quantify. Furthermore, adsorption may be partially reversible, thus masquerading as storage of the traced water. The comparative recovery of two simultaneously injected tracers with different tendency to adsorption (but otherwise similar properties) may indicate major interference. Reversible effects such as the pH and temperature dependence of fluorescence may be removed through suitable analytical procedures and numerical processing (P.L.Smart and Laidlaw 1977).

Variations in background are more difficult to indentify and control. However, the emission spectrum of background fluorescence is usually considerably broader than that of the fluorescent dye and this may be detected with a filter combination acting well outside the adsorption and emission maxima of the tracer in question. The ratio between the value at the wavelength of interest and the offset value will be fairly constant in negative samples regardless of background levels, and will only depart from this in samples containing fluorescent dye (P.L.Smart pers.comm.). In effect this is employing a filter fluorometer as a coarse spectrofluorometer. Using a spectrofluorometer in this way to remove background, Behrens et al.(1980) report detectable levels of 0.0001 microgrammes per litre for eosine. Alternatively, Book et al.(1983) have suggested employing the high temperature coefficient of Rhodamine dyes to distinguish the tracer from less temperature sensitive, masking background.

A tracer exhibiting little or no loss or decay in application is termed conservative. In a broader sense a tracer is conservative if the measured concentration is the true concentration. At Castleguard, the rapid flow of water through discrete conduits will limit dye losses, although cascading flow in vadose shafts means a large wetted area for a given discharge, and may induce some tracer adsorption. The adsorption of Rhodamine onto dolomite laminae was observed at injection sites, which might imply significant losses in passage through the dolomites of the Stephen Formation. In general, the dyes used were reasonably conservative in the narrow sense. In the broader sense, the temperature dependence of Rhodamine W.T., and the higher background of green dyes required attention (see Appendix A.4).

2) Sampling Effects

Assuming an appropriate spatial distribution of sampling locations, there are two important effects of sampling on the form of the breakthrough curve: aliasing and contamination.

Aliasing occurs when the sampling frequency is too low to be representative of significant frequencies in the breakthrough curve. Figure 5.1 shows two breakthrough curves from adjacent springs. Bi-modal peaks are seen in the lower graph (from a continuosly sampled spring), while they have been lost on the upper graph due to an inadequate sampling frequency. Such data may be quite unsuitable for numerical processing, although in this case recognition of the problem in the field allowed its correction on subsequent traces.

Contamination may be a serious problem in high resolution field fluorometry. Fastidious cleanliness with respect to all dye handling is essential and precautions must be taken with sample bottles which have previously contained high concentration of dye. Random use of sampling bottles (and especially caps) means that most contamination error from this source will be "whitened", and thus recognisable by its low serial correlation. When estimating approximate curves from noisy data, the one tailed nature of contamination makes conventional least squares models inappropriate.

The most reliable method of avoiding both aliasing and contamination is to use flow-through fluorometry, and a variable speed recorder, so that (given such real time data) a tactical response to dye arrival is then possible. For example, the sampling interval of discrete samplers can be increased as the dye arrives. Figure 5.2 shows curves from two adjacent springs; the lower graph is from a continuous flow record, the upper one from a discrete sampler. Apart from other effects, the "signal-to-noise" ratio is clearly superior in the continuously sampled series.

3) Network Effects

The organised flow in a karst aquifer means that the conduit network is relatively simple compared to that of fracture or porous aquifers. Unfortunately, this means that karst aquifers are rather idiosyncratic, and not particularly amenable to the statistical approximations used in the latter cases. They are thus far more difficult to study.

The first network property is length, and time of travel is a measure of distance travelled and the mode of flow. Unfortunately, hydraulic variables are also important in controlling tracer velocity and must also be considered in its interpretation.

Network divergence seems perhaps a trivial interpretation of tracer appearing at more than one spring. However, it has already been used to demonstrate (Chapter 4.3.3) that cave floods are not derived from the Castleguard Valley aquifer. Furthermore, the simultaneous arrival of the Meadows dye trace at the Red and Big Spring showed independent underground routing, as opposed to the alternative hypothesis of re-entry of the Red Spring water into the aquifer (Figure 4.20). Divergence is also a demonstration of the distributory nature of the valley aquifer, a fundamental design feature of the aquifer model presented in the next chapter.

Quantitatively, the relative recovery of dye at various springs shows the relative division of tracer from the original injection site. In practice, waters are not necessarily well-mixed before separation. If this were the case, and other waters were not involved, the integral of the time-concentration curve would be equal at each spring, because each spring would then constitutes a sample of a dye injection which has been fully mixed in the parent conduit prior to distribution.

Convergence occurs when two tracer routes rejoin before emergence at the spring. The result is usually polymodal breakthrough curves, because of contrasts in the time of travel, discharge and dispersion between the alternative routes (e.g.Figure 5.1). Theoretically, the characteristics of the two flow routes are separable, but some a priori expression for the form of the breakthrough curve is needed. This has not been attempted at present. Such an interpretation of polymodal curves also undermines the use of a geometric (or harmonic) first moment of the breakthrough curve as a transit time parameter. Time from injection to the maximum concentration is used as an unambiguous time-of-travel parameter here.

Figure 5.2a is a remarkable example of a compound breakthrough curve. The first arrival time of the tracer is 0.5h after injection 3.5 km away and 700 m higher. Water is unlikely to travel at 8 km per hour underground (the previous first arrival "record", at least in Canada, being 2.33 km/hr; Brown et al. 1969). The actual recorded concentrations were well above expected background (4.727 kg of dye was injected and the given figures have been reduced to values equivalent to a one kilogram injection). Furthermore, the steady background prior to the test, and the absence of a similar early rise (despite 4.5 times greater average concentration) at another spring on Meadows Creek suggest that this is not background fluctuations. The dye employed was Lissamine FF, which is a powder. In this case, the dye was not mixed and dissolved before injection into the shaft sinkhole, (the procedure usually recommended). A portion of the dye therefore entered the aquifer as small "flocs". It is possible that some of this material may have been carried rapidly through the vadose zone by descending draughts. This "aeolian" flow route therefore connects only to the Big Spring conduit, and appears to have a very short phreatic component. Descending streams can themselves induce air movements in caves (Wigley and Brown 1976). However, such a wind is unlikely to be entrained by the descent of the Frost Pot Stream alone (approximately 2 l/s). Alternatively, if the wind is convective in origin, then a vadose exit point must exist somewhere down in the Castleguard Valley.

4) Hydraulic Effects

The explanatory potential of applying engineering hydraulics to karst water tracing can seldom be realised, because the system contains too many unknowns. For example "flood routing" degenerates into "pulse train analysis". There are two main areas discussed here; dispersion and dilution.

Theoretical models of dispersion are reasonably developed (Taylor 1954, Liu 1977), but experimental data show a need for field calibration. Stanton and Smart (1981) and P.L.Smart (1981) show that dispersion in karst conduits increases markedly with decreasing discharge, which is in accord with both theoretical (e.g.Liu 1977) and empirical work (e.g.Beltaos 1982). In general, dispersion increases with distance, time of travel and the roughness of the channel. However, Brady and Johnson (1981) emphasise that dispersion is strongly reach- dependent, which is why only very vague approximations of flow route geometry can be estimated from dispersion in karst applications. As an example, Atkinson et al.(1983) consider that the friction factor (f) should be around 0.33-0.90 in the well-developed conduit like Castleguard I.

In practice, values of 0.87-2.31 were found, and attributed to constrictions and contortions in the air flow route. The enormous friction factors (65-140) calculated by Atkinson (1977) for an unexplored conduit, must result from passage collapse or other such occlusions.

One explanation of the usually skewed form of breakthrough curves is storage in and release from "dead zones" in the channel (Nordin and Troutman 1980, Thackston and Schnelle 1970, Valentine and Wood 1977). Thackston and Schnelle provide an empirical relationship between the volume of dead zones, and the friction factor (f), and show that it is possible (but analytically complex) to estimate the volume of dead zones from the measured dispersion. Atkinson et al.(1983) have emphasised the need for estimates of f in karst aquifers and tracing may provide a further means of making such estimates. Similarly, Brady and Johnson (1981) provide equations for estimating hydraulic radius from dispersion studies. In the present situation, however, the compound structure of the flow route would invalidate any such calculations. However, "dispersion factors" may be calculated for comparative purposes, because they reflect mixing processes in the traced channel, which in turn give clues to the "style" of the flow route; discrete conduit, irregular conduit, or broken anastamosing channel. Such inspection of (natural) tracer dispersion provided the qualitative criterion for the selection of isotope mixing models in Chapter 4.3.3.

Dispersion of the tracer in an unbranched conduit will reduce the maximum concentration with the inverse of square root distance (Church and Kellerhals 1970), although the time-concentration integral will remain constant with distance travelled. Another source of reduction in tracer concentrations will be streams tributary to the tracer flow route, and they will proportionally decrease the time-concentration integral. Dilution sources at Castleguard will be streams in the vadose zone, the major flow from the Icefield and contributions from elsewhere in the aquifer, e.g. the Cave underflow. In addition, the Meadows Creek, Cave Stream and possibly the Castleguard River may lose water into the aquifer at certain times.

The diurnal character of the discharge regime at Castleguard reported in the previous chapter here becomes a major constraint. Much of the above discussion may only be relevant under conditions of steady or gradually varying flow. The varying contribution of water will not only vary dilution of a tracer, but will alter aquifer configuration during the course of a trace. Under some conditions such effects may be valuable (see below), but in a poorly defined aquifer the main result is an increased indeterminacy.

Unfortunately, the highest frequency variable flow, the surging of the Big Spring, did not occur during a continuously monitored trace, as this may have provided interesting information concerning the surge mechanism.

5) Storage and Delay

The delivery of a flood pulse through a karst system takes place at a faster rate than the delivery of water initially labelling the pulse. In free surface streams the celerity of these two impulses has been studied (Brady and Johnson 1981, Buchanan 1967, Glover and Johnson 1974) and is predictable, given morphologic data and empirical calibration.

In pure pressure conduits the celerity is increased by the rapid transmission of the flood impulse (Ashton 1966). However, in practise the flood pulse will be initially employed in filling storage and will thus bring about a storage-induced delay (Palmer 1981a). The nature of this delay is considered more fully in the next chapter. However, two end-member forms of storage can occur in karst aquifers; inline (throughflow) storage, and off-line (in-out) storage (Schotterer et al.1979).

In-line stores are more or less enlarged channels which receive input at one end and discharge from the other. In karst they may act as simple delay mechanisms and simply be a loop in a branching pipe network (e.g.Milanovic 1981, Simeoni in Bogli 1981), or else intermittently active conduits holding tracer until floods drive it out (e.g.P.L.Smart 1981, Mangin 1975).

Off-line stores are those which require a flow reversal to change from recharge to discharge, and are akin to bank storage in a river (e.g.Atkinson et al.1973). The systemes annexes of Mangin (1975) are composite stores recharged both from the vadose zone and backflooding conduits.

The two end-member stores may be identified if a flood wave of labelled water passes through the aquifer. An off-line store will be recharged by the flood pulse and will drain on the recession, providing a tracer pulse during falling stage. An intermittent in-line store will be recharged by the flood, but any trapped dye will only be driven out in subsequent floods. At higher flows such a store becomes a second flow route. Figure 5.3 shows a set of dye traces to the Big Spring where such a sequence may be inferred. However, other replicate traces produced single peaked breakthrough curves.

The character of the storage medium may be reflected in the form of the secondary pulse. In Figure 5.3, the release is fairly discrete, supporting the notion of conduit storage. Figure 5.4 is from a trace into a heavily choked sinkhole from which dye was released only slowly by flushing. The result is an unusually strong tail to the trace.

It is clear that technological advances have dramatically increased the quality and quantity of information obtainable from fluorometric tracing. The potential for artificial tracer application in karst is very much greater than the foregoing review might suggest. However, most practical applications will never realise this potential, because of the indeterminate nature of the flow route geometry and unsteady flow conditons. The advantage of a structural-process approach is that assumptions become explicit. It provides a framework in which "likely" conditions can be envisaged and the tracer data evaluated accordingly, using appropriate rather than an abritrary set of processes.

5.3. QUANTITATIVE TRACING IN THE CASTLEGUARD AQUIFER

5.3.1. INTRODUCTION AND RESULTS

The overall pattern of traces in the Castleguard area was reviewed in Section 4.1 and Figures 4.1-4.3. Quantitative traces were made from the south Benches to a few of the springs in the Castleguard valley (see Section

3.2.4 and Tables 3.4a,b and 3.5).

The flow routes are inferred to be constricted vadose shafts of limited discharge, leading down to a major conduit level which is more or less phreatic. The conduit discharges as a series of relatively immature springs.

The breakthrough curves demonstrate very rapid passage of tracer with little dispersion (except for an injection into a choked sinkhole). The tests covered a wide range of discharge conditions, but were complicated by the markedly unsteady conditions, especially for springs in the overflow condition.

Table 5.1 summarises the tracer breakthrough information for each trace. Columns are as follows:

1) Reference name: From-number-To; see Table 3.4a and b.

2) Area: the area in parts per billion-hours of the time concentration breakthrough curve. Estimate includes exponential extrapolation unless stated otherwise.

3) Peak: the maximum tracer concentration calculated for the quadratic curve fitting the three points around the maximum measured concentration.

4) Time of travel of tracer from time of injection to peak concentration (the least ambiguous measure, given polymodal

breakthrough curves, and variable tails).

5) Q: Discharge; "Estimated" assuming perfect mixing, inversely proportional to Area. "Observed" is the estimated total spring flow, using total measured spring flow plus 2.5 cubic meter per second (2.0 for trace 13) as a conservative estimate of unmeasured underflow.

6) Disp. is the dimensionless variance of the breakthrough curve (Brady and Johnson 1981) which is equal to to the dispersion coefficient divided by tracer velocity and distance. The actual values of both the latter are unknown.

7) Recovery is the percentage of the original dye estimated to have passed through each spring. Calculated as the integral of discharge and concentration with respect to time. For Big Spring discharge measurements are available. Assuming the Tangle Spring trace to be representative of the recovery from Meadows Creek Springs, recovery may be estimated for this spring group. For Watchman, Artesian and the Castleguard River Springs discharge estimates of 0.5, 1.0 and 1.5 cubic meters per second are used respectively.

The traces MD12BS and MA13BS have missing data due to machine malfunction; the former trace in particular is quite unrepresentative. Some curves were sufficiently documented so as not to need exponential extrapolation, in others the tail contribution by extrapolation was excessive. In the latter case estimates without extrapolation are provided where useful.

The estimated discharges are generally greater than the total observed discharge, which implies either large undiscovered springs, or preferential routing of the dye to unsampled springs. The variability in discharge estimates between springs and the poor recovery at high discharges show incomplete mixing in the aquifer, and supports the latter hypothesis, in which sampled sites are unrepresentative, being over-diluted. At low discharges Tangle Creek Springs (FP11TC, MA13TC) give estimates very close to observed values, implying that local dilution has ceased.

The 1979 and 1980 Big Spring traces are shown with discharge series in Figures 5.5a,b and 5.6a,b. Most of the 1980 traces are shown in Figure 5.7a,b. (Note the concentration scales with care).

5.3.2. VARIATIONS BETWEEN SPRINGS

All springs sampled in the Castleguard Valley were positively traced and they may thus be regarded as a distributary group. However, relationships between springs were unequal and varied with discharge and injection site.

Similar breakthrough curves suggest that springs are discharging from the same conduit, in which case the area under the breakthrough curve should be the same, and differences in the peak value, dispersion and travel time will reflect different distances of travel.

In trace 10 (Figure 5.8) the highest overall resolution was obtained revealing a bimodal breakthrough curve at four widely separated springs. This characteristic was clearly gained before distribution to the various springs, but has remained despite subsequent modification of the tracer cloud.

The Castleguard and Big Spring traces appear to have a close relationship when sampled correctly. The Watchman Spring (Figure 5.2) is in the same immediate family, but has been diluted by some other water source. The irregularity of this record may reflect incomplete mixing of these waters.

The Artesian Spring and Tangle Spring are clearly distributaries with similar characheristics, but the relationship breaks down under low discharge (Figure 5.9). Inspection of Figure 5.7b suggests that this may also be the case for the Castleguard River Springs. In fact, the Big Spring was dry during trace 13 and recovery was from

"Gravel Spring" nearby, so it can only be stated that the Gravel Spring was somewhat dilute compared to Castleguard Spring.

Traces to Artesian Spring also showed long tails which suggests some dead-zone storage. The difference in arrival time between Tangle and Artesian Spring is 6.1 h, and their separation is 2 km, giving a minimum velocity of 0.3 km/h. In contrast, the Big Spring to Castleguard Spring velocity is 0.61 km/h. The residual head on the Artesian Spring suggests that it drains an immature, confined conduit with a relatively high effective roughness, and some lateral storage.

The relationship between the "Big Spring group" and "Meadows Creek group" is more variable, depending on discharge and injection point. In general, the Big Spring is more concentrated and has the larger mass recovery under high discharge, but this effect is reversed under low flow, or when injection is from Frost Pot (e.g.Figure 5.10).

5.3.3. VARIATIONS DEPENDENT ON INJECTION POINT

The result of injection into a choked sinkhole (Figure 5.4) is a clear example of control by injection site. This result is "external", however, compared to other effects.

The polymodal breakthrough curves (Figure 5.3) were obtained from four out of seven well-defined traces from Met Sink. The effect was seen at all springs monitored, which suggests a division in the associated shaft system before entry into the Castleguard II conduit. This may occur at the injection site itself where water sinks at several points. However, polymodal curves are not alway obtained. In part, this reflects sampling frequency, but Figure 5.5 shows that polymodality only occurs during periods of strong diurnal flow variability. This is in accord with the "in-line storage" hypothesis described above, although inspection of the last two traces of 1979 (Figure 5.5) shows the curves are somewhat more complex, and may reflect an off-line component also.

The unimomodal traces from Met Sink occur when spring flow is relatively steady, suggesting that the polymodal curves may be an unsteady flow phenomenon, induced by interaction between the shaft and the conduit. For example, if head is rising rapidly in the main conduit the water level may rise in the shaft system. If shaft discharge is insufficient to keep pace with rising water levels, a net flow will occur into the shaft from the conduit system, preventing the tracer from continuing to enter the conduit. As a result, a tracer hiatus will be observed in the tracer concentration curve at the spring.

The longer delays between pulses are harder to account for, and it may be that the polymodality has multiple causes.

Both traces from Frost Pot (traces 7 and 11) were preferentially routed to the Meadows Creek. This was also the case for traces made from Midway and Main Sinks (traces 12 and 13). The latter probably constitute a discharge related phenomenon, because both injection sites were close to Met Sink and on adjacent fractures.

5.3.4. VARIATIONS WITH DISCHARGE

The relationship of tracer behaviour to discharge is difficult to assess, because the appropriate measure of discharge is not obvious. Alternatives are the input discharge, total spring flow, or individual spring flow, although only the latter values are really available. There is also a choice of tracer parameters available, such as various travel time estimators, dilution parameters, dispersion and recovery.

The relationship between time of travel and discharge has been shown to be of morphological significance (e.g.Brady and Johnson 1981, Glover and Johnson 1974, P.L.Smart 1981). In this case, an appropriate time of travel estimate is time of peak concentration after injection. This avoids the complications of more sophisticated parameters caused by polymodal breakthrough curves, extensive tails, or high backgrounds. Figure 5.11 shows travel time in relation to average Big Spring discharge over the days of dominant breakthrough. The seven Met Sink traces were fitted to a power function, but the relationship proved weak (r-squared=0.63, insignificant at the 0.01 level) and the slope was considerably less than the value of -1 expected in a phreatic system (P.L.Smart 1981). The implication is that travel time is controlled largely by the vadose component of the route, in which case, inflow discharge at the sink will be the most important control. Unfortunately, not only are inflow data unavailable, but the high variability of this inflow (see Figures 4.7 and 4.8) means that unsteady flow phenomena are probably controlling tracer behaviour more than discharge per se. Furthermore, the underflow-overflow spring hierarchy means that the Big Spring discharge is also an inappropriate parameter.

Figure 5.12 shows travel time against total measured spring discharge, which is probably an underestimate of actual spring flow by between 2 and 5 cubic metres per second, most of which is derived from underflows and therefore relatively steady. The results depend upon injection site, but discharge response is quite inconsistent, for while Big Spring travel times gradually increase as discharge falls, Tangle and Artesian Spring decrease to a minimum and then remain relatively steady. In part, inflow behaviour is controlling these data; e.g.the inflow discharge was a minimum for trace 12, although spring flow was maintained by other sources at this time. Trace 12 is considered anomalous in this respect and may be tentatively rejected.

The major problem is accounting for the different response of travel times to increasing discharge. The Big Spring behaviour is roughly that of a vadose system, which is compatible with its overflow status. It is also expected that times of travel will decrease with increasing discharge in the underflow (e.g.P.L.Smart 1981), but this is not borne out by the data.

Although several distributary configurations can be envisaged which limit the rate of increase of velocity with discharge in one conduit compared to another, none are able to account for a decreased velocity at high discharges. Table 5.1 shows considerable dilution of the Meadows Creek Springs, which may partly result from incomplete mixing before distribution. However, it is also known that exotic (unlabelled) water is discharging from Meadows Creek Springs (Chapter 4), having been derived either from sinks in the bed of the creek, or as underflows of the cave floods, or both. If this component is entering the Meadows Creek conduit, then it will be competing with flow from the main conduit. The hydraulic gradient of the distributary may therefore be reduced by the head of the exotic system (Figure 5.13). In contrast, when the external source is inactive, the hydraulic gradients may favour discharge to Meadows Creek, as the tracer data appear to suggest.

Figure 5.14 shows the comparative tracer recovery of the different springs in relation to total measured discharge. The graph is influenced by changes in spring discharge as much as by changes in dye distribution. Thus in general, underflow springs will have a decreasing overall significance as total discharge rises (e.g. the Tangle Spring). A dimensionless plot of tracer recovery against the proportion of total spring discharge is an improvement, but difficult to interpret (Figure 5.15). However, the ratio of the previous two variables may be plotted against total spring discharge. In this case, the ordinate is dimensionless and is defined as

$$\left(\int C_i\,Q_i\,dt\,/\,M\right)/\left(\int Q_i\,dt/\Sigma\int Q_i\,dt\right)$$

where C is concentration

Q is spring discharge

M is the mass of tracer injected

While roughly proportional to the integral of the time-concentration curve, these values are corrected for system discharge. In Figure 5.16 they are plotted against an estimate of total spring discharge (obtained by adding 2.5 cubic metres per second to measured discharge (2.0 for trace 13) based on field observation). For a perfectly mixed, distributory system values should lie along the 1.0 line. Departure from 1.0 and separation of data points in a particular trace reflect uneven tracer distribution and/or dilution by exotic (unlabelled) water.

Several points are raised by this interesting graph, (bearing in mind that recovery is estimated from the total CQ integral including tail estimation errors):

1) The close relationship between Big and Castleguard Springs is apparent, although trace 13 suggests additional dilution of the Big Spring similar to that of Watchman Spring in trace 11. Trace MA13BS was recovered from the Gravel Spring and appeared to have suffered dilution by unlabelled water in the alluvial aquifer.

2) The similarity of the Tangle and Artesian Spring results justifies their joint assignment as the Meadows Creek Group.

3) Traces 8 and 11 were from Frost Pot, and demonstrate the preferred routing to Meadows Creek springs from this injection site.

4) As discharge falls, so a disproportionate quantity of tracer is routed to Meadows Springs (ratios above 1).

5) If some springs show a deficiency in tracer recovery in perfectly mixed system, then this should be compensated for by excess tracer recovery elsewhere. This is not the case and incomplete mixing is inferred.

5.4. THE CASTLEGUARD VALLEY AQUIFER

The tracer results allow a preliminary, simplified model of the aquifer to be developed. The elements are listed below with a brief summary of available evidence.

5.4.1. THE SHAFT SYSTEM

For Met Sink, four out of seven well defined traces produced bimodal breakthrough curves which may be indicative of a dual flow route in the vadose zone.

Frost Pot shaft also appears to have two components; one taking water to the Castleguard II conduit, the other possibly conveying air to the Big Spring conduit downstream of the divergence to Meadows Creek.

The Met Sink and Frost Pot shafts enter Castleguard

II on opposing sides, a comparatively short distance upstream of the divergence. In this way, dye always enters both distributaries, but is not completely mixed at the junction.

5.4.2 THE CONDUIT SYSTEM

The division into the two distributaries is probably phreatic, because both routes always receive dye, regardless of discharge.

The conduit feeding the Big Spring per se is an overflow and non-operative at low discharges. The decrease in dye delivery to the Castleguard River Group is thus partly a result of changed system configuration. The low dye recovery from Big Spring under full flow may show preferential routing to unsampled overflows.

Figure 5.16 shows a dilution effect on Big Spring at all discharges. Whether this is discharge from an independent karst aquifer, or an alluvial aquifer is not known. The situation appears similar to that in the Muotatal (Behrens et al 1981) where two karst aquifers flank an alluviated valley. Tracing has shown the numerous springs in the valley bottom to be various mixtures of karst and alluvial water. At Castleguard, the alluvium extends only a very short distance up valley, and the karstic Watchman Spring also exhibited dilution, so an independent flanking karst aquifer may be hypothesised.

The Meadows Creek branch of Castleguard II also encounters some exotic dilution water, which appears to limit discharge and thus tracer velocity and concentration. The lagged diurnal pulses in the Meadows Creek (II) (Figure 4.13) may stem from this contribution.

Under low flow, the competition relaxes, dilution of Tangle Spring ceases, but continues at Artesian Spring. This dilution water may be a lower level remnant of the more general dilution source operating at greater discharges. Meadows Creek bed losses, or simply the contribution of a semi-autonomous fracture flow aquifer.

Figure 5.17 is an attempt to combine the above observations into a conceptual physical model of the conduit system. However, note that the above discussion draws heavily on virtually every anomaly, as well as on consistent results. While employing the latter phenomenon has a certain scientific credibility, interpretation of singularities is at best the practice of speculation. However, the general structure conceived here may be employed and examined in the following chapter on aquifer modelling.

5.5. CONCLUSIONS

Advances in technology have meant that high precision, quantitative fluorometric tracing can now be undertaken in relatively remote areas. Considerable demands are made on the operator, however, both in terms of the understanding of the techniques and materials, and in the industry and care which such a program necessitates.

The improved quality of tracer information has created a need for suitable interpretive tools. Bearing in mind previous and contemporary work, a crude attempt has been made to develop such tools using a structure-process approach. The probable structure of the traced route must be determined using geologic, geomorphic and hydrologic information (including tracer data). Given the layout of the aquifer, five groups of controlling process are recognised; those characteristic of the chosen tracer, sampling scheme, the effects of conduit network, hydraulic processes, and storage within the aquifer. This approach is applied with some success to the description of the form and function of the Castleguard Valley aquifer. This site proved suitable for such an investigation, having brief flow-through and tracer purging times, but it also suffers from being an exceptionally complex aquifer experiencing a wide variety of rapidly varying discharges.

Although the potential for breakthrough curve interpretation in karst aquifers is very large, it is unlikely to be realised, because only rather meaningless average flow route properties can be estimated. Furthermore, karst aquifer behaviour is often flow dependent so that numerous replicate traces are necessary in interpretive studies. The indeterminacy introduced by unsteady flow phenomena further limits possible applications.

These problems were very much in evidence at Castleguard, and, in addition, great difficulty was had in obtaining anything approaching experimental control over the investigation. As a consequence, the aquifer model developed is highly speculative.

REF.	AREA (ppb.h)	PEAK (ppb)	PEAK.T. (h)	Q(m3/s) EST.	Q(m3/s) OBS.	DISP.	RECOVERY (%)	NOTES
MS1BS	(4.47	0.898	5.43	62.2	5.02	.0117	8.2)	2,3,4
MS2BS	12.6	2.75	13.7	22.1	2.94	.0126	13.1	2,4
MS3BS	10.5	1.95	15.0	26.6	1.35	.0140	5.0	2,4
MS4BS	11.6	1.43	17.7	24.0	1.11	.0214	4.3	2,4
MS5BS	22.8	3.84	11.6	12.2	.91	.0524	6.0	4
MS7AS	(20.4	0.821	14.8	13.6	11.30	.161	21.5)	5,1
FP8BS	8.31	3.27	4.88	33.4	11.30	.0058	13.5	2
FP8AS	10.9	1.55	11.8	25.5	"	.384	16.6	5
MS9BS	15.2	3.72	5.80	18.2	10.57	.0148	25.2	2
MS9CG	18.0	3.51	8.52	15.4	"	.0239	9.7	5
MS9TC	3.76	1.02	8.19	74.0	"	.0137	5.5	2
MS9AS	4.44	0.640	12.1	62.5	"	.0283	1.6	5
MS10BS	23.1	5.62	5.13	12.0	9.28	.0122	28.2	2
MS10CG	22.5	4.52	6.52	12.4	"	.0273	12.2	5
MS10TC	4.61	1.13	8.19	60.3	"	.0129	5.9	2
MS10AS	6.47	0.769	10.6	43.0	"	.0425	2.3	2,5
FP11BS	(20.3	3.21	7.90	13.7	6.70	.338	10.9)	1
" "	13.7	"	"	20.2	"			2
FP11WS	3.81	0.976	8.31	72.8	"	.0062	0.7	2,5
FP11TC	60.2	12.3	9.86	4.62	"	.182	49.7	
MD12BS	(6.49	0.731	15.3	42.8	5.37	.0432	1.8)	2,3
MD12TC	54.2	4.27	16.5	5.13	"	.0406	38.3	1
" "	47.0	"	"	5.92	"			2
MD12AS	(60.76	1.41	21.7	4.57	"	.276	21.9)	1,5
" "	18.8	"	"	14.1	"			2
MA13BS	(27.6	4.53	8.62	10.1	4.12	.0400	2.8)	3
MA13CG	40.7	6.48	9.14	6.82	"	.0529	14.8	5
MA13TC	110.	11.6	9.34	2.56	"	.491	67.2	

Notes:
1. Large exponential contribution from tail.
2. Estimates without exponential extrapolation.
3. Missing data.
4. Observed discharge is only Big Spring
5. Discharge estimated for recovery calculation.

Table 5.1. Parameters of breakthrough curves. The observed discharge is a conservative estimate of total springflow. Parentheses enclose doubtful data.

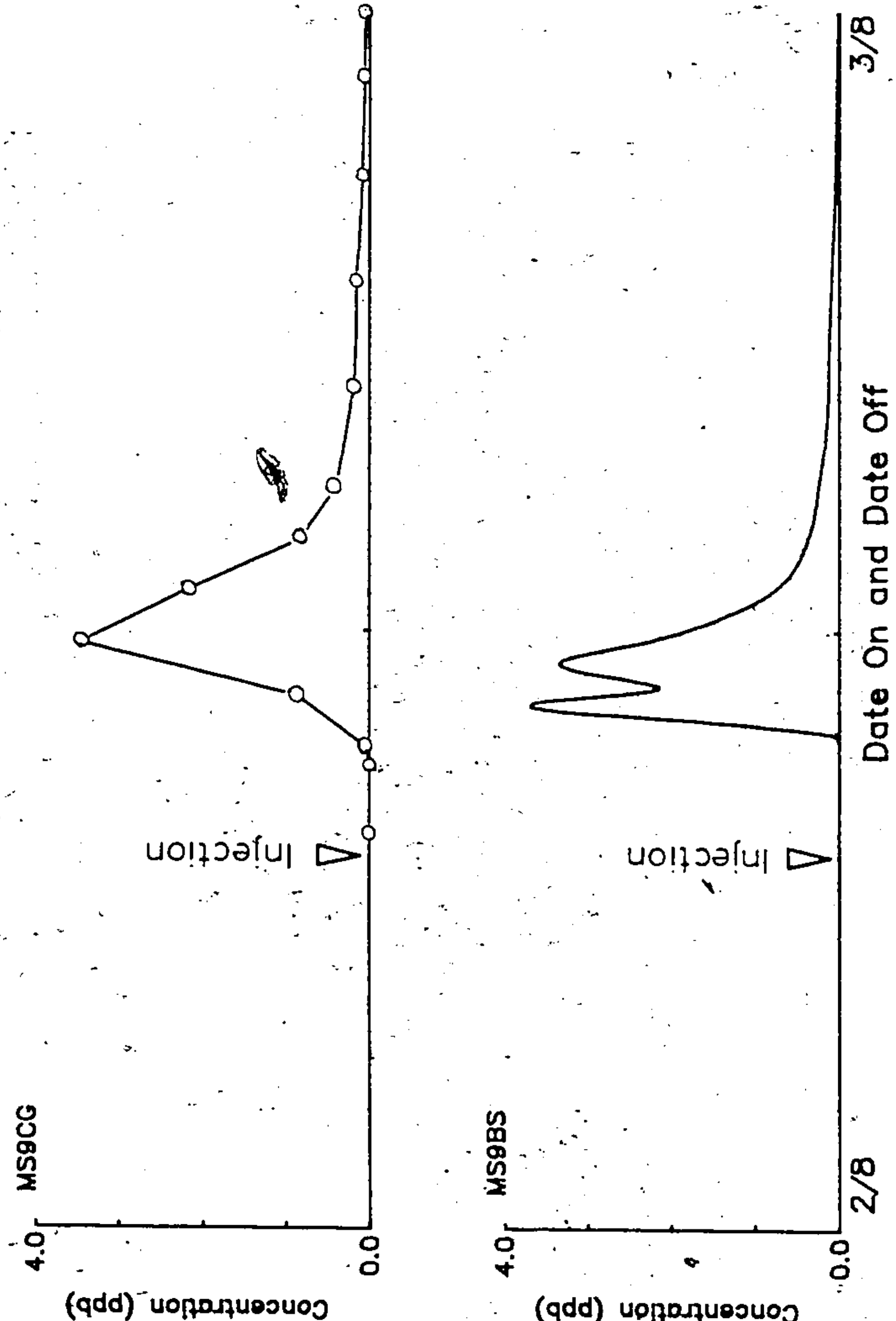


Figure 5.1. The effect of aliasing on the form of the breakthrough curve

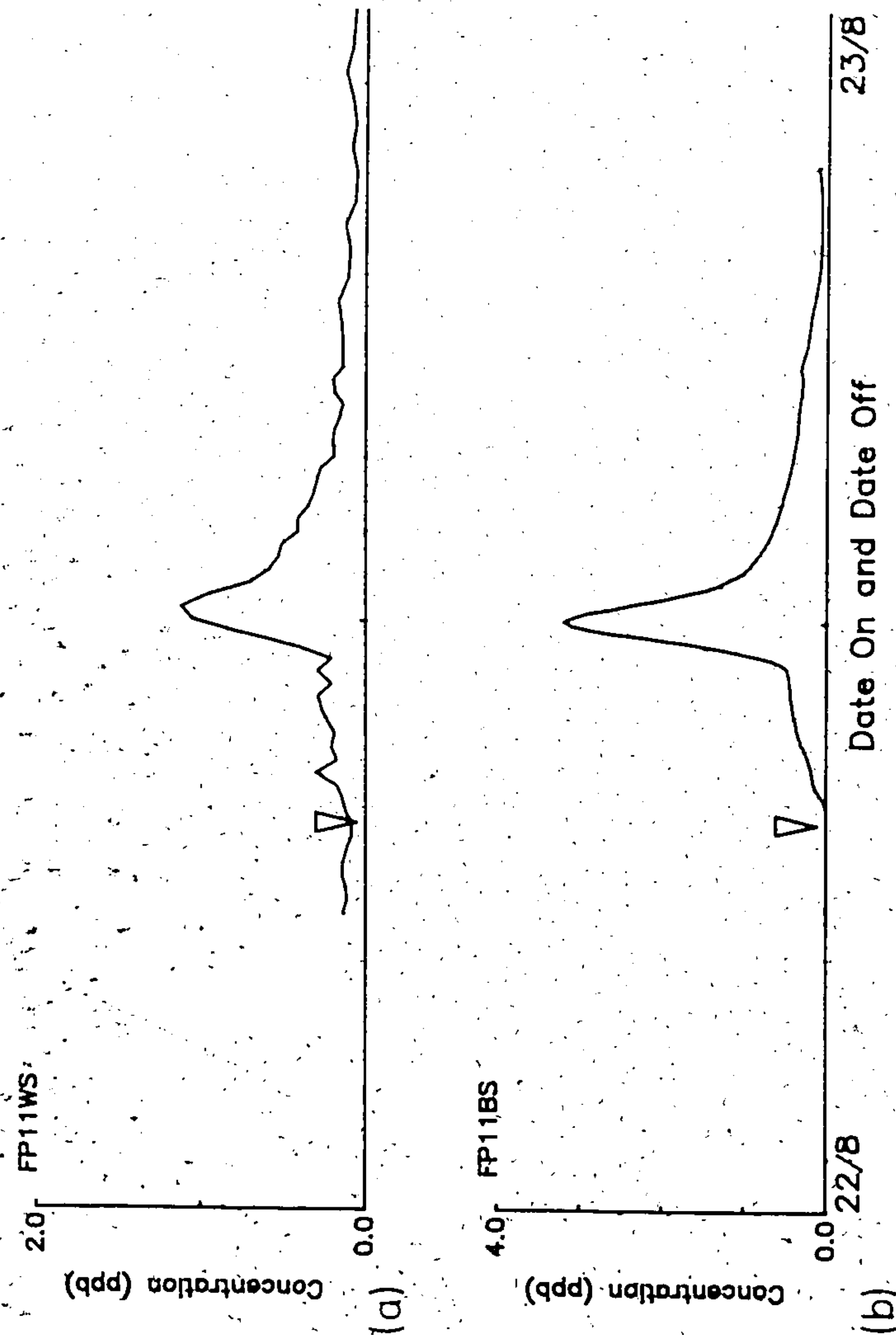


Figure 5.2. The effect of contamination on the clarity of the breakthrough curve

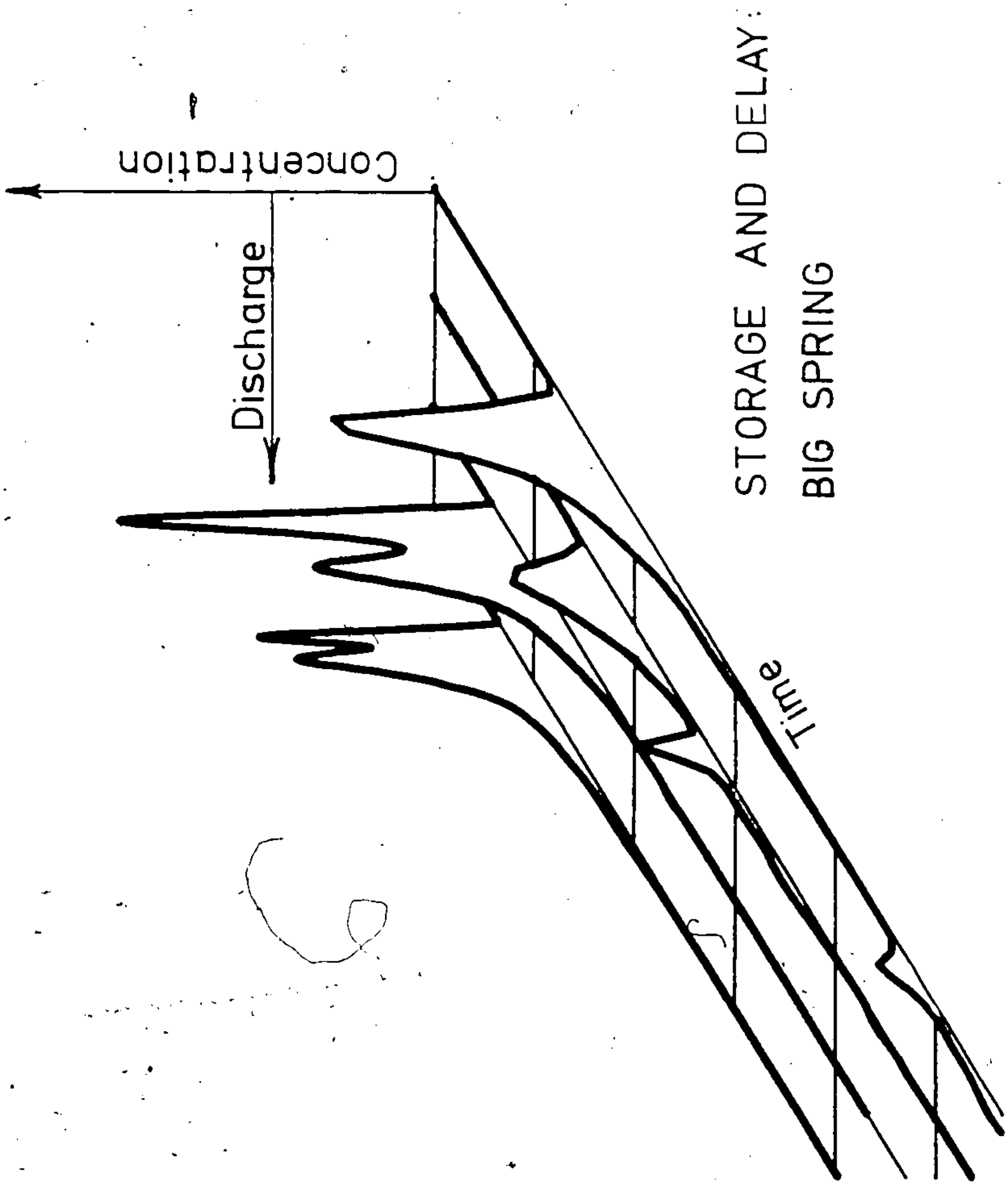


Figure 5.3. The gradual separation of a second flow pulse into "In-line" storage as discharge falls

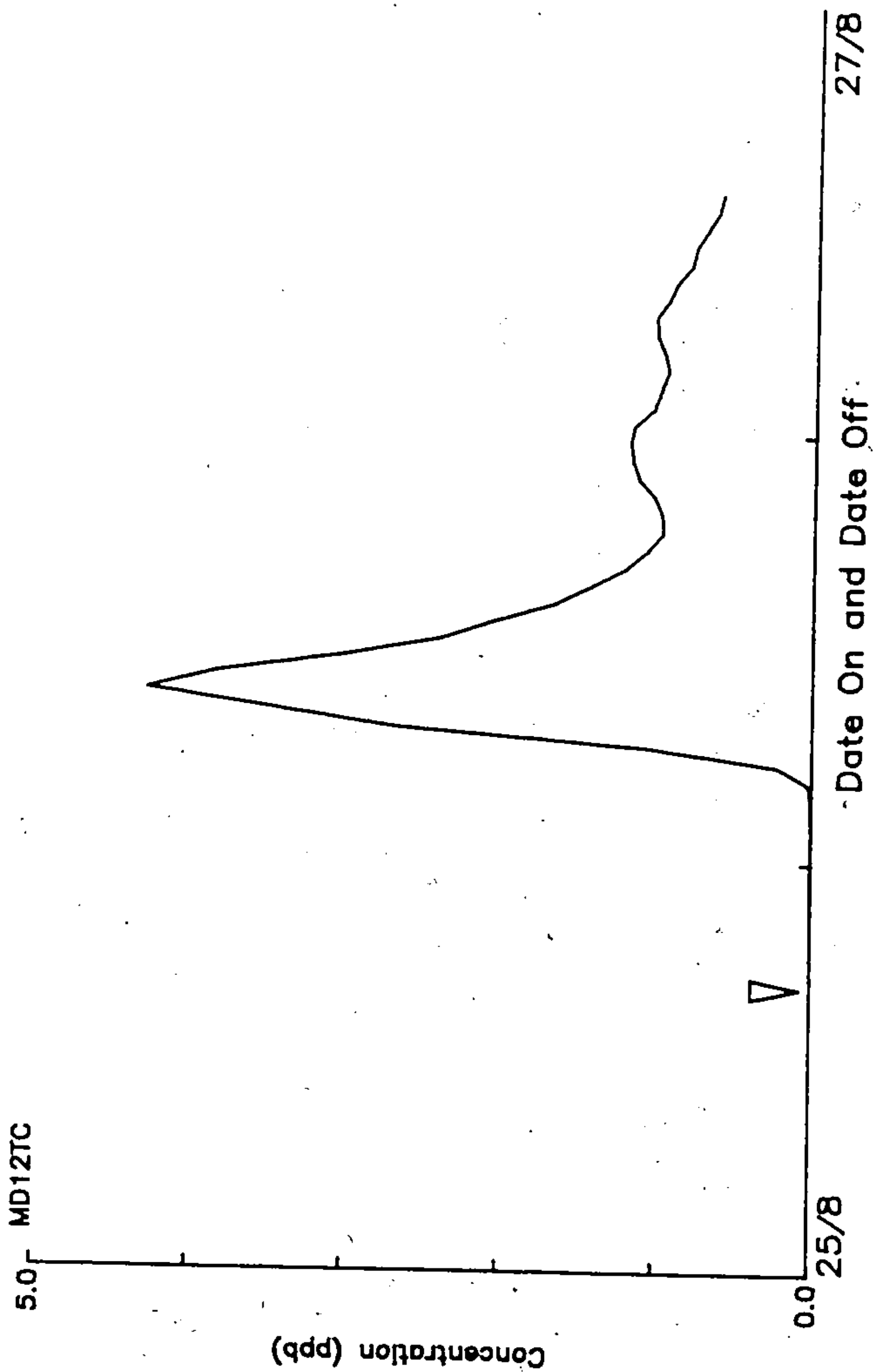


Figure 5.4. An extended tail caused by dye storage in a choked sinkhole

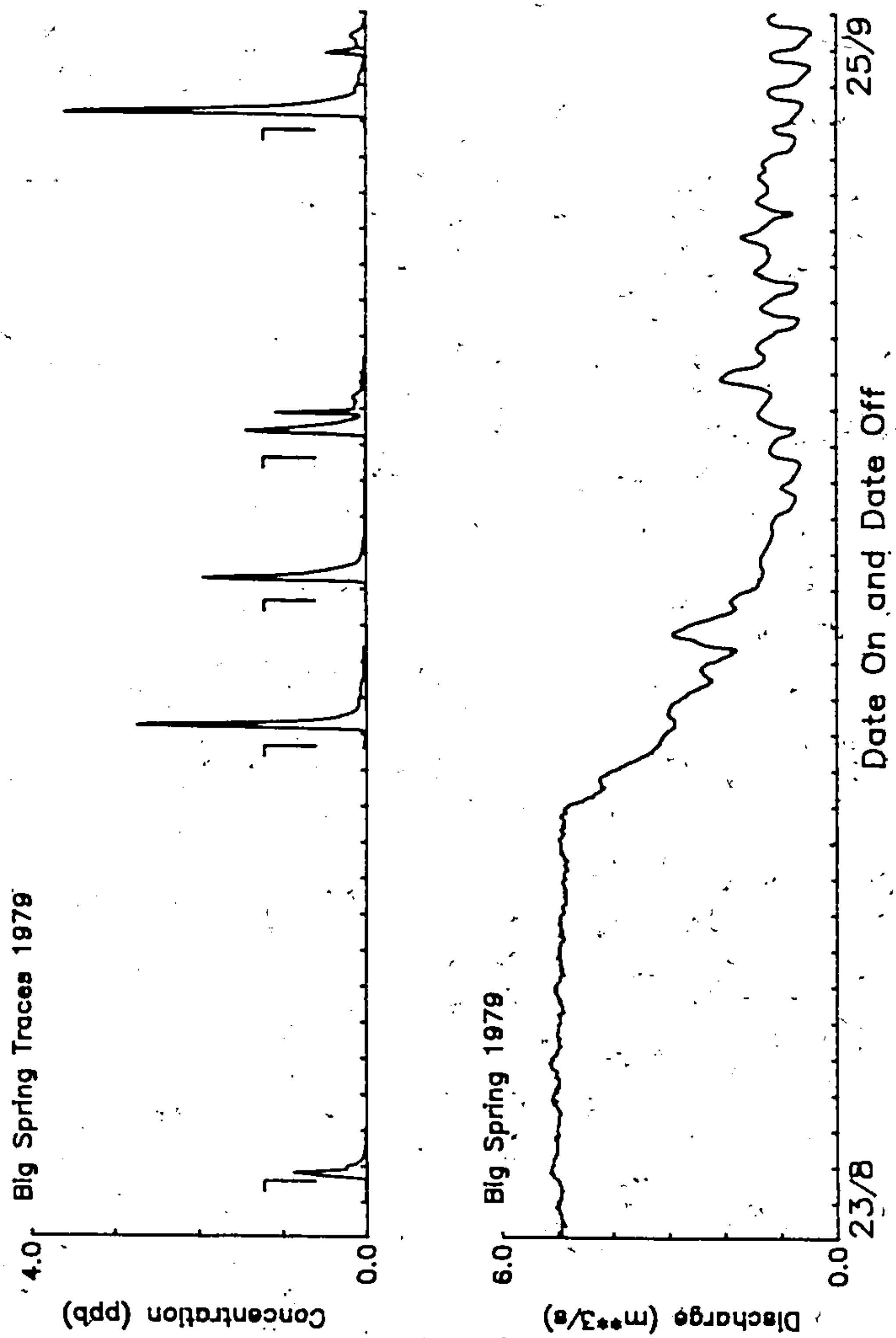


Figure 5.5. Big Spring Traces 1, 2, 3, 4, 5 1979 with spring discharge

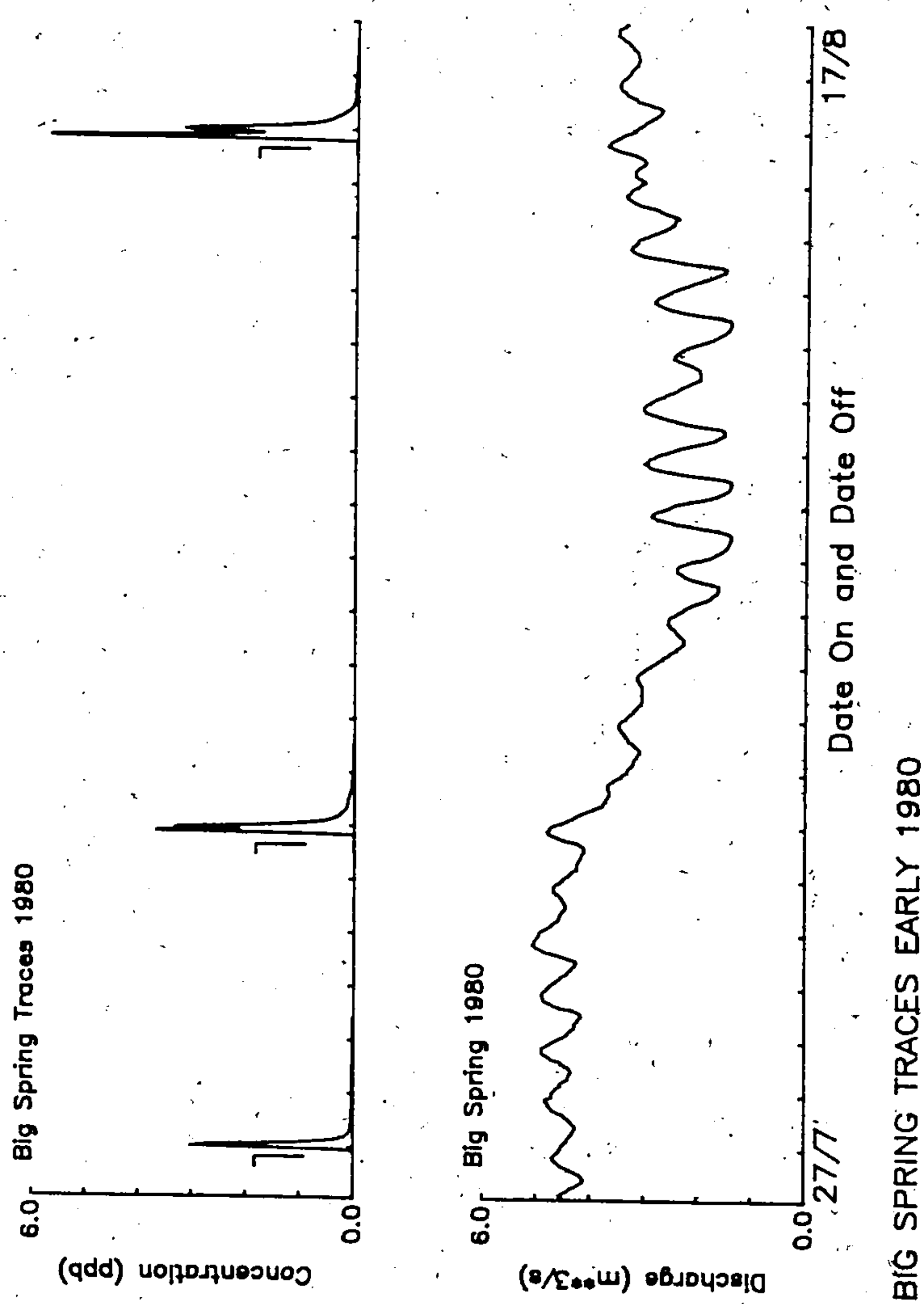


Figure 5.6a. Big Spring Traces 8, 9, 10 1980 with spring discharge

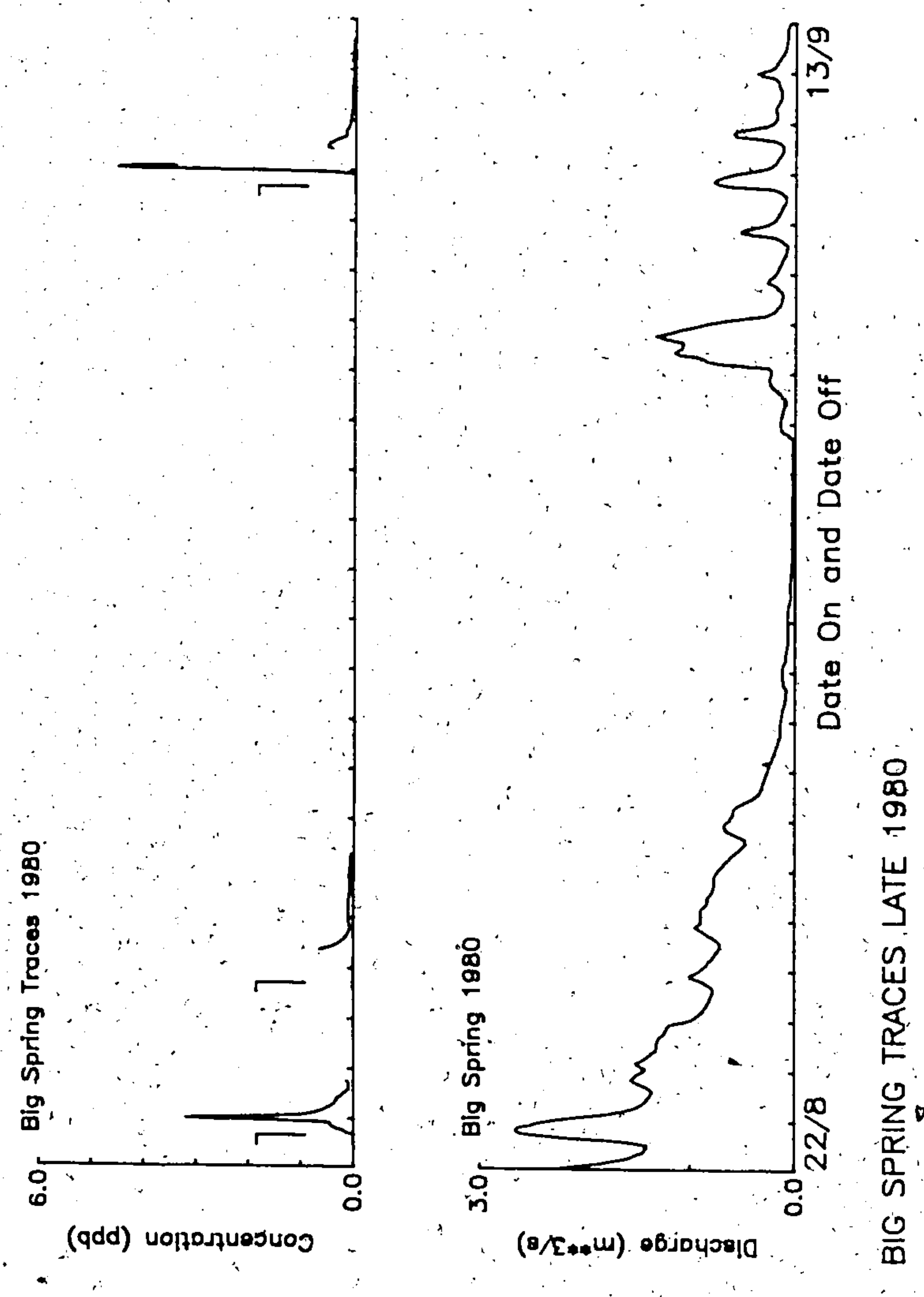


Figure 5.6b. Big Spring Traces 11, 12, 13 1980 with spring discharge

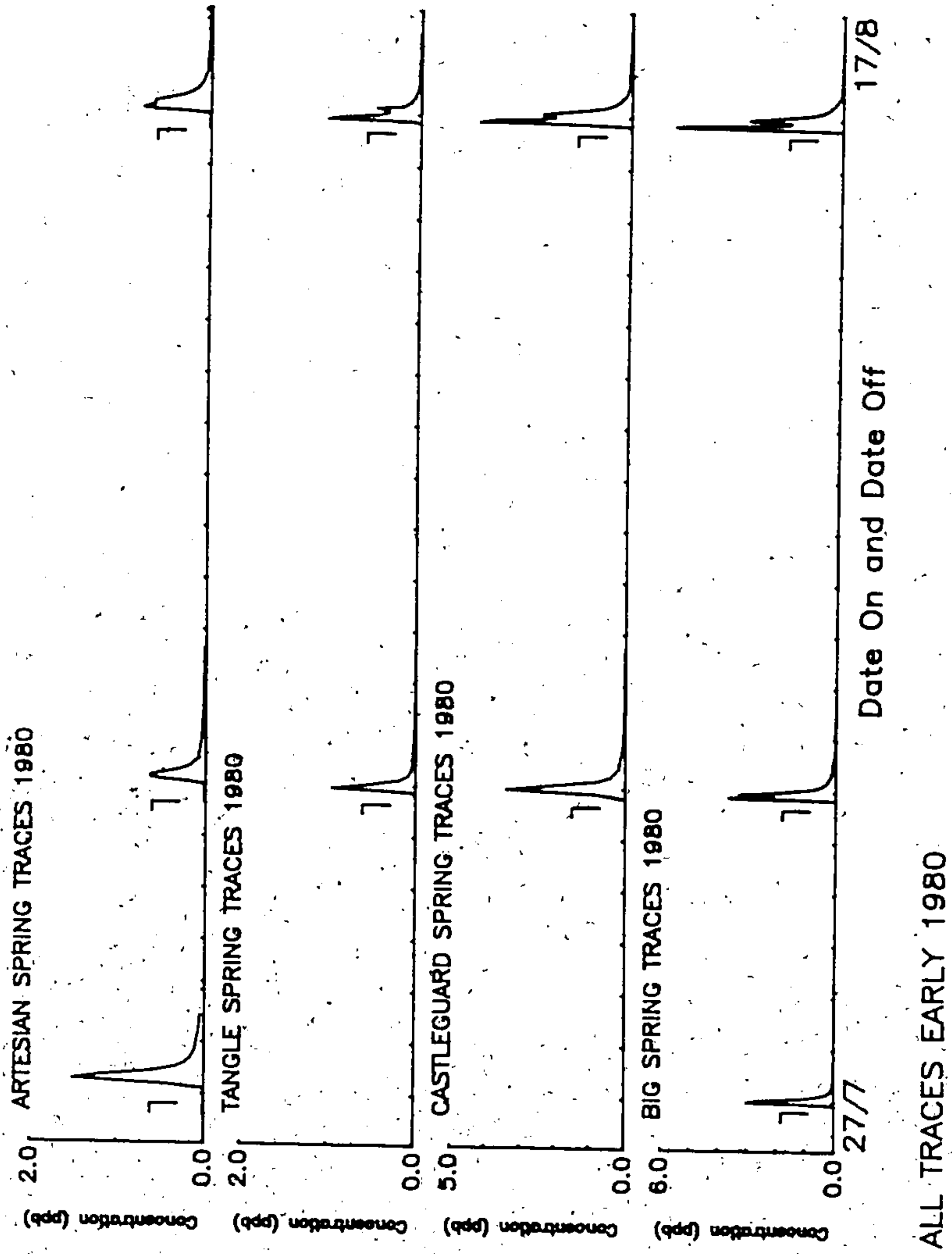


Figure 5.7a. Breakthrough curves for traces 8, 9, 10 1980

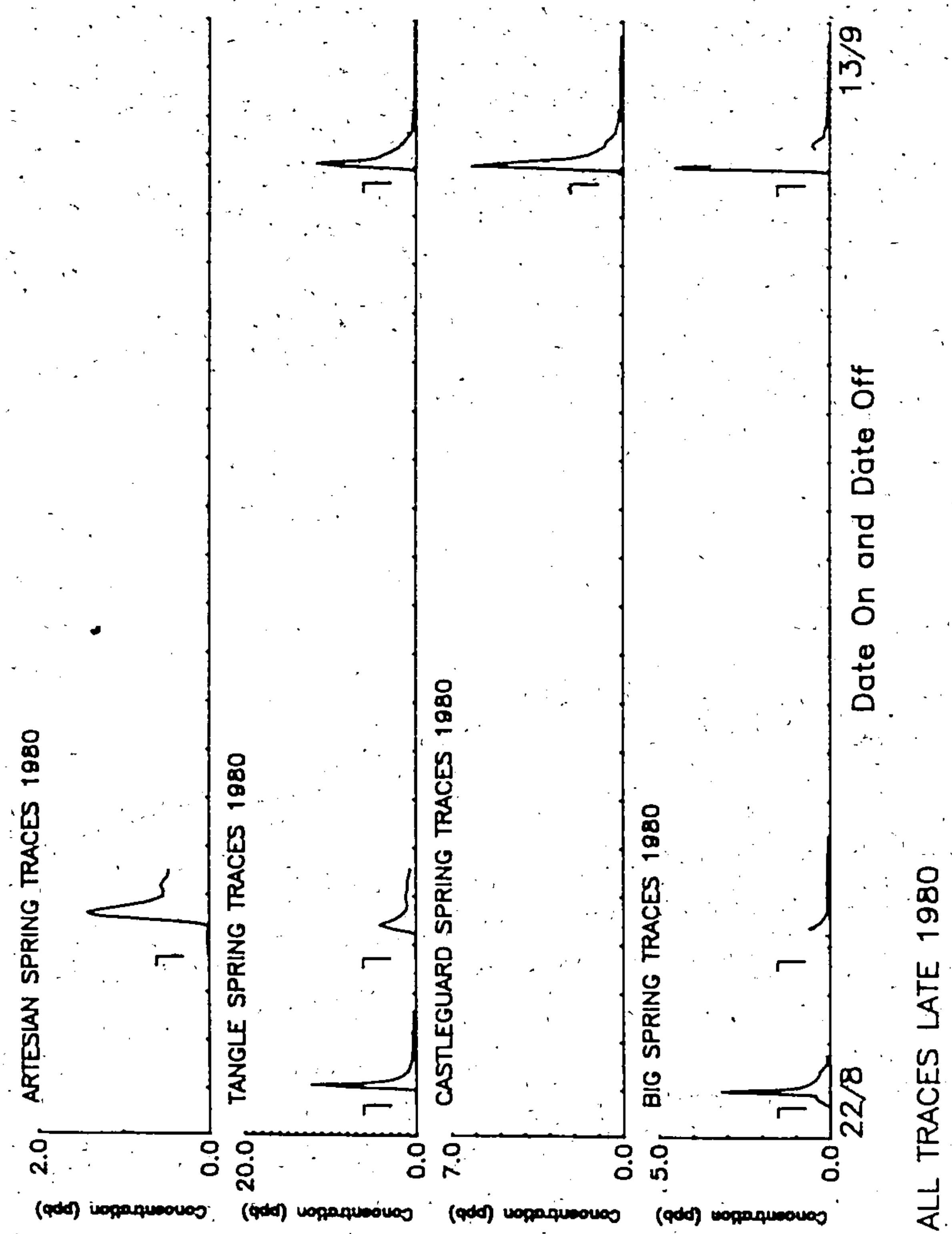


Figure 5.7b. Breakthrough curves for traces 11, 12, 13 1980

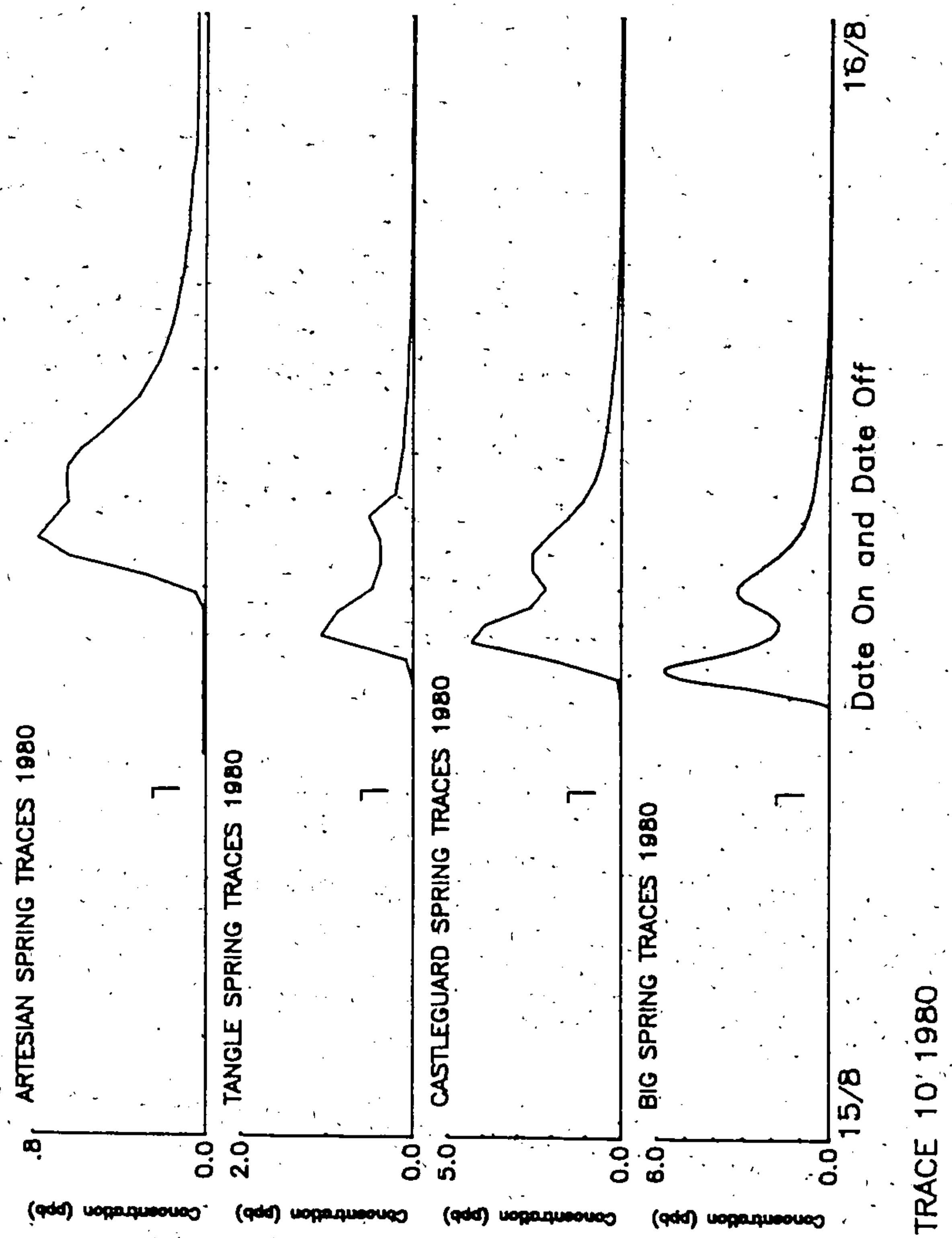


Figure 5.8. The bimodal character of trace 10 was recorded at all sites because of high frequency sampling

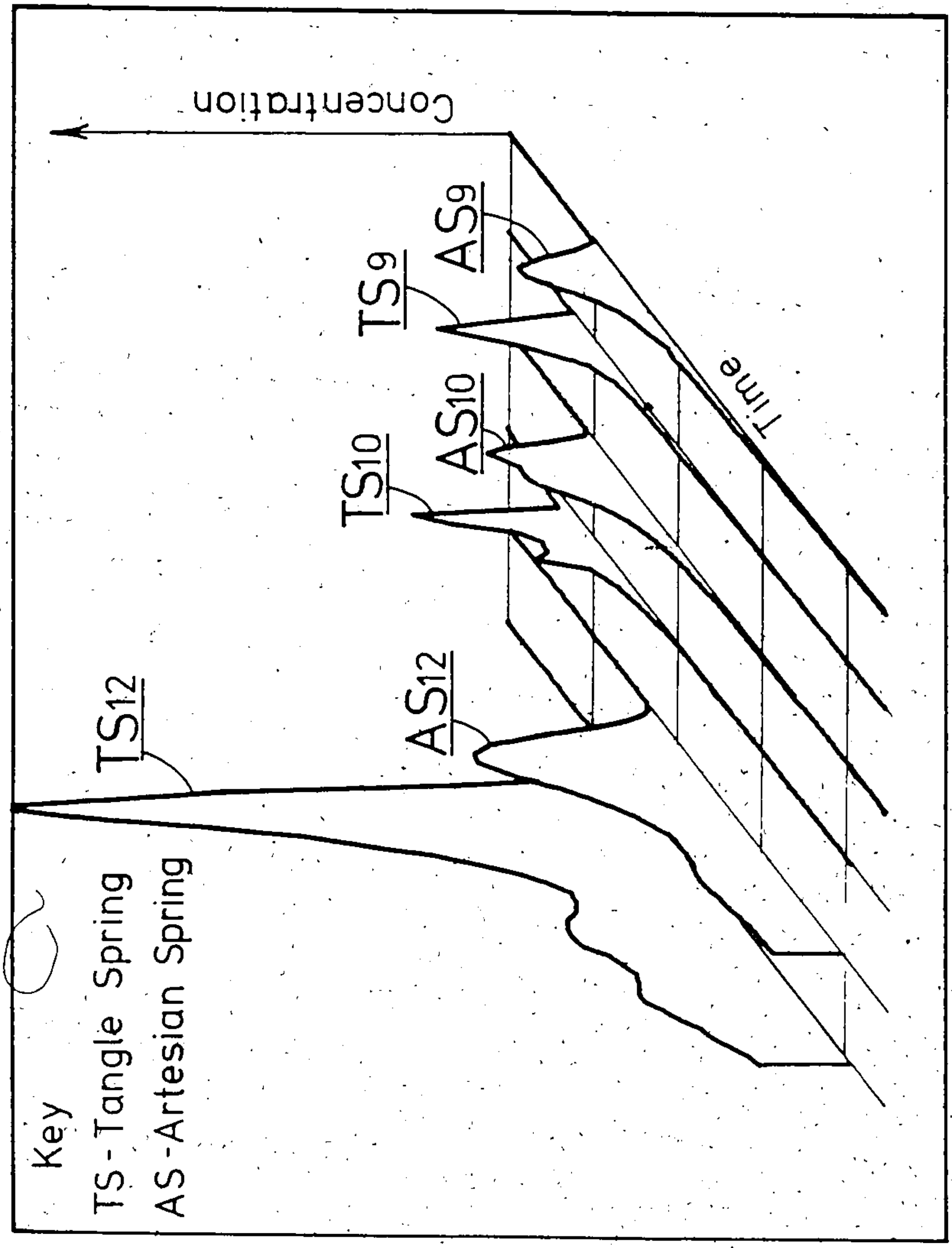


Figure 5.9. Tangle and Artesian Springs: traces 9, 10 and 12

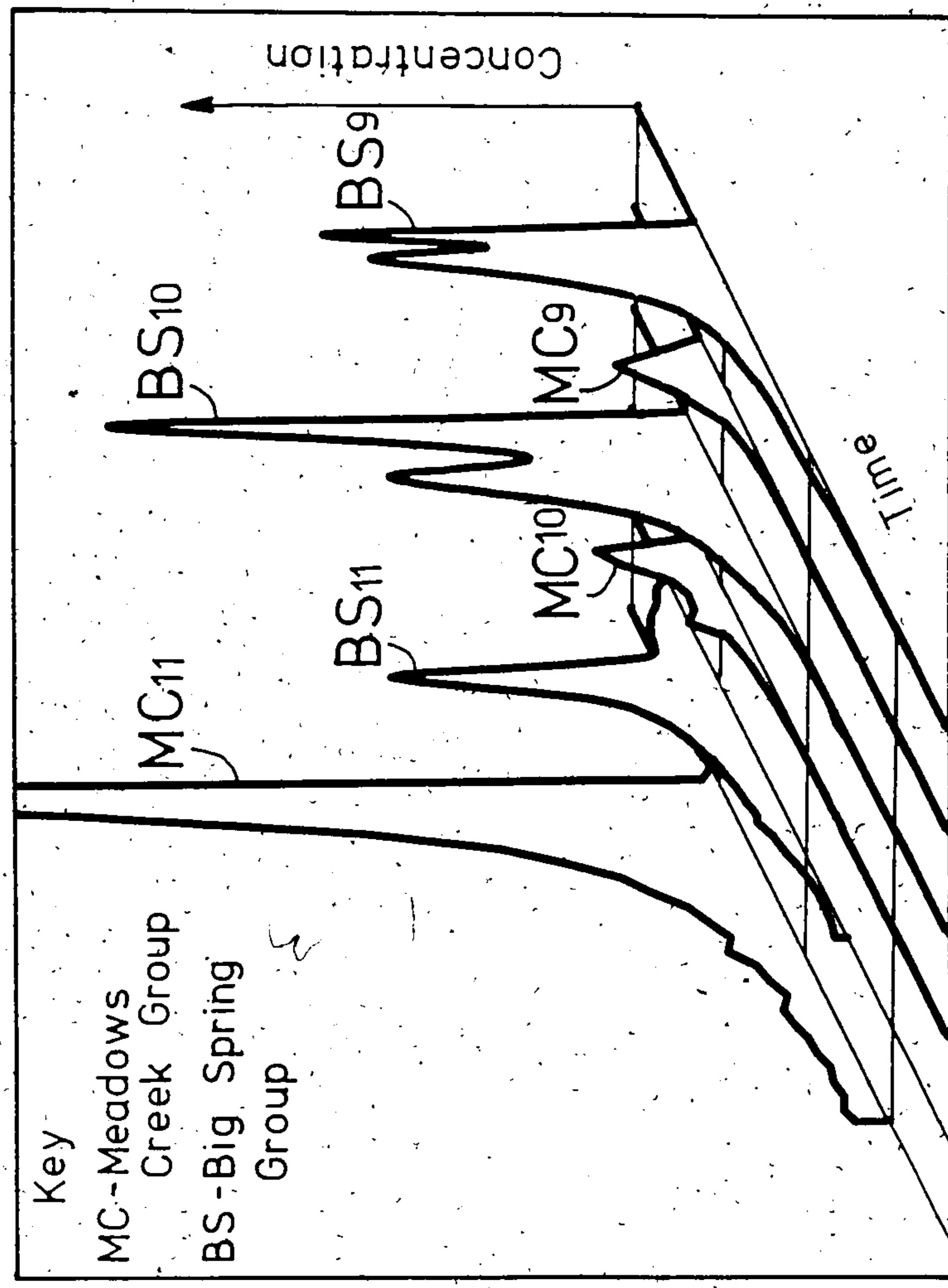


Figure 5.10. Big and Tangle Spring traces 9, 10 and 11

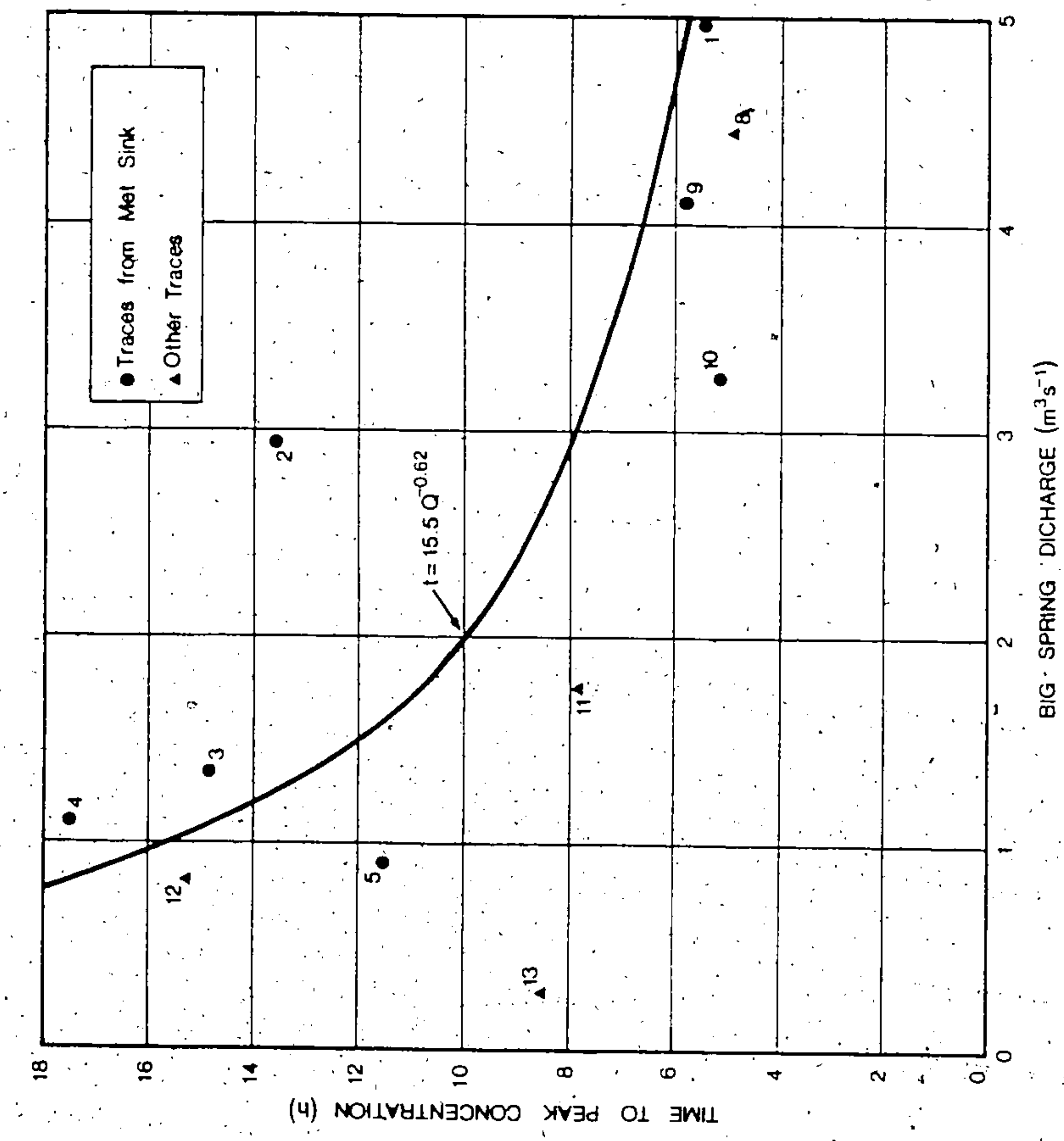


Figure 5.11. Time to peak concentration versus Big Spring Discharge

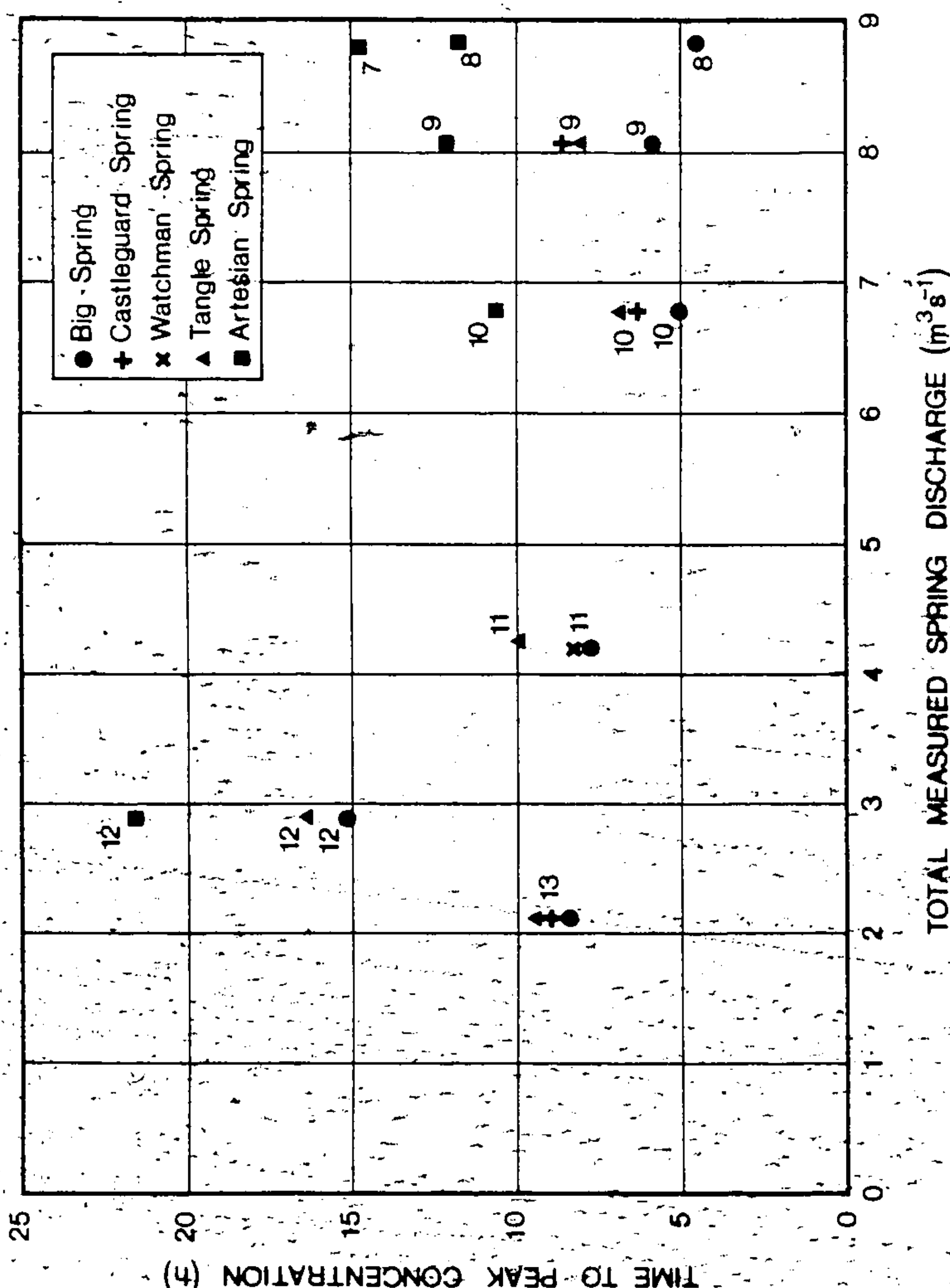
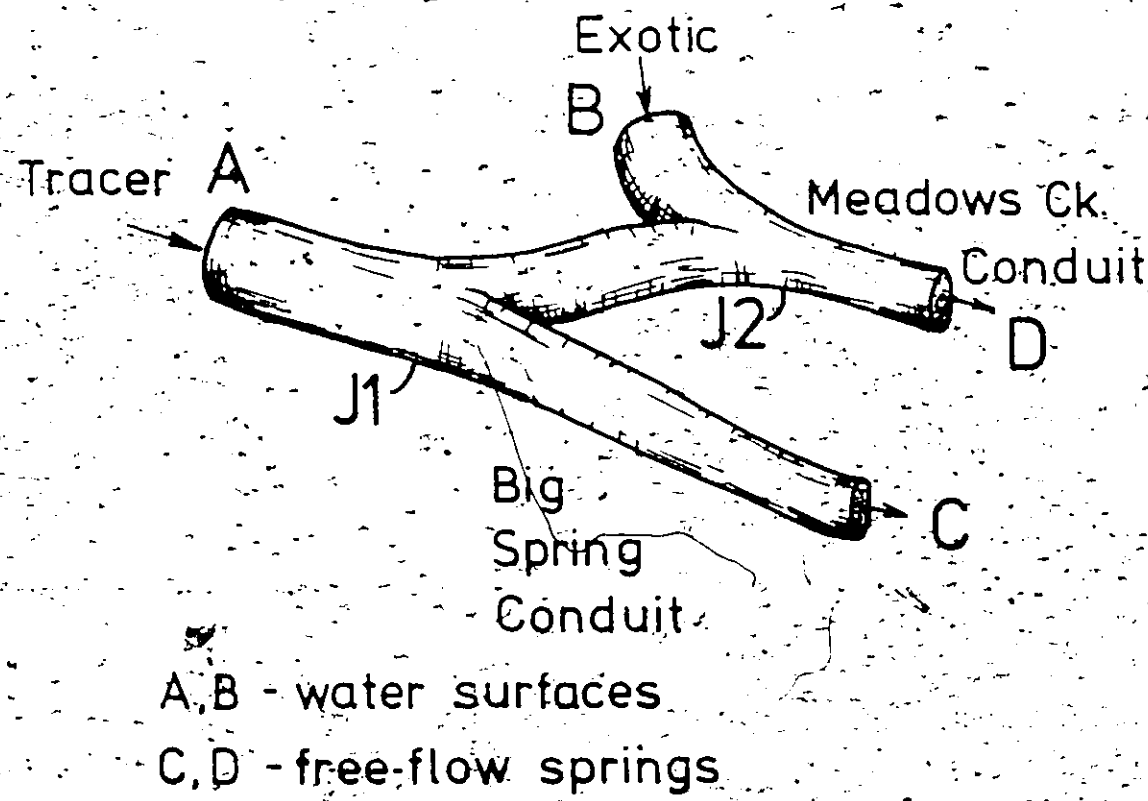


Figure 5.12. Time to peak concentration versus total measured springflow



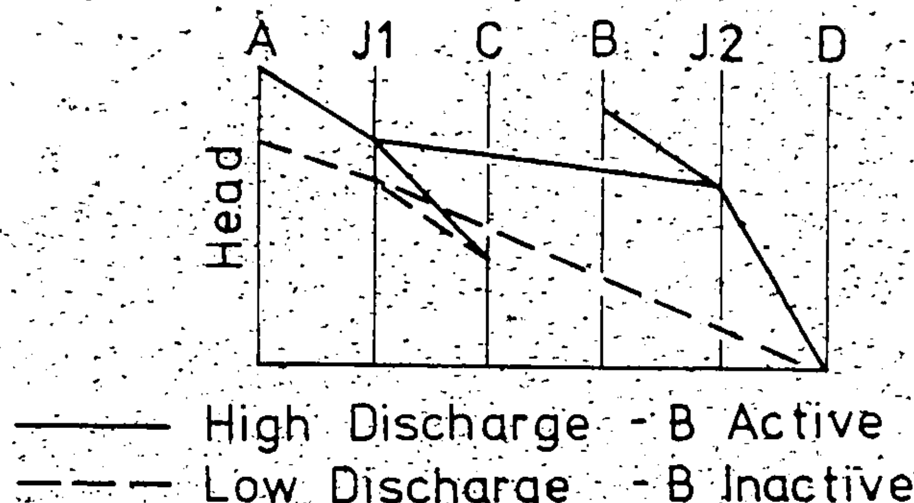


Figure 5.13. Model for the retardation of travel time to Meadows Creek Springs as discharge from an external origin competes with Castleguard II

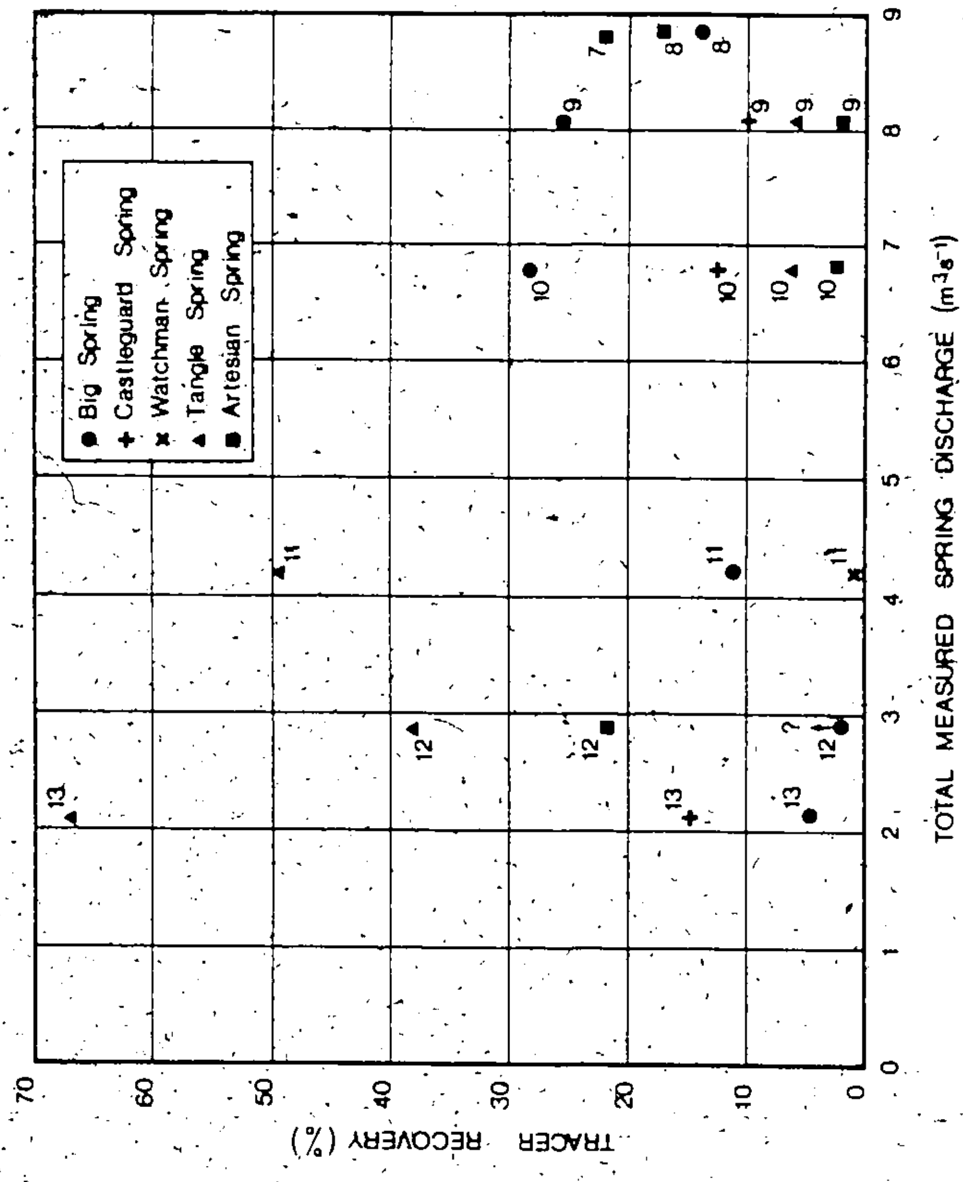


Figure 5.14. Tracer recovery versus total measured spring discharge

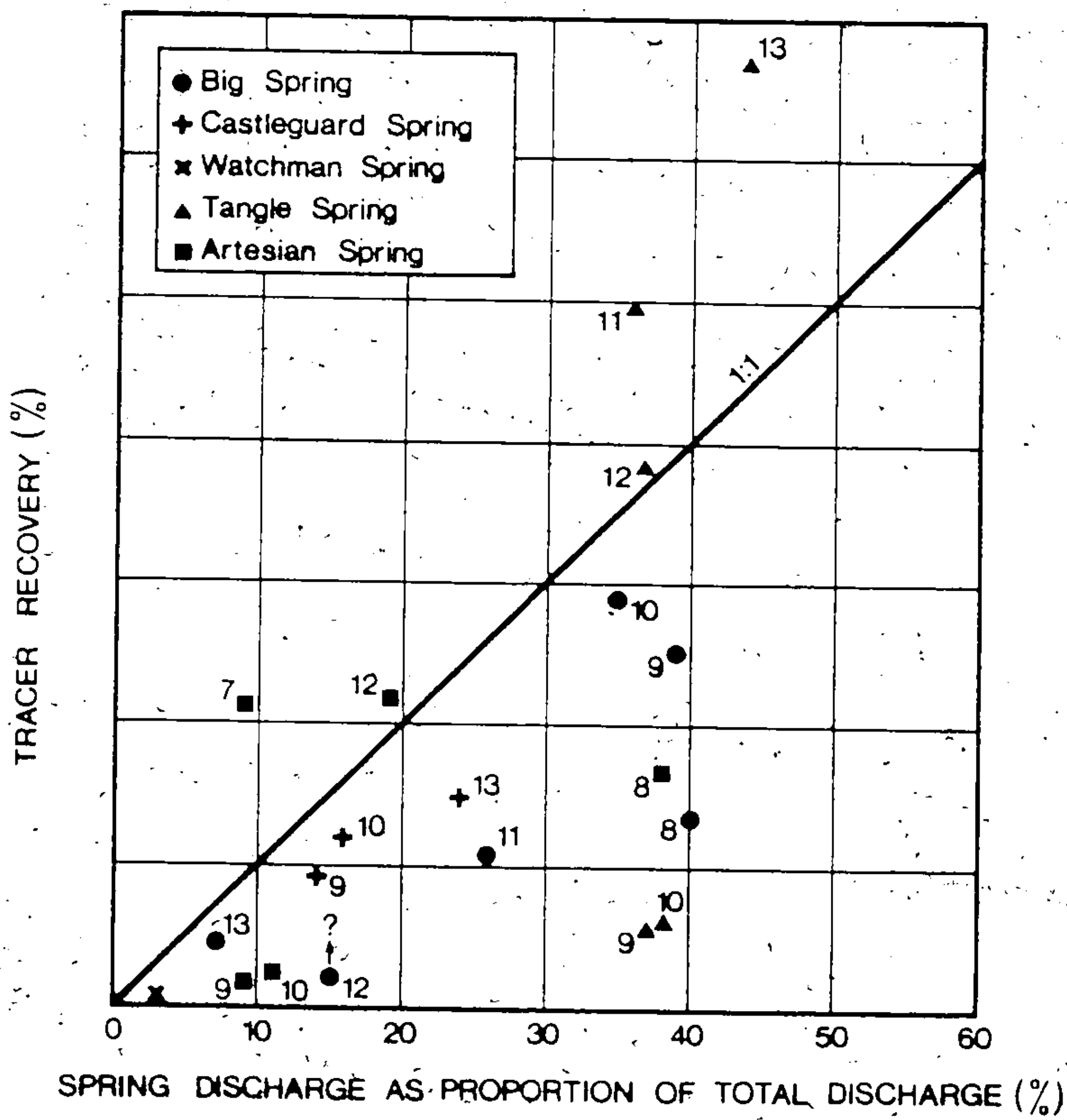


Figure 5.15. Tracer recovery versus spring discharge as a proportion of total discharge

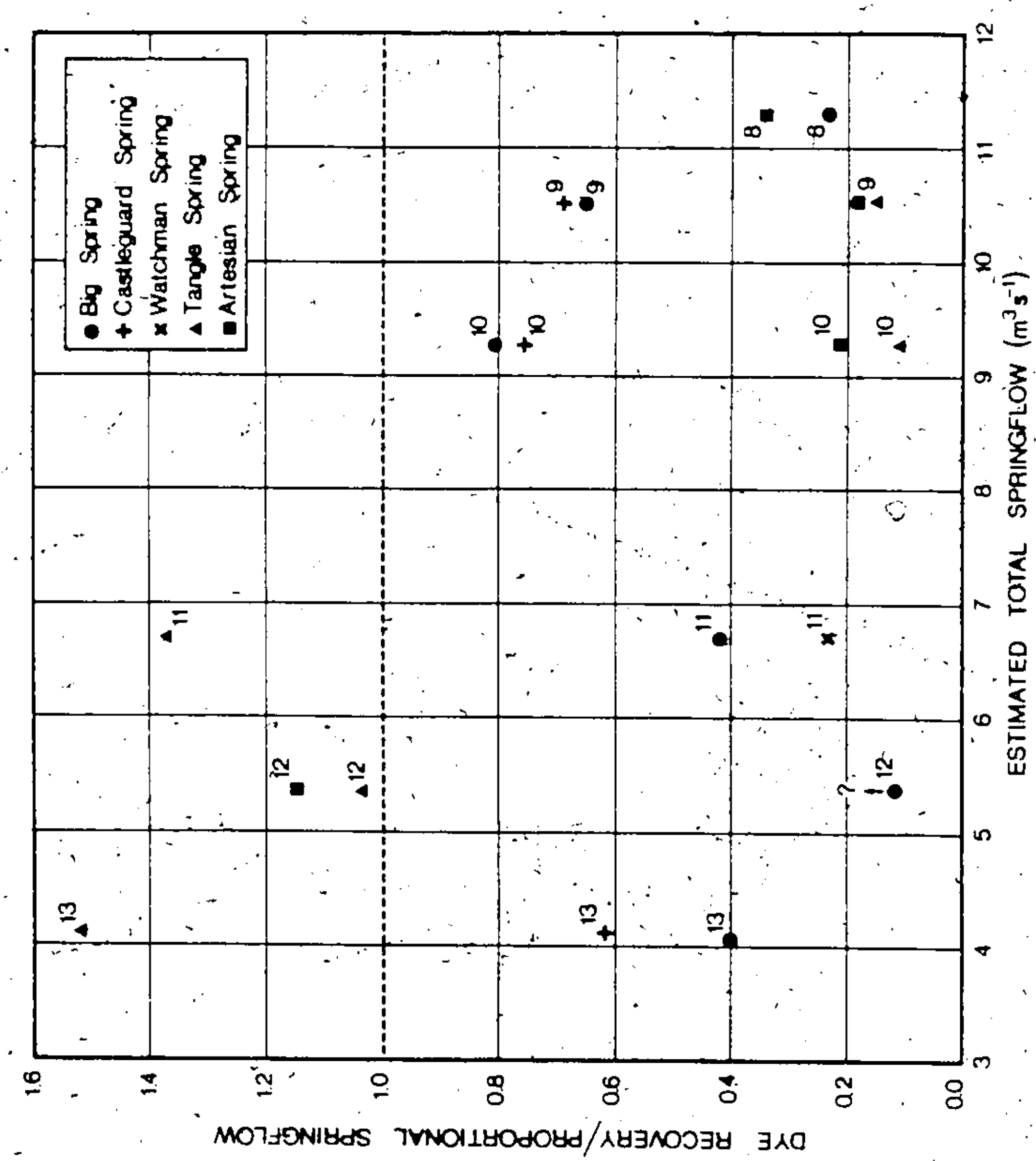


Figure 5.16. Tracer recovery over spring discharge as a proportion of total discharge versus total discharge

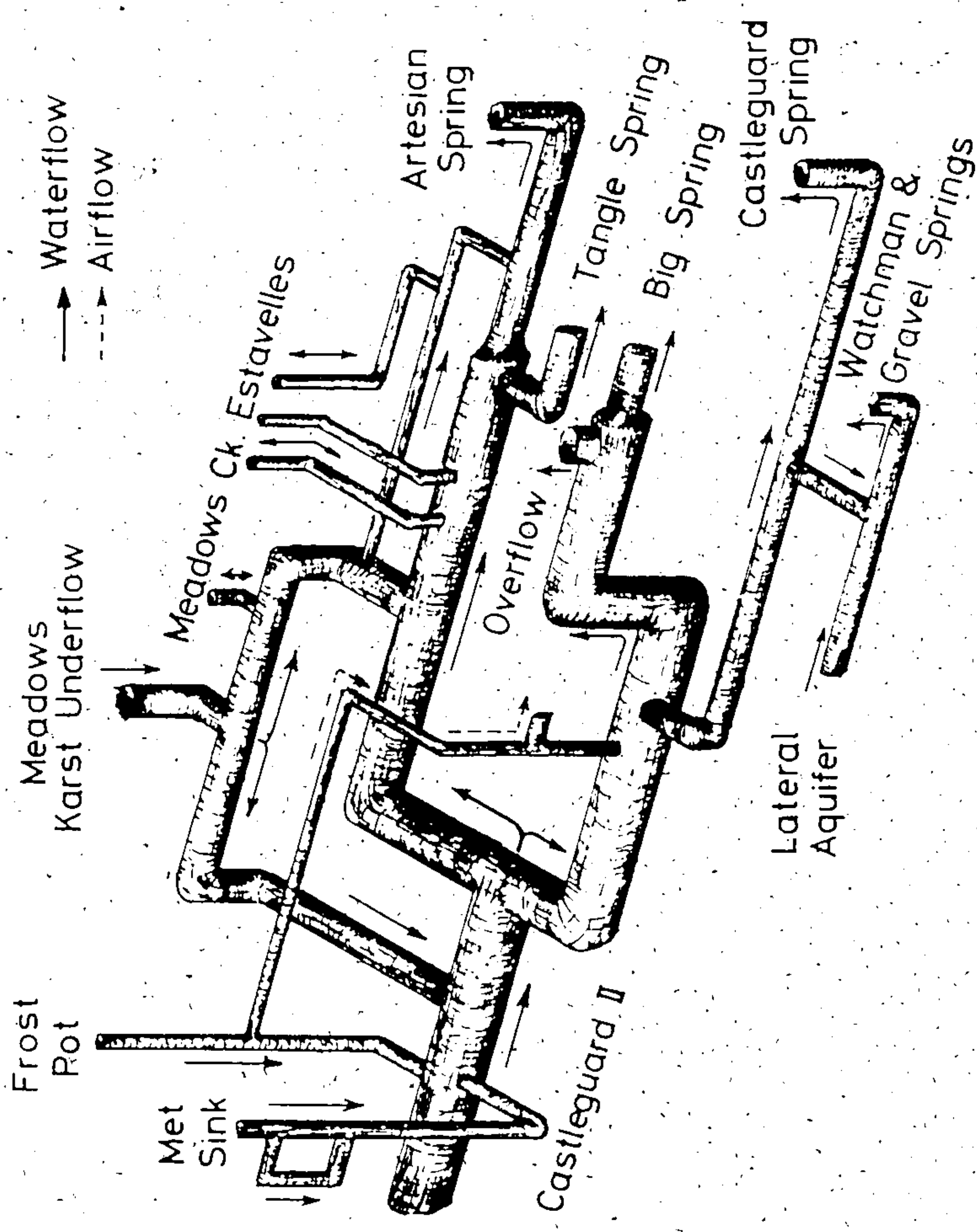


Figure 5.17. A conceptual physical model of the Castleguard Valley aquifer, based on dye tracer information

CHAPTER SIX

HYDROGRAPH SIMULATION AND ANALYSIS

The preceding two chapters have demonstrated that springs in the Castleguard Aquifer are members of a vertical hierarchy and that they are supplied by a well-developed conduit network. This chapter attempts to investigate the nature of a conduit aquifer from basic principles. Starting with simple laws of pipeflow, it considers the effect of conduit configuration on the time distribution of head in a conduit aquifer. These preliminaries lead to the development of a simple numerical model, which is used to demonstrate the peculiarities of conduit aquifers, and to simulate the general behaviour of the Valley aquifer.

6.1. FUNDAMENTAL PROCESSES

6.1.1. PIPEFLOW

A single, uniform phreatic conduit conveying water through an otherwise impermeable body of limestone is the simplest conceivable karst aquifer. Assuming no change in conduit volume with discharge, aquifer behaviour may be described by simple pipe flow equations such as the Darcy-Weisbach equation, which for tubular conduits is,

$$Q = A\,(4\,h\,r\,g\,/\,f\,L)^{0.5}$$

where Q is discharge
A is conduit cross-sectional area
h is head
r is pipe radius
g is gravitational acceleration (9.81 m/s/s)
f is a friction factor
L is length

Rearranging the equation,

$$Q = h^{0.5}\;r^{2.5}\;.\;(4\,g\,/\,f\,L)^{0.5}$$

Over geomorphic time (Schumm and Lichty 1965) radius, length and friction factor will all evolve, but discharge is by far most sensitive to differences in conduit radius. For the shorter time scales characteristic of hydrologic events, discharge is controlled by head, and conduit geometry may be taken as fixed. In this case, for a given conduit

$$Q = C\,h^{0.5}$$

where

$$C = r^{2.5} \, (4 g / f L)^{0.5}$$

This function has been graphed in Figure 6.1 for pipes of various radius, and demonstrates the importance of pipe radius. The conduit length has been set arbitrarily at 1km, and the friction factor taken as 0.33, following the empirical results of Atkinson et al.(1983). More complex and realistic pipes with constrictions, sharp bends, and other such discontinuities will have higher friction factors. However, since these parameters are not of direct concern here, an unrealistically smooth pipe will behave in the same manner as a wider, or shorter, rougher one and in this respect the parameters are interchangeable.

6.1.2. VARIABILITY OF DISCHARGE AND HEAD

The variability of discharge over time may be expressed as a probability of flow exceedance. Although the distribution of flow exceedances may be quite complex, let it be assumed that an exponential function is a suitable approximation in which

$$Pq(Q) = K \exp - K Q$$

for Q greater than or equal to zero

where

$Pq(Q)$ is the probability of flow exceeding a given flow Q.

K is the reciprocal of mean discharge

Figure 6.2 is a semi-logarithmic plot of the exponential probability density function for a hydrological system with an assumed mean discharge of one cubic metre per second. A flow of 5.9 cubic metres per second is expected to be equalled or exceeded for 1 day per year. The second abscissa gives the equivalent head developed in a karst aquifer at the upstream end of a 1 km long by 0.5 m radius conduit. The head equivalent to the one day per year flood is 950 m!

Two important conclusions may be drawn from this obervation:

1) In some karst aquifers the head development will be limited by surface topography; any excess head leading to overland flow. This provides an important process for dry valley formation, especially in constricted or less mature karst aquifers. Hanwell and Newson (1970) and Tratman (1963) describe floods from parts of the British Isles in which karst discharge capacity was exceeded and extensive surface runoff occurred. Significantly, the estimated recurrence interval for Hanwell and Newson's flood was

lower at karst springs than that estimated for the rainstorm, and considerably less than the catastrophic surface erosion appeared to suggest.

2) Alternatively, the configuration of the karst aquifer may differ from that assumed in the model. The system could have substantial storage capacity which may regulate floods, or the conduit system may be different.

In reality, surface floods are much less frequent in karst areas than Figure 6.2 might imply, which suggests that the chosen aquifer configration is untypical and that the second inference should be more closely investigated.

6.1.3. THE IMPORTANCE OF AQUIFER FORM

1) Matrix Storage

The storage capacity of karst aquifers is made up of conduit, fissure and pore spaces, in order of permeability to recharge. Recharge of fractures and pores is non-turbulent under which condition flow rate is directly proportional to head. Although the permeability of these storage media is less than that of the conduit storage, their recharge will become progressively more important as head rises compared to conduit discharge. This is an important observation, which should be reflected in the flow duration series for karst basins compared to conventional systems. E.L.White and Reich (1970) report relatively low mean annual floods from karst drainage basins, although further analysis (E.L.White 1976,1977) showed that some karst basins moderate floods and sustain flows more successfully than others. However, this interesting topic is probably not relevant to the situation at Castleguard in which matrix storage is apparently unimportant.

2) Conduit Geometry and Topology

Speleological research has demonstrated that almost all karst systems are more complex than the single conduit used in the example above. Some of the reasons for this are reviewed by Ford and Ewers (1978) and Ewers (1982). In many cases the conduit is simply larger than that in the model and is thus capable of conveying considerably greater discharges without developing excessive head. In others, the conduit is vadose, and increases in water depth can accommodate large increases in discharge according to the effects of hydraulic geometry (e.g.Leopold and Maddock 1953). Figure 6.3 shows this effect, and the sudden decrease in conveyance engendered by closing of a channel. The nature of the transition is strongly shape dependent, but in practice natural flow irregularities preclude the discharge excursion seen just prior to closure. (Dr.A.Smith pers.comm.).

In many caves compound and branching conduits exist, constituting parallel systems, or loops. These develop both as a result of internal speleogenesis, and in response to external geomorphic evolution and perturbation. Figure 6.4 shows the probability density function (p.d.f.) of head in various conduit systems assuming an exponentially distributed probability of discharge exceedance with a mean discharge of one cubic metre per second. The head exceedance probabilities are derived from the p.d.f. of discharge converted to the equivalent steady state head on the given system. Systems 1,2 and 3 are single horizontal conduits of 0.4, 0.5, and 1.0m radius respectively. System 4 consists of two 0.5m conduits vertically seperated by 20m, and system 5 is two 0.4m conduits, with 100m vertical seperation.

The impact of conduit growth is clear; well-developed karst does not develop significant head, and correspondingly will have less dynamic exchange between the conduit and the host aquifer.

For geomorphically evolved karst, a second conduit at a different level (even 100 m above the first) has a marked effect in reducing the expected head distribution to a reasonable level. P.L.Smart (1981) has presented tracer evidence for this effect, in which tracer velocities were constant above a certain discharge when an overflow conduit was active, suggesting a steady head despite varying discharge.

The spring discharge time series will show the effect of sudden changes of internal configuration. At flow initiation, a conduit is extremely sensitive to head variations because:

$$dQ/dh = C / 2h^{0.5}$$

This is graphed for a 0.5m pipe in Figure 6.1. While the derivative of discharge with respect to head is greater than one, discharge will change at a faster rate than head up to the point where

$$h = C^2 / 4$$

This means that inflow variance will be predominantly visible in the overflow discharge, especially if the overflow conduit (and thus "C") is large.

These principles are illustrated in Figure 6.5, which is a simple example of a "static" model of an overflow-underflow conduit hierarchy in which the storage term has been ignored, and the time axis therefore remains unlabelled. In this case, the estimates are instantaneous head solutions to an arbitrary input discharge, and are based on solution of the Manning equation for various conduits with a Manning's "n" = 0.03, and horizontal distance = 1 km. The greater complexity under open-closed channel conditions is shown for a simpler configuration in Figure 6.6. In both cases the high variability of the active overflow in comparison to the underflow is evident.

Although these models are somewhat crude, they explain the anomalous behaviour of the Castleguard Springs noted in Chapter 4. In particular, they illustrate the ability of a spring to behave as both an overflow and underflow, depending upon the state of the system.

Thrailkill (1968, 1974) suggested that the discharge pattern in a conduit aquifer was not affected by flow regime (laminar or turbulent), nor was flow rate significantly different in conduits at different levels in the aquifer, providing they had common inflow and outflow points. However, as Palmer (1981) showed, conduit aquifer behaviour is strongly dependent on flow regime. Furthermore, as the present work demonstrates, in aquifers experiencing significant water level changes the relative elevation and geometry of the conduits is of great importance in dictating their flood response.

6.1.4. CONDUIT STORAGE

The examples above show the general form of response in a karst aquifer, but ignore any moderation of the flood waves caused by changes in storage. Dynamic conduit storage occurs as water levels rise above the "normal" phreatic level, into a zone called the epiphreas. Ashton (1966) states that "an increase in the head of water in this section [the epiphreas] will cause an almost instantaneous increase of flow throughout the phreas, and so flood pulses will be passed from this point through the phreas practically without delay." While this is true with respect to initial flood response, it only applies to the rest of the hydrograph in a system without storage, or in a frictionless conduit (approximated by a very large, short, unobstructed channel). In general, a finite increase in head is necessary in order to increase conduit discharge. This requires a certain volume of water, which will be abstracted from the leading edge of the flood pulse, and released on falling stage. The responsiveness of a system depends on the area of the water surface at the input (A), the resistance of the conduit to flow (1/C), and the head (h). Palmer (1981) has illustrated this effect for a single discrete conduit (Figure 6.7).

Conventional flood-pulse models of karst aquifers (Figure 6.8, a) are now seen to be inadequate, because of the delayed transmission of a flood pulse through both the phreatic and vadose zones. The phreatic delay is incurred by storage in the epiphreas, which results in a small, but significant change in the aquifer model (Figure 6.8, b). A component of this "dynamic" storage will be in abandoned conduits which may ultimately be reactivated. At this point the entire configuration of the aquifer will change and discontinuities may be observed in spring behaviour.

The remainder of this chapter is concerned with illustrating these points using both a simulation model and data from Castleguard, perhaps the most extreme conduit aquifer ever studied in detail.

6.2 SIMULATION MODEL OF A CONDUIT AQUIFER

The karst aquifers defined in Chapter Three are at present too complex to be realistically modelled. However, an initial understanding of how a conduit aquifer behaves can be developed and may be subsequently expanded to encompass more realistic situations. The present model consists of a pure conduit aquifer in which phreatic conduits rise from a vertical shaft system. The input hydrograph is applied directly to the shaft phreas so that no vadose component is involved.

6.2.1. MODEL DESIGN

1) System Configuration

The active part of the model is designed to be entirely phreatic, and pipes must be entirely filled before they discharge water from their outlet. The pipes themselves are unbranched, but rise from a mutual, vertical phreatic shaft at their upstream end. The horizontal cross-sectional area of the shaft is specified and each pipe has a defined length, radius, roughness and inflow and outflow elevation. The outflow elevation is always greater than or equal to the inflow crown elevation. In this way, the pipe must be filled before it starts to flow at its outlet, and acts as a part of the dynamic storage of the system (see the inset diagram in Figure 6.5). Compound pipes are not considered.

2) Input Hydrograph Generation

The input to the system is assumed to be applied directly to the phreas, and routing effects must be explicitly included in the generating function, although a phase shift operator can be used to delay or advance the hydrograph.

Inflow was modelled as two superimposed (additive) "diurnal" flood events, each lasting 48 hours and mutually lagged by 24 hours. In this way, the recession component of a given day continues for the next 24 hours, after which a new (or recursive) parameter set is used. The input hydrograph is defined as:

Q(t) = P0 + P1(t-tmin)**P2 exp - P3(t-tmin)

summed for the two overlapping hydrographs

where Q(t) = input discharge at time t (h)
P0 = parameter provided by user
P1 = parameter provided by user
P2 = parameter provided by user (usually 1.4)
P3 = parameter provided by user (usually 0.25)
t = time of day (plus 24 h for earlier flood)
tmin = 24.0 - time of daily minimum (phase-shift operator)

In practise changes only in P0 and P1 were necessary to provide a suitable diurnal inflow. Transient flood events were simulated by setting P1, P2 and P3 at 100, 4.0 and 4.0 respectively.

3) System Response

The initial water level (h) in the aquifer is specifed with the initial time (t) and time increment (dt).

The first calculation is the estimation of inflow discharge (Qin) at time t. Then a scan of conduit "status" is made. Pipes with outlet elevations below the given water level are active, and their instantaneous discharge at the initial head calculated using the Darcy-Weisbach equation. Pipe discharge is summed to give outflow discharge (Qout). Pipes with inlets below water level, but not flowing are part of the dynamic storage, while those with inlets above present water level are disregarded.

The net discharge is calculated as Qin - Qout, and this will cause a rise or fall in water level. The change in water level (dh) is defined by

$$dh = ((Qin - Qout)\ dt)/A$$

where A is the aquifer cross-sectional area in a horizontal plane at h, and

dt is the specified time increment.

Area is an important variable; if large, the resulting head change is relatively small, and vice versa. Area is calculated as the shaft cross-sectional area plus the cross sectional area (in a horizontal plane) of any non-discharging pipes with an inflow elevation below present water level. There are thus discontinuities in the cross-sectional area of the aquifer as certain pipes become active (or inactive) as storage media or discharge conduits. The head response to a given net discharge is thus dependent on both aquifer configuration and water level.

The most marked discontinuity is the increase in outflow and decrease in aquifer area caused when a pipe first starts to flow. The crude extrapolation over the given time increment is then unrealistc. This problem is avoided by ruling that a predicted level cannot directly change the status of the aquifer (ie. no pipe should start or stop flowing in the given time interval). In such cases, the time interval is halved until two conditions are met: first, the predicted level must be in the same domain as the initial level; and second, the level change must be less than half of that between the initial level and the level at which configuration changes. This "procrastination" routine operates until the predicted level is less than some critical value from the critical pipe overflow (1mm was used in this case), and then aquifer level is set at that critical level.

If the pipe on the point of becoming active is relatively unresistant to flow and the change in water surface area very great, then numerical instabilities still result which can only be avoided by taking very small time steps, or else by introducing an optimisation routine.

Both are expensive options; in one example a time step of 0.36 seconds was necessary in order to avoid instability. Alternatively, the aquifer can be designed so that dramatic changes in form do not occur.

4) Model Reliability

The major imprecision in the present method is the linear extrapolation of instantaneous flow rates over the succeeding time interval. While inflow discharge can be easily integrated over the given time interval, the outflow is more dificult to predict. In general, outflow rates are overestimated slightly, depending upon the time increment and system configuration.

No minor head losses are considered, they are implicitly included in the friction factor, which (somewhat unrealistically) is independant of pipe discharge. There is no open-channel flow, and the sharp transition from open channel to conduit flow (Figure 6.3) is thus avoided.

Conduits which are active as storage media are considered "frictionless" ie. there is an instantaneous matching of the head in the shaft and the pipe system. Estimation of more realistic behaviour is feasible, but rather arbitrary assumptions are necessary concerning conduit configuration and behaviour.

In general, the model is more abrupt than a real aquifer would be, having rapid transitions of behaviour which friction and storage would ameliorate in reality. Conversely, the steepening effect in a flood pulse passing down a dry channel is not considered, nor are effects like "hydraulic damming" (e.g. Mohring et al. 1983), although these may be important mechanisms in generating the Cave floods (Section 4.3.3).

6.2.2. SIMULATION RESULTS AND INTERPRETATION

1) Epiphreatic Delay (Runs 1 and 2)

Figure 6.9 shows the delay in flood transmission caused by varying the area of the epiphreatic zone in a single pipe system (Table 6.1). In the first example the losses into storage are trivial and there is no perceptible delay in flood peak, and negligible distortion in the transmission of the input flood. In the second example, using the same conduit system, but with a shaft area of 1000 square metres, a significant proportion of the input flow is temporarily drawn off into conduit storage, resulting in a delay (1.75 h) and deformation of the flood pulse in passing through the aquifer.

Delay in time to peak is not necessarily a vadose phenomenon, delay in initiation of the flood is a better (but less easily determined) parameter. However, the simulation of transient events below (Section 5.2.5) shows that even time of initiation may be an ambiguous parameter.

2) Multiple Conduit Systems (Runs 3 and 4)

An aquifer with two or three vertically separated conduits will show sharp hydrologic discontinuity at the initiation or cessation of flow in the superior active conduit. Such an aquifer was used in runs 3 and 4 (Tables 6.3 and 6.4) and the results are illustrated in Figures 6.10 and 6.11. The input from individual conduits shows the characteristics of underflow and overflow springs as presented in Chapter Four; strong peaks in the overflow, and truncated hydrographs at the underflow.

Storage in conduits is responsible in part for the assymetry of the flood pulse; delaying activation of a spring on rising stage, and sustaining flows during falling stage. This accounts for the convex form of the underflow output curve (e.g.pipe 1, Figure 6.10) even when the recharge hydrograph is concave in shape.

Where conduits rejoin within the aquifer their discharge may be combined, thus generating more complex hydrographs (e.g.Figure 6.12 from Run 4). If the outflow is not free, feedback may occur resulting in non-additive convergence of the contributaries (such as that identified from dye tracing results, Figure 5.13, Section 5.3.4). No such mechanisms are considered here.

3) Recession of a Conduit Aquifer (Runs 5 and 6)

If recharge of a conduit aquifer ceases at some arbitrary point, then the drainage of the water stored in the conduits will have a characteristc recession. Palmer (1981) has shown that drainage by turbulent flow (as implied in the use of the Darcy-Weisbach equation) does not produce an exponential recession. Figure 6.13 shows how a three conduit system (Table 5) will drain; the overflow ceasing relatively suddenly and the underflows showing marked changes in behaviour as the aquifer configuration evolves. When conduits are combined (Figure 6.14), yet greater complexity emerges.

These "recession curves" bear a striking resemblance to compound recessions reported in the literature (e.g.Mijatoric 1968, Milanovic 1975, 1981, Torbarov 1975) which are commonly interpreted as reflecting the progressive drainage of distinct storage media. Apart from questioning how storage in pores, fissures and conduits can be "independent" in a karst aquifer, it is shown here that drainage of a pure conduit aquifer can produce apparently compound recessions. Their form reflects the active conduit topology and geometry as much was storage volumes per se.

Recognising the compound storage in most karst aquifers, a simulation was run on a conduit system of 9.6 thousand cubic metres initial storage (Same form as Run 5, Table 6.5) which was also receiving an independant contribution from a linear reservoir (e.g.a porous medium) with an initial volume of 0.29 million cubic metres, and a recession coefficient of 0.025 per hour (Run 6). Initially, all three conduits were active and shared approximately exponential recessions (Figure 6.15), but all at different rates implying (according to conventional wisdom) quite distinct characteristic aquifers feeding each conduit.

Although these simulations are very simplistic, it is clear that in a conduit aquifer the recession curve is not necessarily a straightforward phenomenon. It is unfortunate that the method of recession analysis is so firmly established in the karst water resource evaluation literature (e.g.Atkinson 1977, Burdon and Papakis 1963, Burger and Dubertret 1975, Karanjac and Altug 1980).

This difficulty has already been recognised from work on chalk and oolite aquifers, which have a multi-component flow system (e.g.Atkinson and Smith 1974, Downing et al.1977, Ineson 1962). These systems create difficulties in interpreting recession and pumping test data (e.g.Foster 1974, Foster and Crease 1975, Foster and Milton 1974). These have led to the identification of a rapid-flow mechanism (e.g.Fox and Rushton 1975, Rushton and Rathod 1979, E.J.Smith 1979), that is only active under certain recharge or water level conditions. This is partly because the hydrogeological properties of these aquifers vary markedly with depth (e.g.Headworth et al.1982), which also means that special analytical techniques are necessary for their study (Rushton and Chan 1976, Rushton and Rathod 1980). One of the most successful models of a partially karstic aquifer is one recently developed for the Lincolnshire Limestone, U.K. (Rushton et al.1982). However, a satisfactory performance was only obtained after employing field values for aquifer parameters at multiple depths at each finite element node, emphasising the difficulty of developing a general model for karst aquifers.

The extension of the techniques developed for the chalk and Lincolnshire Limestone aquifers to more predominantly karstic situations has not yet taken place. The present simulation results emphasise the necessity of such a transfer.

4) Transmission of Transient Events (Run 7)

Rainstorms at Castleguard often produce a brief, but intense transient flood which can be simulated by superimposition of a brief "input pulse" (Qt) on some arbitrary baseflow, where

$$Qt = 100t^{4}\exp(-4t)$$

and t is time (in hours) from flood initiation.

Replicate transient events were superimposed upon a more general exponential recession to test their impact during various aquifer states (Figure 5.16, Table 5.6, Run Number 7). The results of this simulation are probably very misleading, because the aquifer model as described above includes no dynamic terms. These are of paramount importance in the simulation of the translation of sudden discontinuities of discharge through a conduit aquifer.

At high flow, the transient is transmitted rapidly and with little alteration except for a general (but uneven) attenuation. At lower discharges an increasing portion of the flood is retained in storage, leading to greater distortion (Figure 5.16b). There is also greater delay at lower discharges, in particular in the overflow hydrographs. The aggregate output discharge (Figure 5.16b) becomes rather complex as individual pipes produce quite distinct output hydrographs.

5) Multi-day Simulations (Run 8)

Running the simulation model over several days becomes somewhat expensive, especially if very small time steps are necessary in order to avoid numerical instability. Table 6.6 gives the aquifer configuration for Run 8. To stabilise the model, a fourth, large, high outlet conduit has again been included, although it is never active as an overflow, but is a storage element supplementing shaft storage. Figure 6.17 shows the buildup of storage in the aquifer culminating in the characteristic overflow-underflow behaviour described in Chapter 4.

The form of the output hydrographs can be modified indefinitely by adjustment of the inflow and aquifer parameters. Here, however, it is perhaps better to recognise the approximate nature of the model and not to attempt to create an "absolute" Castleguard Aquifer.

6.2.3. CONCLUSIONS FROM THE SIMULATION MODELLING

The simulation experiments have shown that behaviour of a conduit aquifer is strongly conditioned by the pipe geometry and topology. Recession analysis and pulse train analysis have been shown to be potentially misleading investigative techniques for conduit-dominated karst aquifers. Furthermore, the more sophisticated stochastic methods (e.g.Dreiss 1982, Knisel 1972) are probably insufficiently robust to cope with severe system discontinuities and nonlinearities should they occur.

The general results of the simulations are fairly reliable, but there remain several gross simplifications and assumptions in the model. Although these may be overcome by increasing sophistication, more precise system definition is required, which for most non-trivial cases will be increasingly arbitrary, or else expensive to obtain.

One remaining question is similar to that originally posed by White and Schmidt in 1966 (p559): "What proportion of the total groundwater in the aquifer is moving by localised [conduit] flow and what proportion by diffuse flow?" How representative is the simple conduit model of a real karst aquifer? Very few karst aquifers have totally conduit flow, and only high quality field data can reveal the relative importance of conduits in a given situation. In general, these data are not available, but many of the peculiarities of the Castleguard results appear to stem from a dominantly conduit groundwater system. Inspection of the hydrographs in this light provides a little additional insight to the aquifer. In future, far more detailed interpretations may be possible.

6.3. CLOSING REMARKS

6.3.1. INTERPRETATION OF SELECTED CASTLEGUARD HYDROGRAPHS

The overflow and underflow regimes illustrated in Figures 4.11 and 4.12 can be effectively explained using the simulation model.

Some additional detail concerning the Big Spring can be gained from Figure 6.18. In the upper figure, taken from a period when the Big Spring itself was only intermittently active, small underflow springs maintain flow with characteristic diurnal variability. Initiation of the Big Spring causes a dramatic rise in total discharge, while cessation of flow is more gradual. The hydrograph (Fig. 6.18a) is characteristic of the convergence of an overflow and an underflow pipe, in which conduit storage increases on rising stage, and is depleted on falling stage. A similar superimposition of a hydrograph peak is seen at higher discharges (Fig. 6.18b) in the hydrograph from the Big Spring itself, suggesting an internal overflow route. Given the known overflow into Camp Stream, a local conduit network is proposed in Figure 6.19.

The stage record from the Forest Spring (located some 15m below the cave entrance; Fig. 3.3) also shows compound overflow-underflow behaviour (Fig. 6.20). This was initially attributed to the shape of the control section at the stage recorder, but the steps are also seen on the Cave Stream stage record obtained further downstream. Unless there is a fortuitous similarity between the two control sections, then the three levels of discharge may represent three independant conduits which amalgamate upstream of the recorder to produce a compound hydrograph (cf. Fig. 6.12). The Forest Spring indeed consists of two separate outlets, and a conduit network like that in Figure 6.20 may be hypothesised, with the cave as the ultimate controlling overflow.

6.3.2. CONCLUSIONS

The variability in behaviour of the Castleguard Springs has been reasonably well accounted for by the uniquely conduit nature of the karst aquifer. A very simple simulation model has been developed which acceptably mimics the field data. The model's success also undermines the application of flood pulse and recession methods in conduit karst aquifers. The significance of the errors in these techniques depends on the nature of the aquifer in question, but in so far as all karst aquifers exhibit conduit flow, they may no longer be ignored.

Future experiments with the model will include more satisfactory optimisation routines, more realistic exchanges into and out of storage, the addition of porous and fracture storage media, and possibly the addition of a vadose routing algorithm. The difficulty with such developments is in determining realistic form parameters for the inaccessible parts of the aquifer. Arbitrary design is quite unreasonable considering the sensitivity of the model to aquifer form. Rushton et al.(1982) have recently emphasised the need for abundant empirical information for both the conceptual design of, and parameter assignment to, successful models of carbonate aquifers.

The scarcity of reports of underflow-overflow behaviour in the literature is partly because most karst aquifers are less dynamic and subject to less frequent flooding than is Castleguard; consequently, their behaviour is less conspicuous. However, verbal reports of observed overflow-underflow phenomena are relatively common, suggesting that it is the (economically preferable) collective gauging of karst springs which is concealing real aquifer behaviour. It is recommended that karst resurgences should be gauged individually, at least initially, so that aquifer behaviour can be properly identified. Furthermore, the records of conduit springs are themselves amenable to analysis, such that, in future, quite detailed descriptions of the interior of the aquifer may be obtained from suitable hydrologic data.

RUN NUMBER 1

Number of Pipes = 1
Shaft Cross-sectional Area = 10.0 square metres
Initial Aquifer Level = 0.8521 metres

PARAMETER		PIPE NUMBER 1	2	3	4
LENGTH	=	1000.0			
RADIUS	=	1.0			
VOLUME	=	3141.5			
ROUGHNESS	=	0.33			
INFLOW INVERT	=	-2.0			
OUTFLOW CROWN	=	0.0			

Table 6.1. Simulation Run 1: Small Epiphreas, Aquifer Configuration.

RUN NUMBER 2

Number of Pipes = 1
Shaft Cross-sectional Area = 1000.0 square metres
Initial Aquifer Level = 0.8521 metres

PARAMETER		PIPE NUMBER 1	2	3	4
LENGTH	=	1000.0			
RADIUS	=	1.0			
VOLUME	=	3141.5			
ROUGHNESS	=	0.33			
INFLOW INVERT	=	-2.0			
OUTFLOW CROWN	=	0.0			

Table 6.2. Simulation Run 2: Large Epiphreas, Aquifer Configuration.

RUN NUMBER 3

Number of Pipes = 2
Shaft Cross-sectional Area = 10.0 square metres
Initial Aquifer Level = 8.0 metres

PARAMETER		PIPE NUMBER 1	2
LENGTH	=	1000.0	1000.0
RADIUS	=	0.5	2.0
VOLUME	=	785.4	12566.
ROUGHNESS	=	0.33	0.33
INFLOW INVERT	=	-1.0	-1.0
OUTFLOW CROWN	=	0.0	5.0

Table 6.3. Simulation Run 3: Two Conduit Aquifer Configuration.

RUN NUMBER 4

Number of Pipes = 4
Shaft Cross-sectional Area = 100.0 square metres
Initial Aquifer Level = 0.0 metres

PARAMETER		PIPE NUMBER 1	2	3
LENGTH	=	3000.0	3000.0	1000.0
RADIUS	=	0.5	1.0	2.0
VOLUME	=	2356.1	9424.7	12566.
ROUGHNESS	=	0.33	0.33	0.33
INFLOW INVERT	=	-1.0	1.0	2.0
OUTFLOW CROWN	=	0.0	5.0	8.0

Table 6.4. Simulation Run 4: Three Conduit Aquifer Configuration.

RUN NUMBERS 5-6

Number of Pipes = 3
Shaft Cross-sectional Area = 100.0 square metres
Initial Aquifer Level = 100.0 metres

PARAMETER		PIPE NUMBER 1	2	3
LENGTH	=	3000.0	3000.0	3000.0
RADIUS	=	0.33	0.5	0.8
VOLUME	=	1026.3	2356.1	6031.8
ROUGHNESS	=	0.33	0.33	0.33
INFLOW INVERT	=	-0.66	2.0	4.40
OUTFLOW CROWN	=	0.0	5.0	8.0

Table 6.5. Simulation Runs 5-6: Reccession Runs, Aquifer Configuration.

RUN NUMBER 7

Number of Pipes = 4
Shaft Cross-sectional Area = 10.0 square metres
Initial Aquifer Level = 1.0 metres

PARAMETER		PIPE NUMBER 1	2	3	4
LENGTH	=	3000.0	3000.0	1000.0	3000.0
RADIUS	=	0.33	0.5	1.0	1.0
VOLUME	=	1026.3	2356.	3142.	9425.
ROUGHNESS	=	0.33	0.33	0.33	0.33
INFLOW INVERT	=	-0.66	-1.0	-2.0	-2.0
OUTFLOW CROWN	=	0.0	5.0	8.0	30.0

Table 6.6. Simulation Run 7: Transient Events, Aquifer Configuration.

RUN NUMBER 8

Number of Pipes = 4
Shaft Cross-sectional Area = 10.0 square metres
Initial Aquifer Level = 1.0 metres

PARAMETER		PIPE NUMBER 1	2	3	4
LENGTH	=	3000.0	3000.0	3000.0	3000.0
RADIUS	=	0.33	1.0	2.0	4.0
VOLUME	=	1026.3	9425.	37700.	150800.
ROUGHNESS	=	0.33	0.33	0.33	0.33
INFLOW INVERT	=	-0.66	1.0	2.0	2.0
OUTFLOW CROWN	=	0.0	5.0	10.0	20.0

Table 6.6. Simulation Run 8: Multi-day Test, Aquifer Configuration

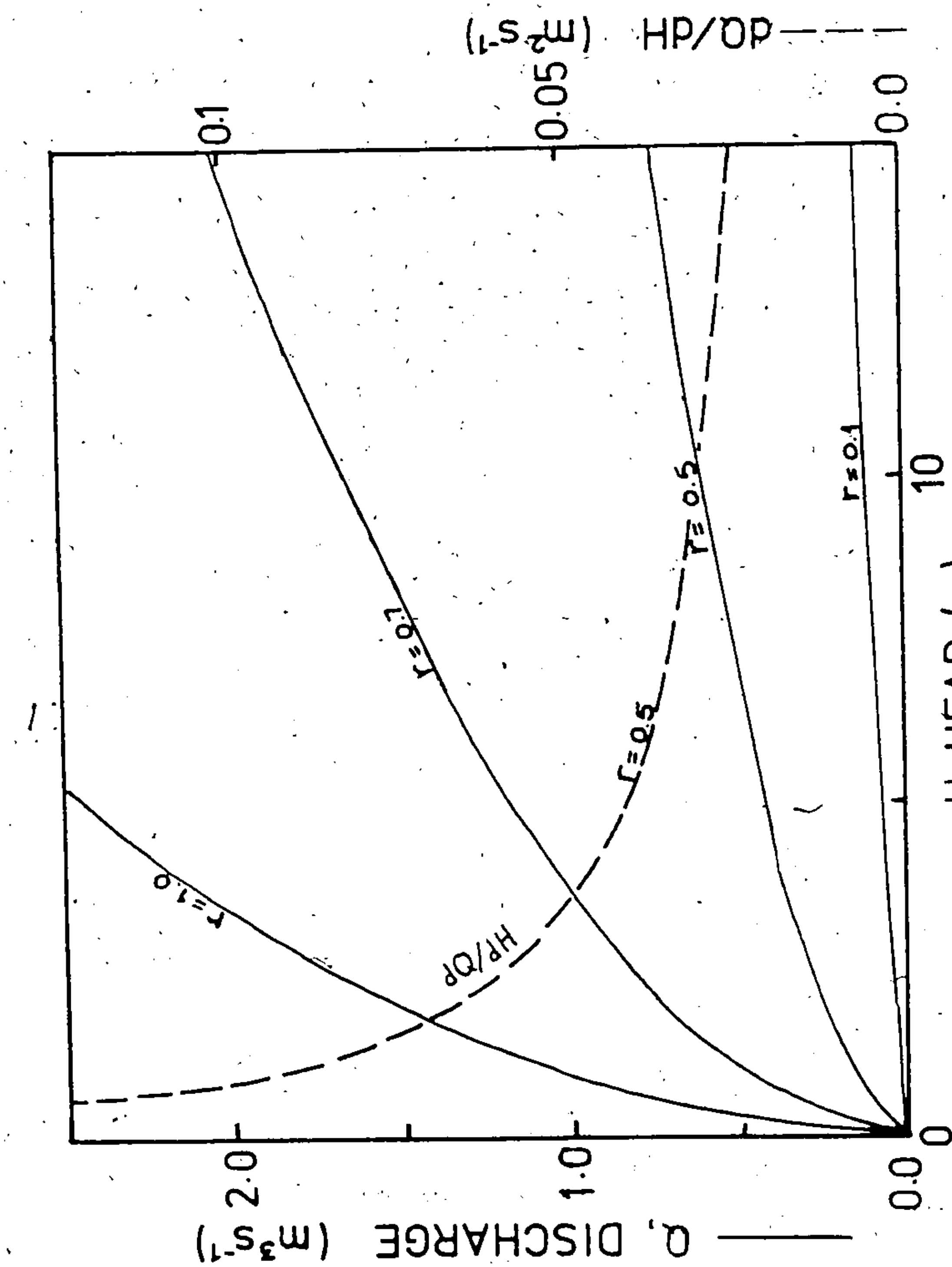


Figure 6.1. Discharge versus head developed on pipes of various radii, and the derivative of discharge with respect to head for a 0.5m radius Conduit

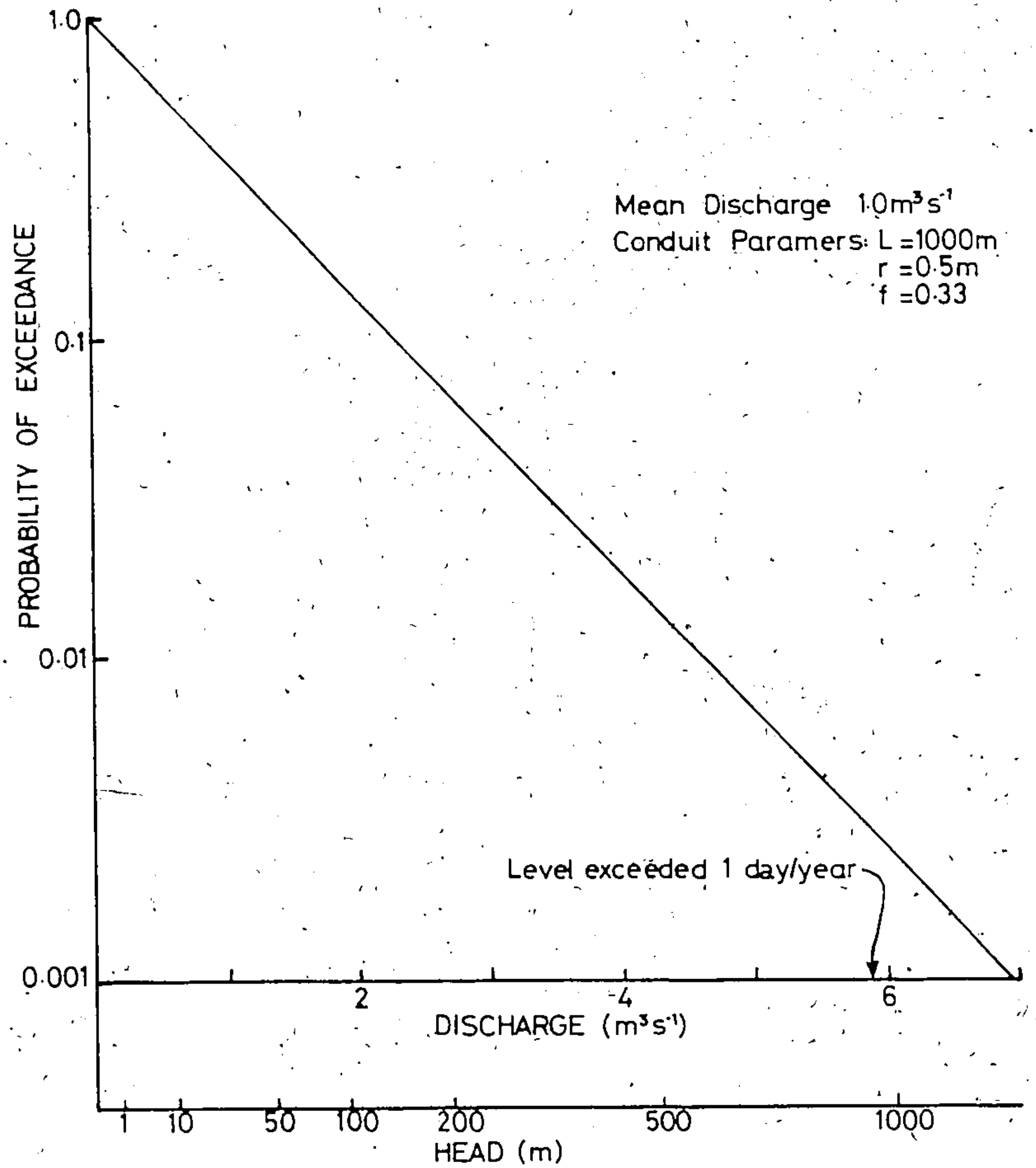


Figure 6.2. Probability density function for an exponential distribution of discharge exceedance with mean discharge 1 cubic metre per second, and the equivalent head developed on a conduit of 0.5m radius

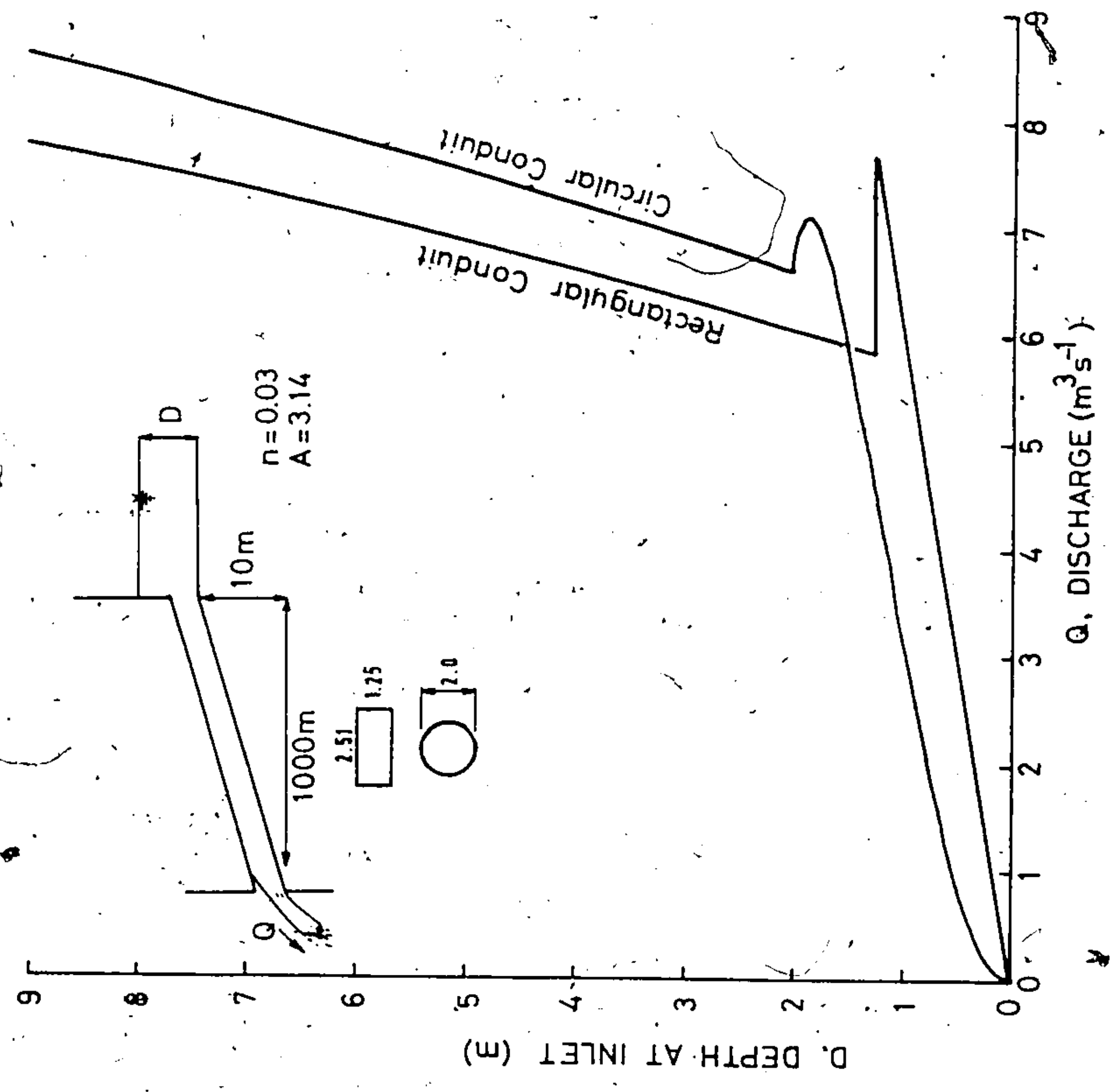


Figure 6.3. Head developed in response to discharge at the inlet of two simple conduits of different shape undergoing a transition from open channel to closed conduit flow

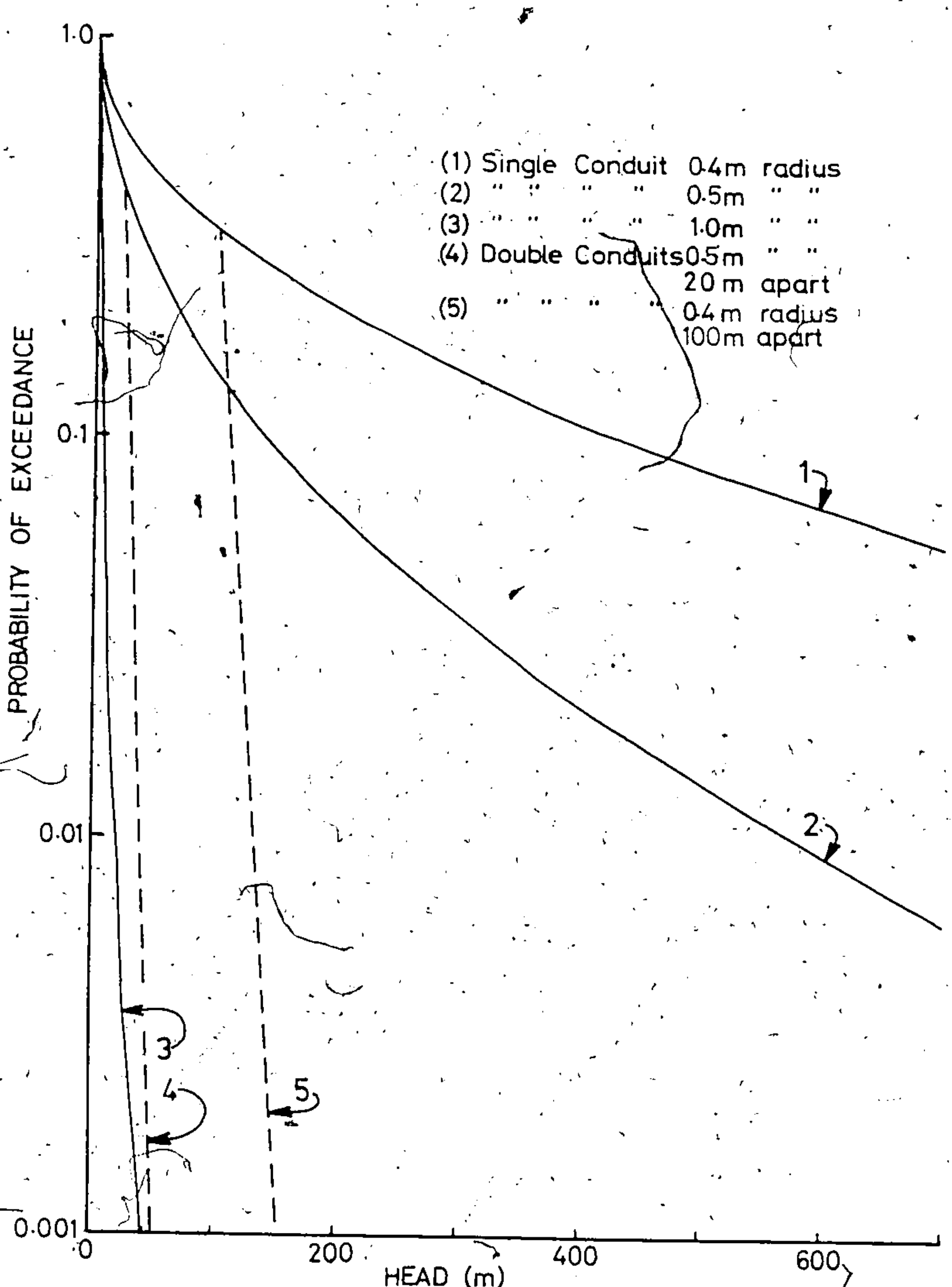


Figure 6.4. Probability density function of head on a simple conduit system for an exponentially distributed discharge exceedance, for various conduit sizes and configurations

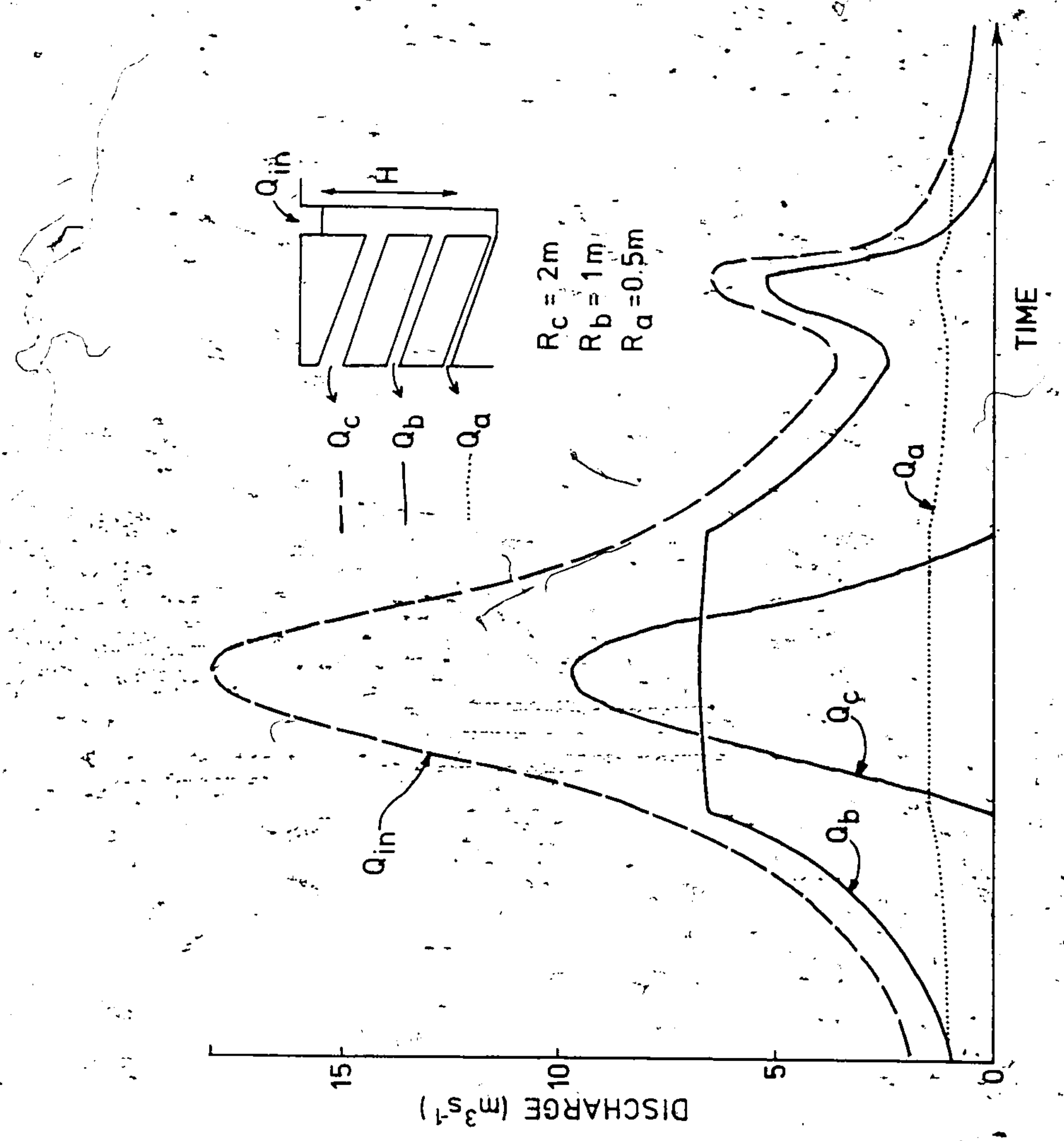


Figure 6.5. Static system model (no storage considered) for a three-component underflow-overflow system responding to an arbitrary input hydrograph

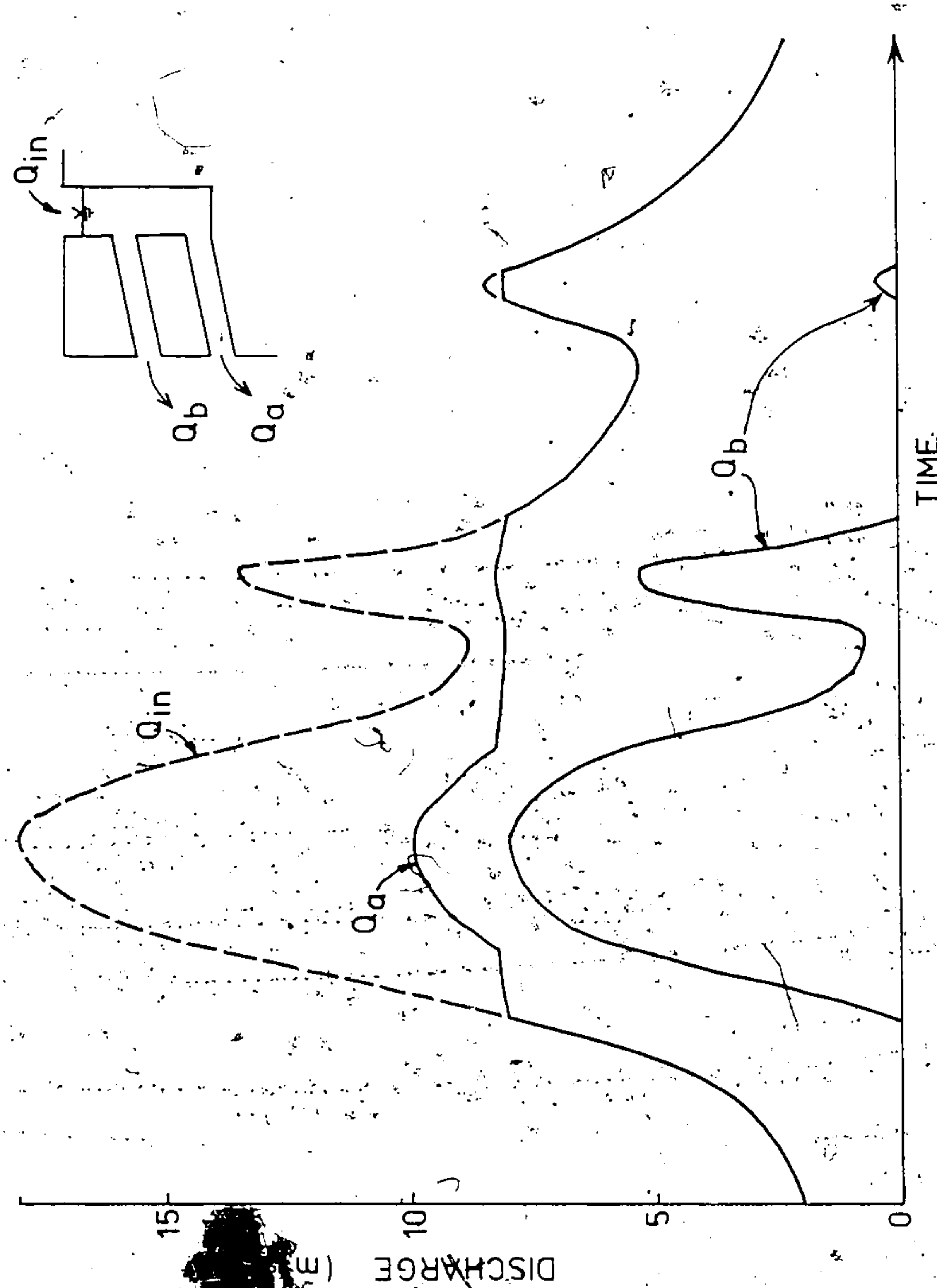


Figure 6.6. Staged system model for a two component, open-closed channel, underflow-overflow system responding to an arbitrary input hydrograph

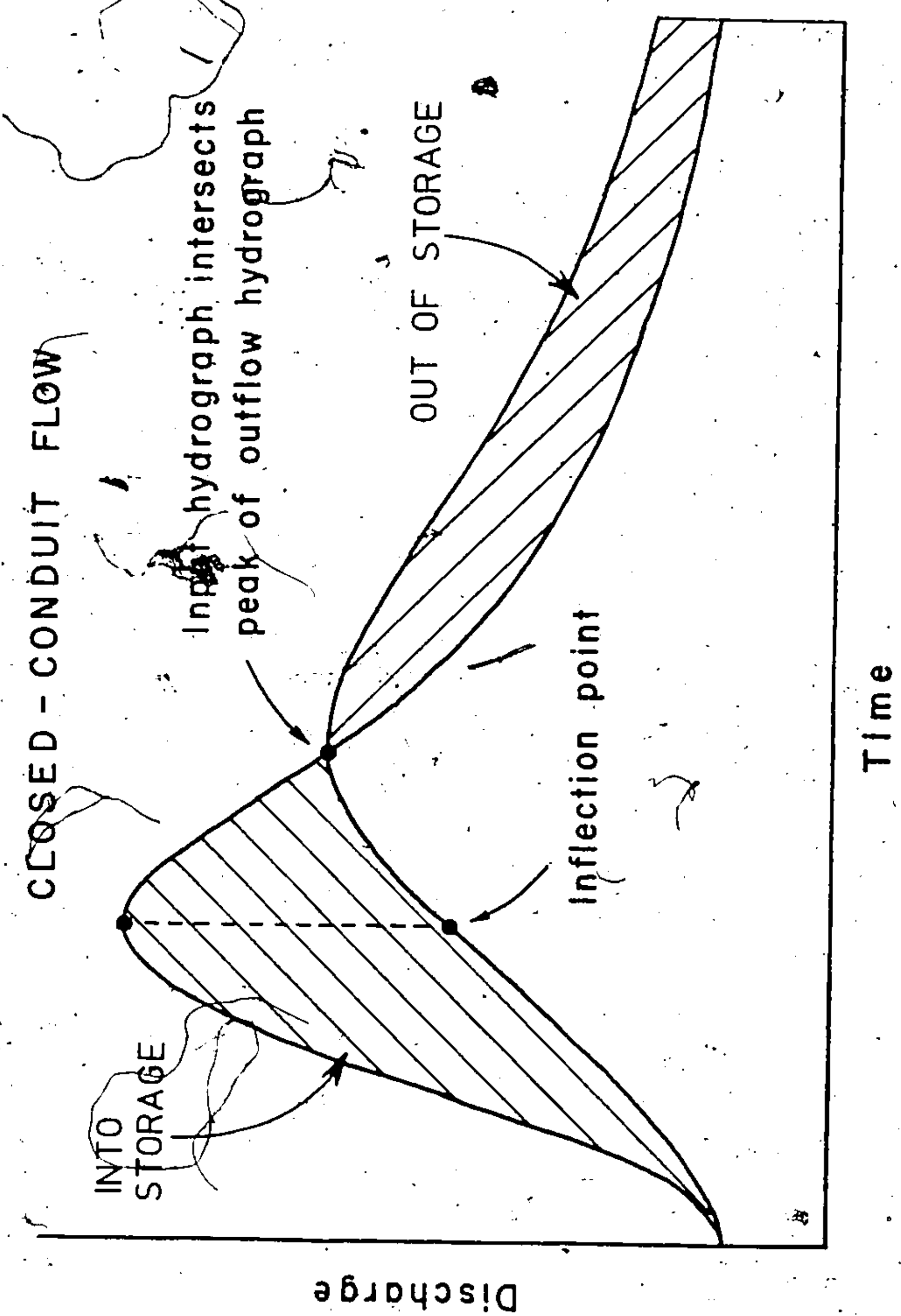


Figure 6.7. Delay of transmission of a flood pulse through a closed conduit by epiphreatic storage (after Palmer, 1981)

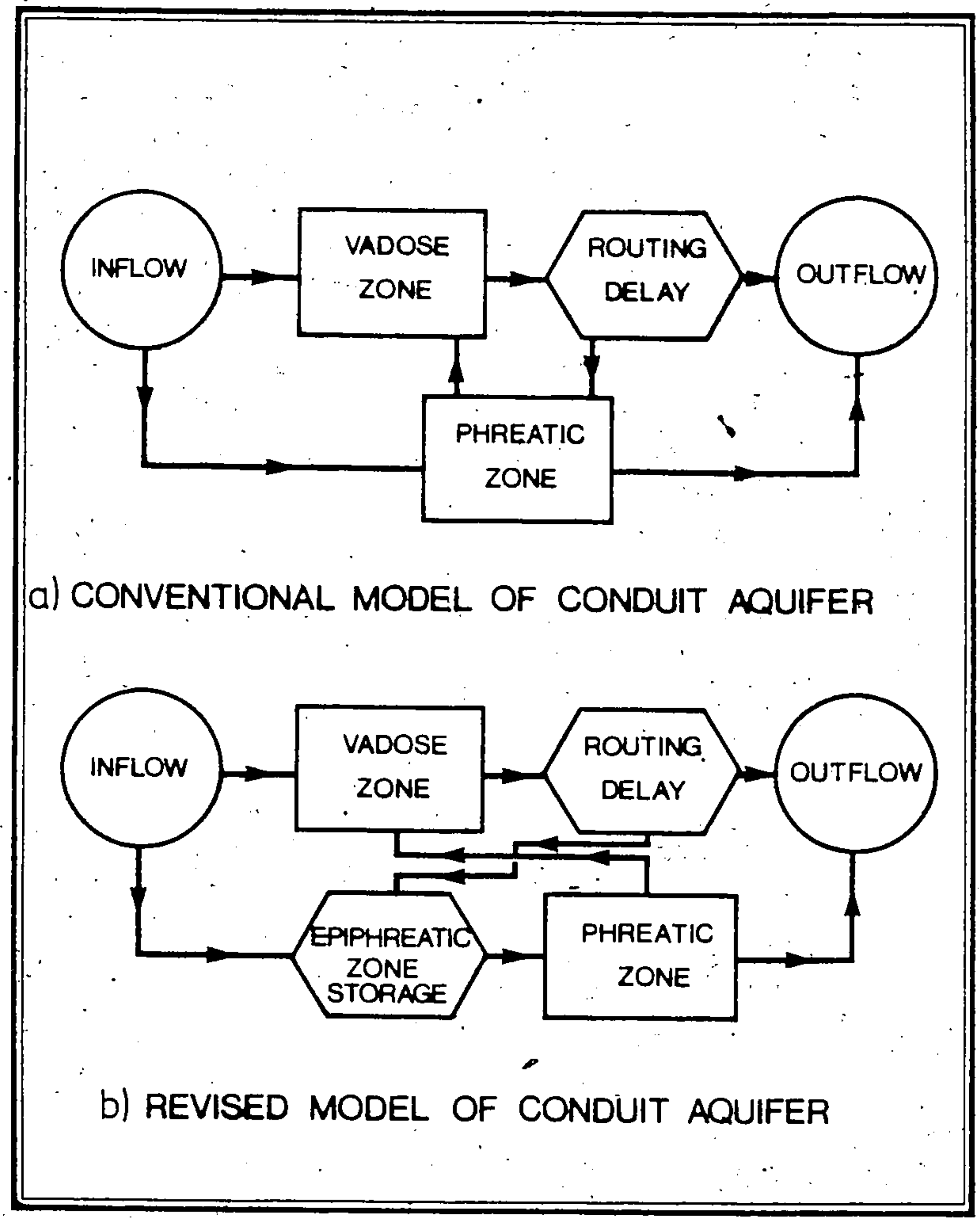
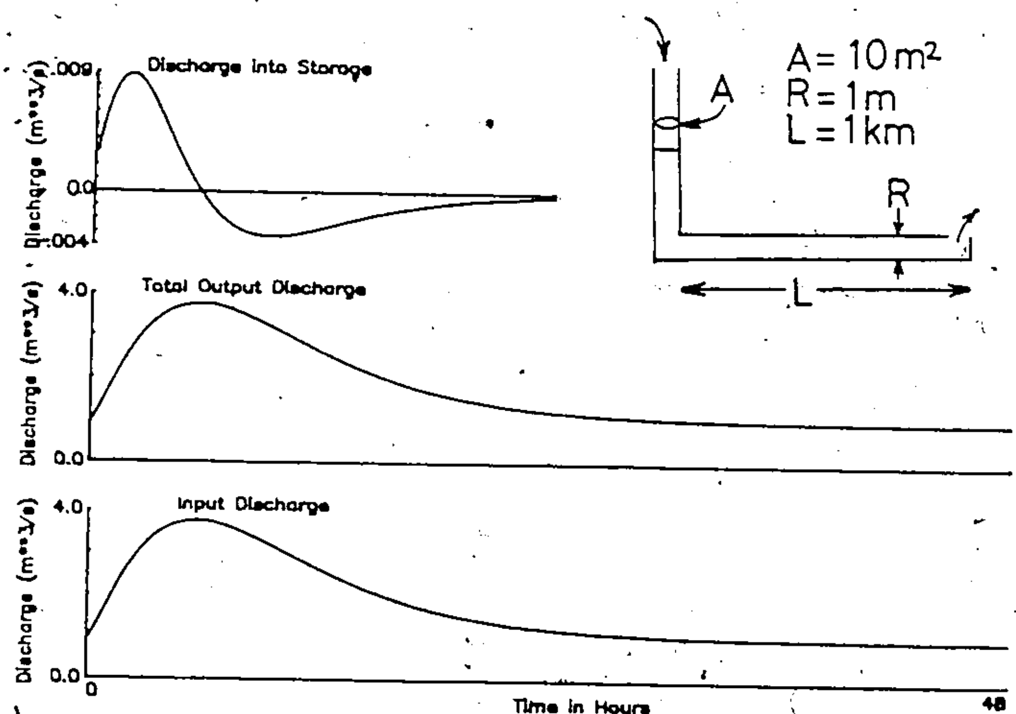
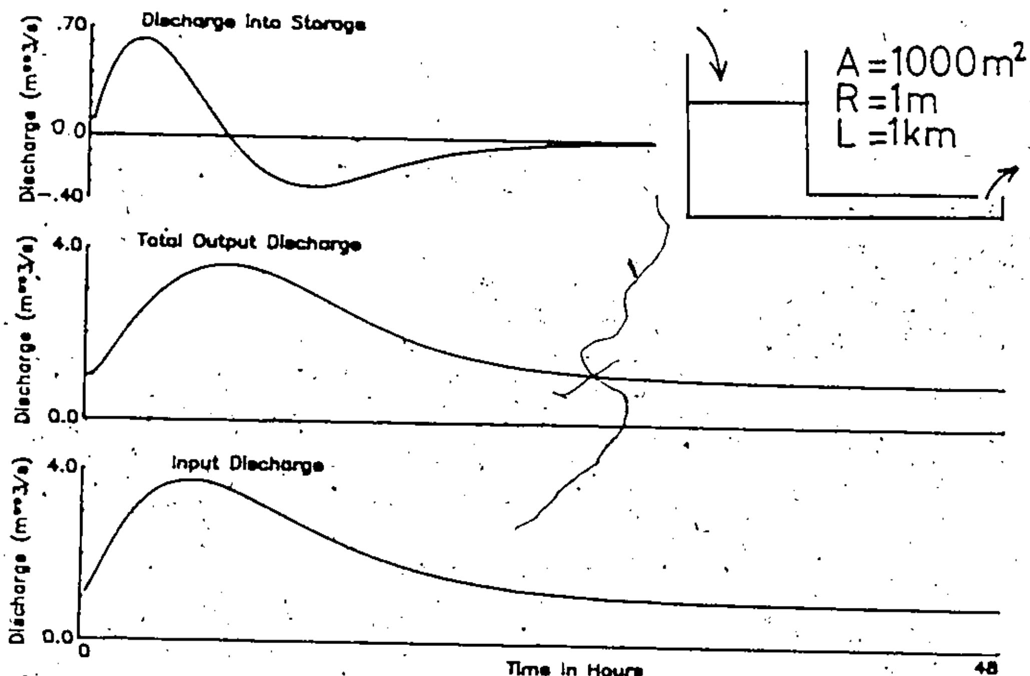


Figure 6.8. Conventional and revised models of a conduit aquifer in pulse train analysis



(a)



(b)

Figure 6.9a. Run 1: Passage of a flood through a single conduit with limited epiphreatic storage

Figure 6.9b. Run 2: The same conduit with extensive epiphreatic storage

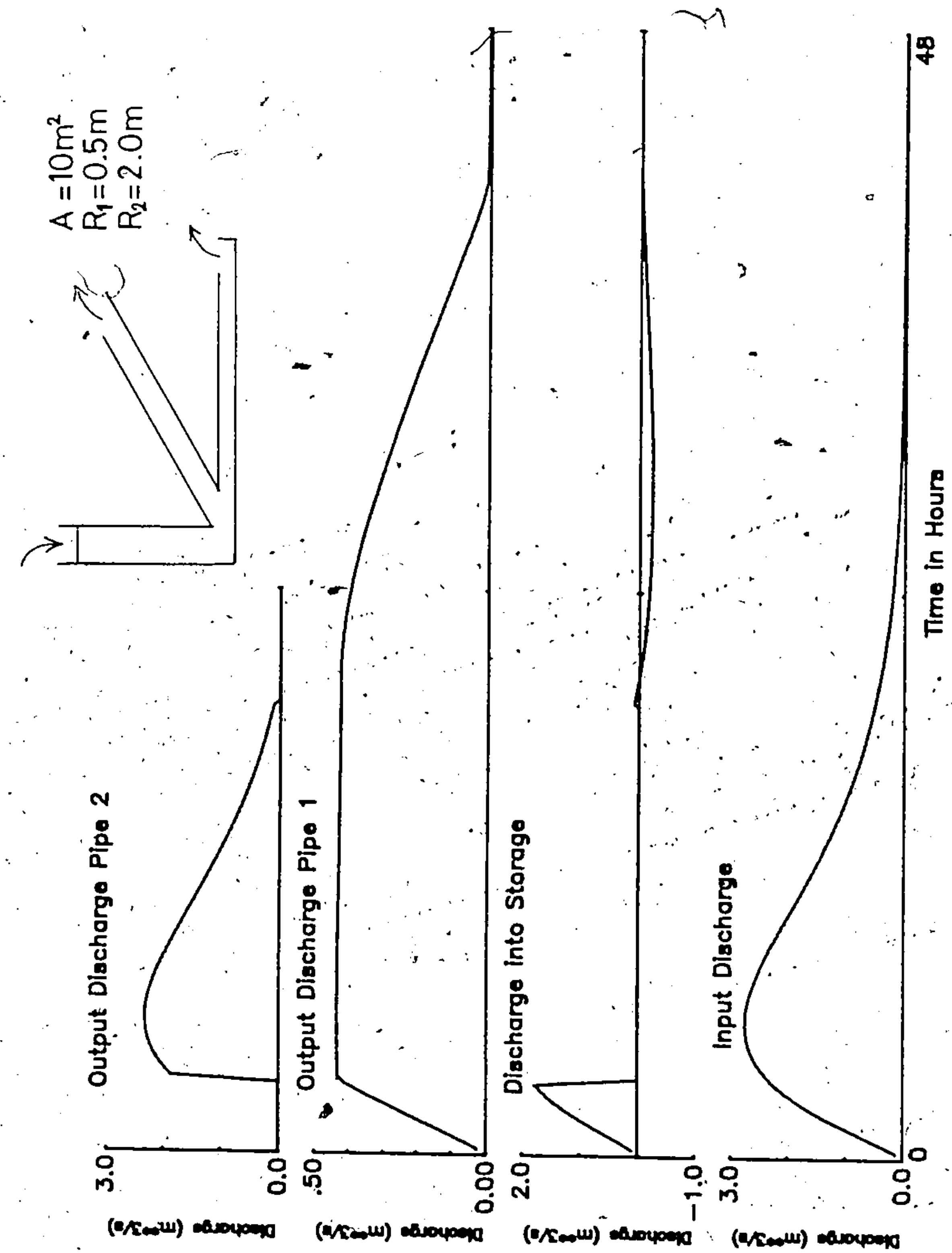


Figure 6.10. Run 3: Passage of a flood through a two conduit aquifer

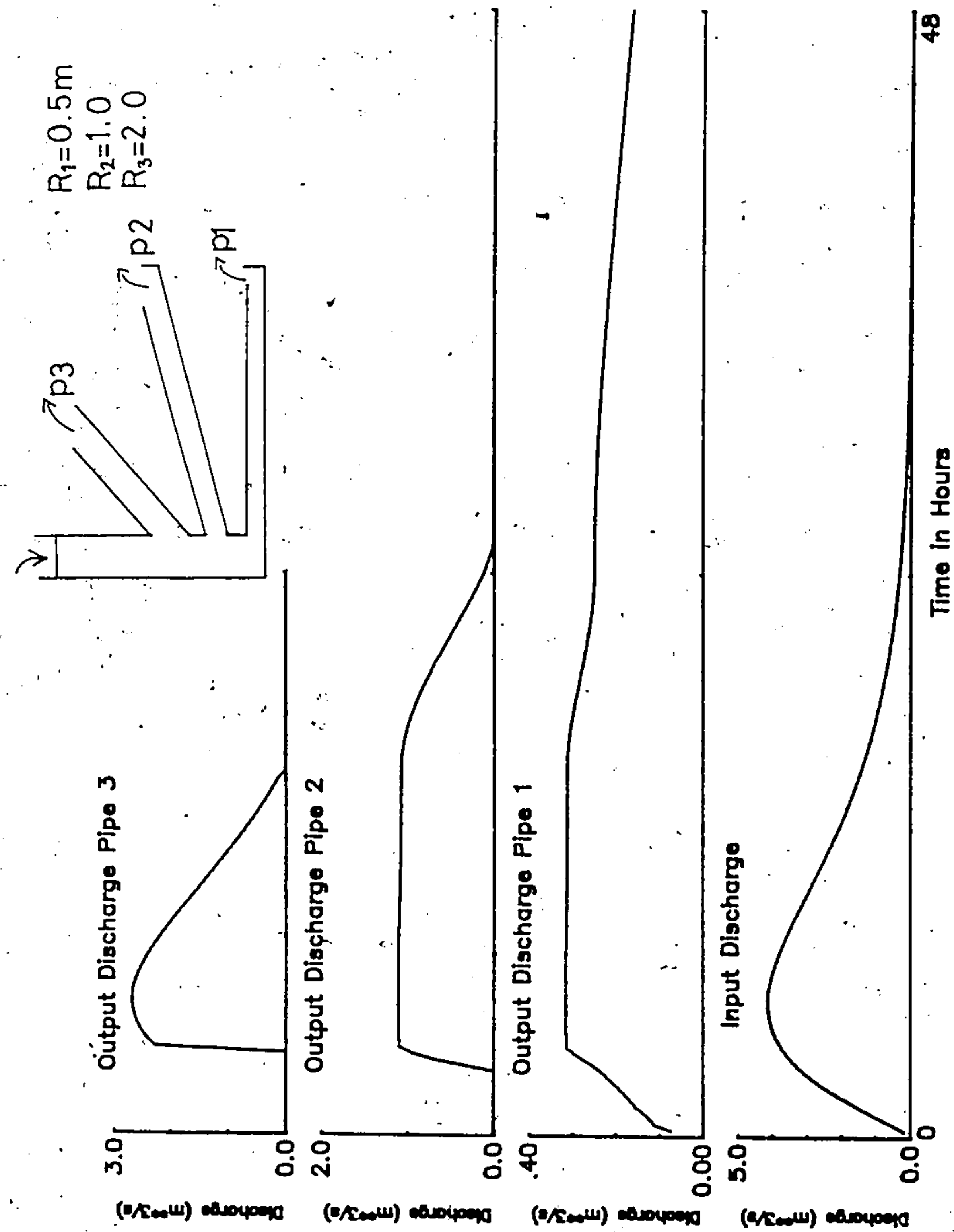


Figure 6.11. Run 4: Passage of a flood through a three-conduit aquifer

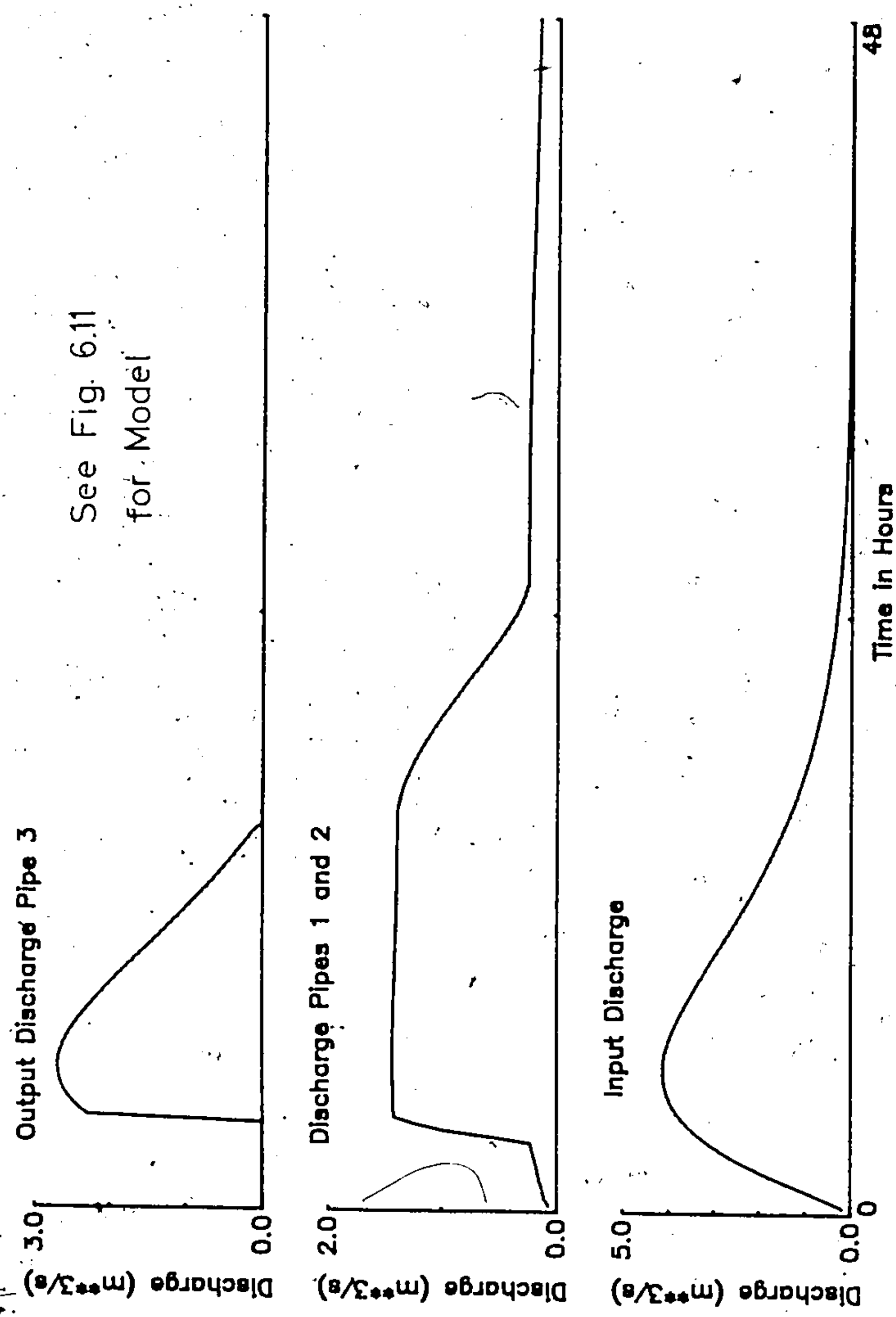


Figure 6.12. Run 4: Complex output hydrographs produced by additive combination of the output from the two inferior conduits

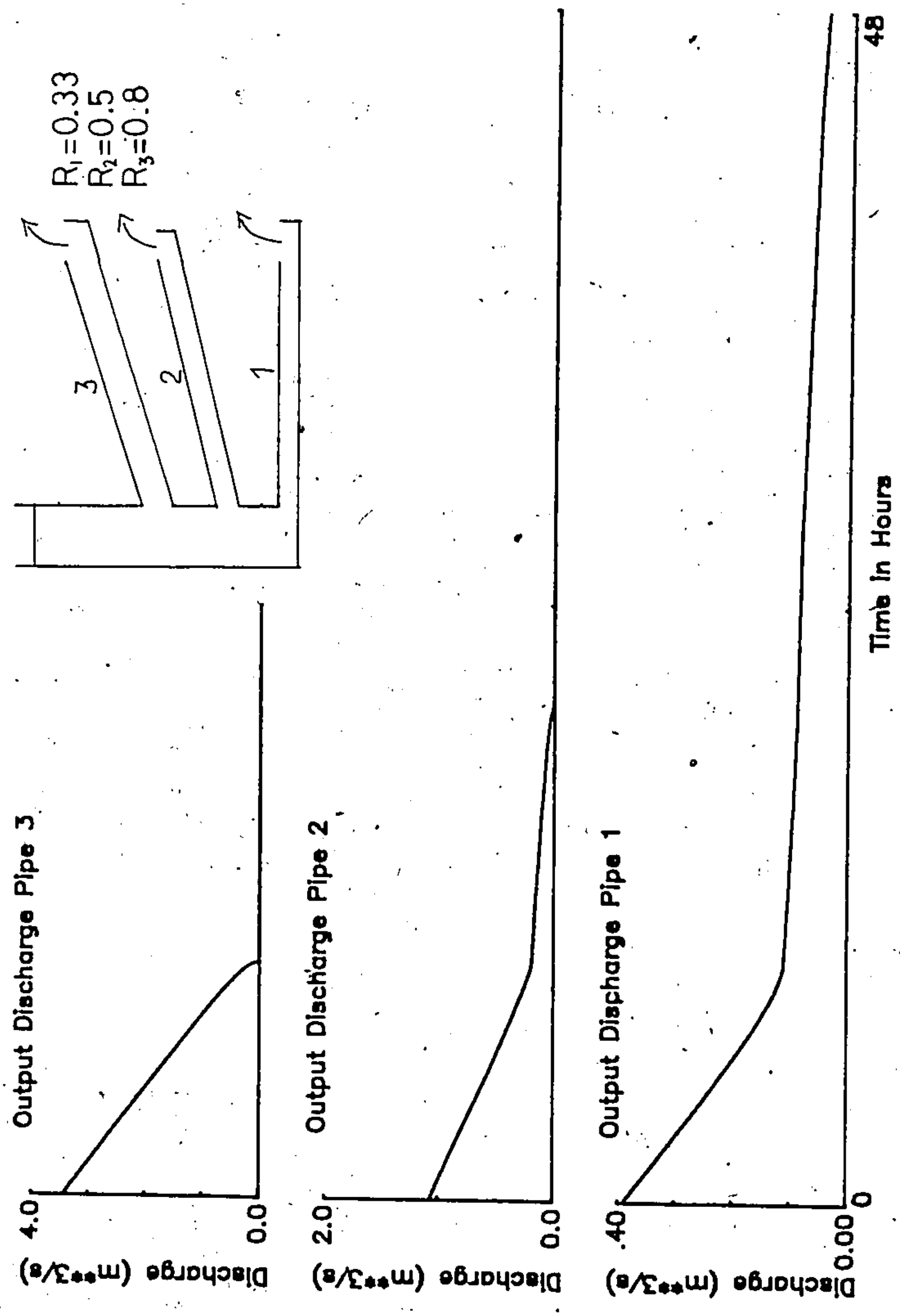


Figure 6.13. Run 5: Passive drainage of a three-conduit aquifer

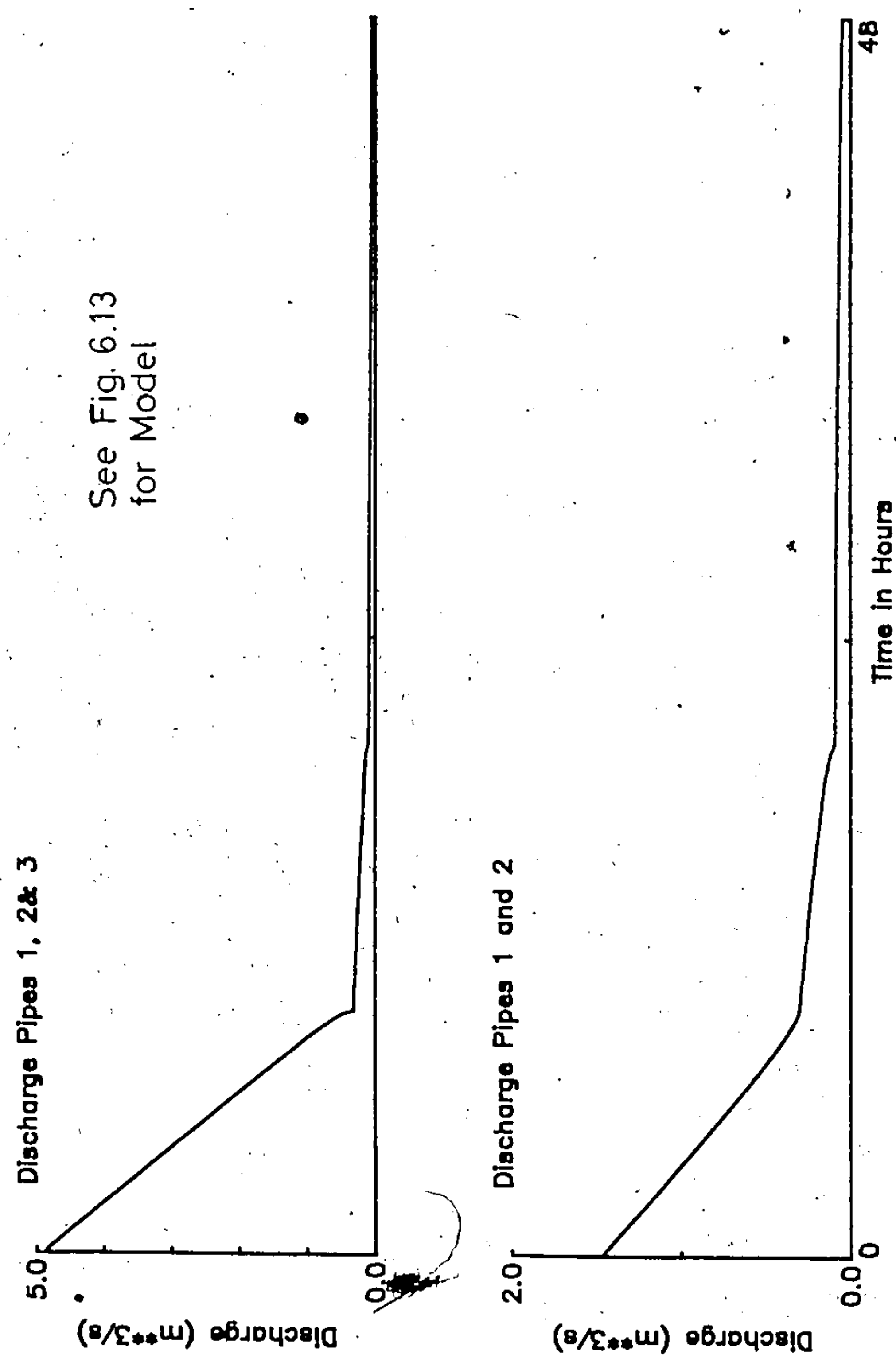


Figure 6.14. Run 5: Combined output of conduits 1 and 2, and 1, 2 and 3 under passive drainage

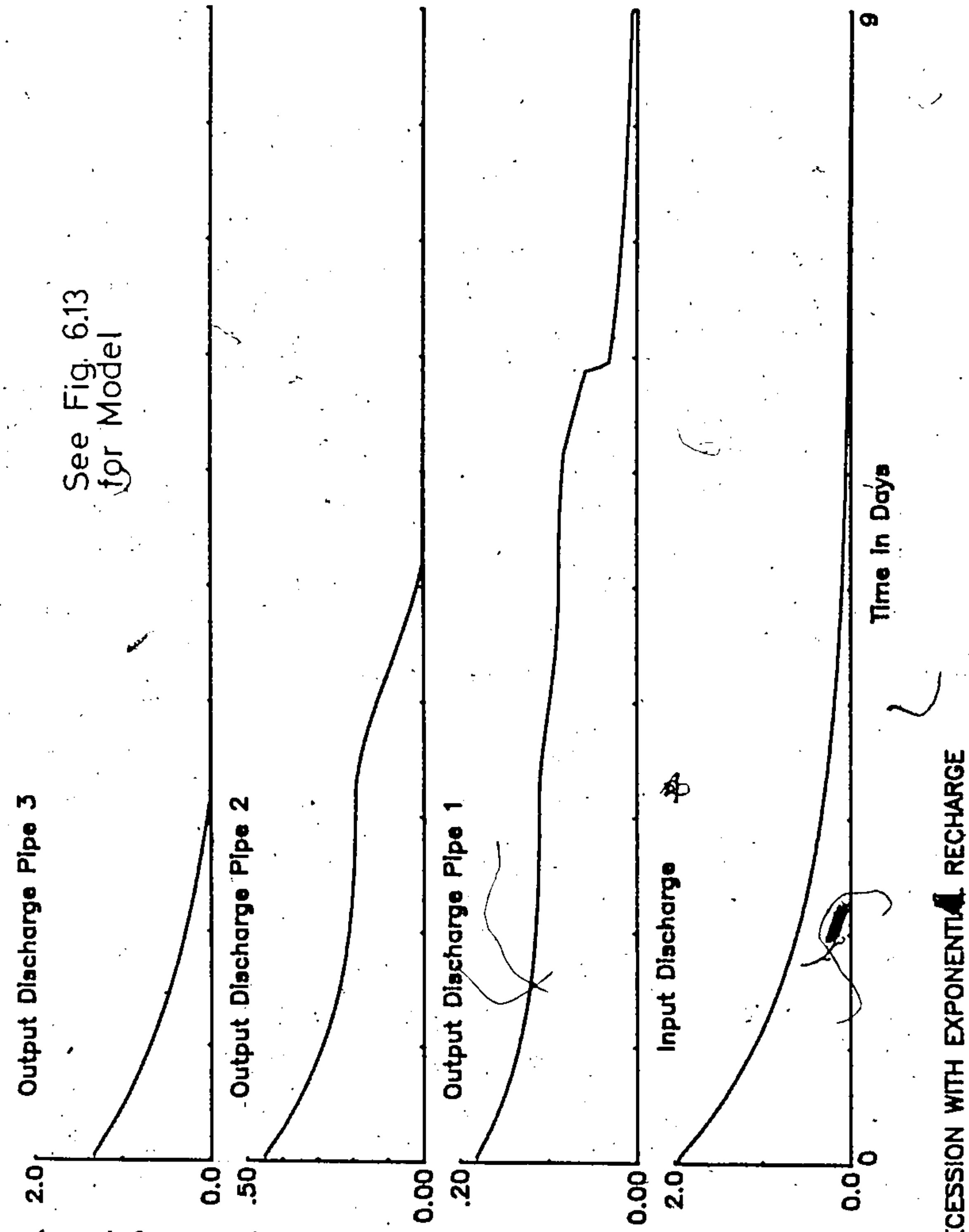


Figure 6.15. Run 6: Exponentially decaying recharge and passive drainage of a three-conduit aquifer

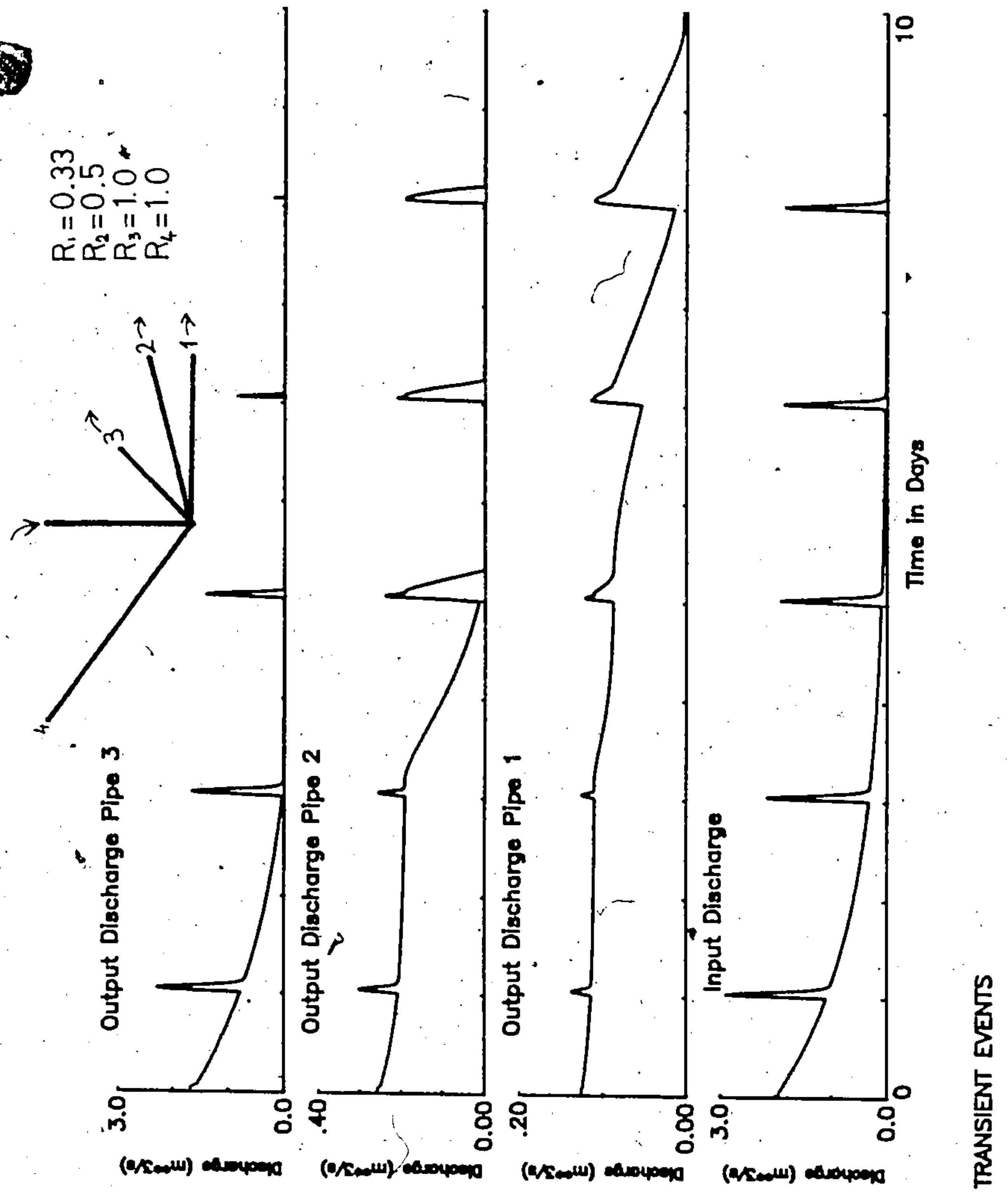


Figure 6.16a. Run 7: Transmission of transient floods through the outlets of a three-conduit aquifer under various baseflow discharges

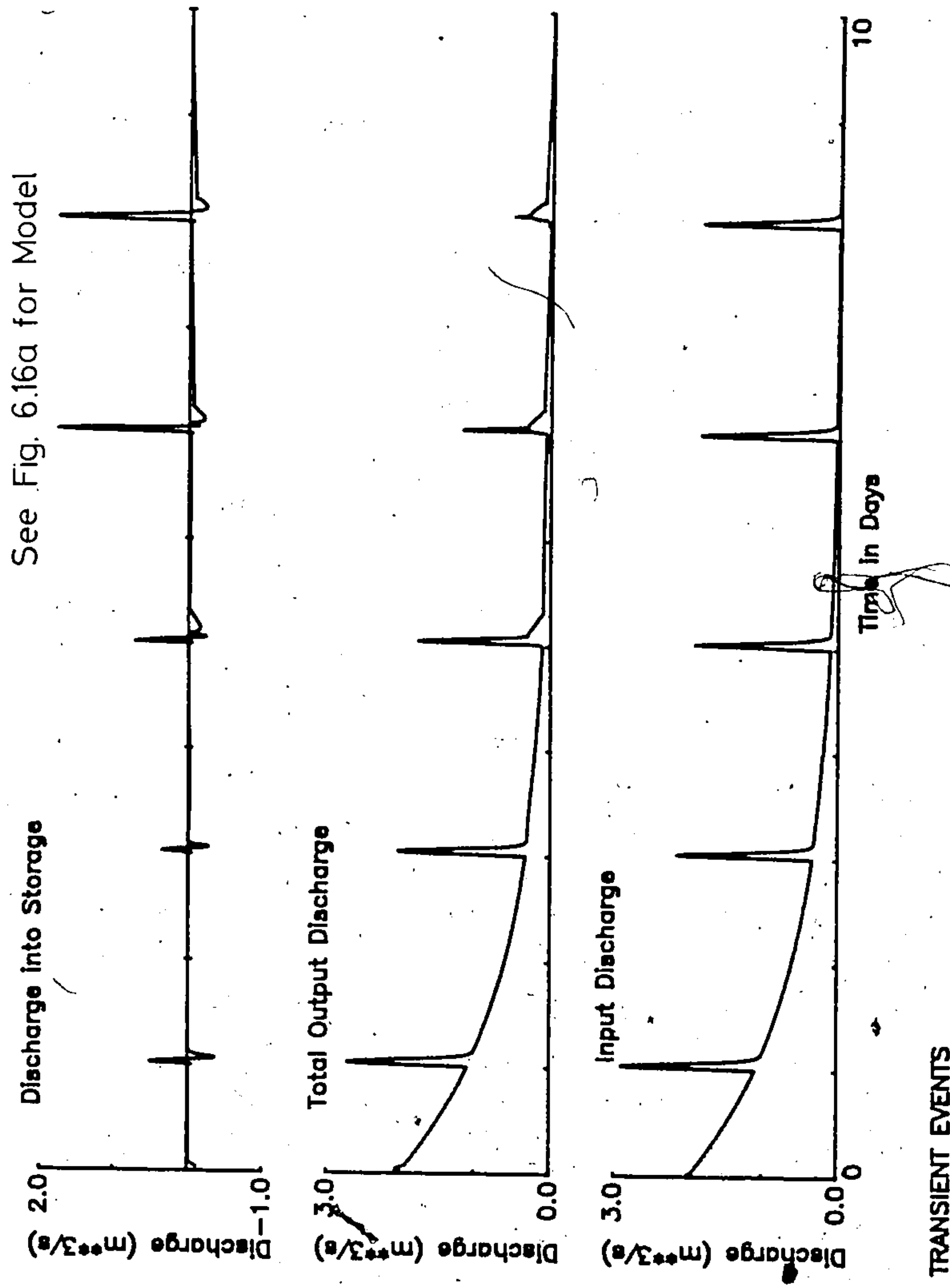


Figure 6.16b. Run 7: Inflow, total outflow and change in storage for transient events

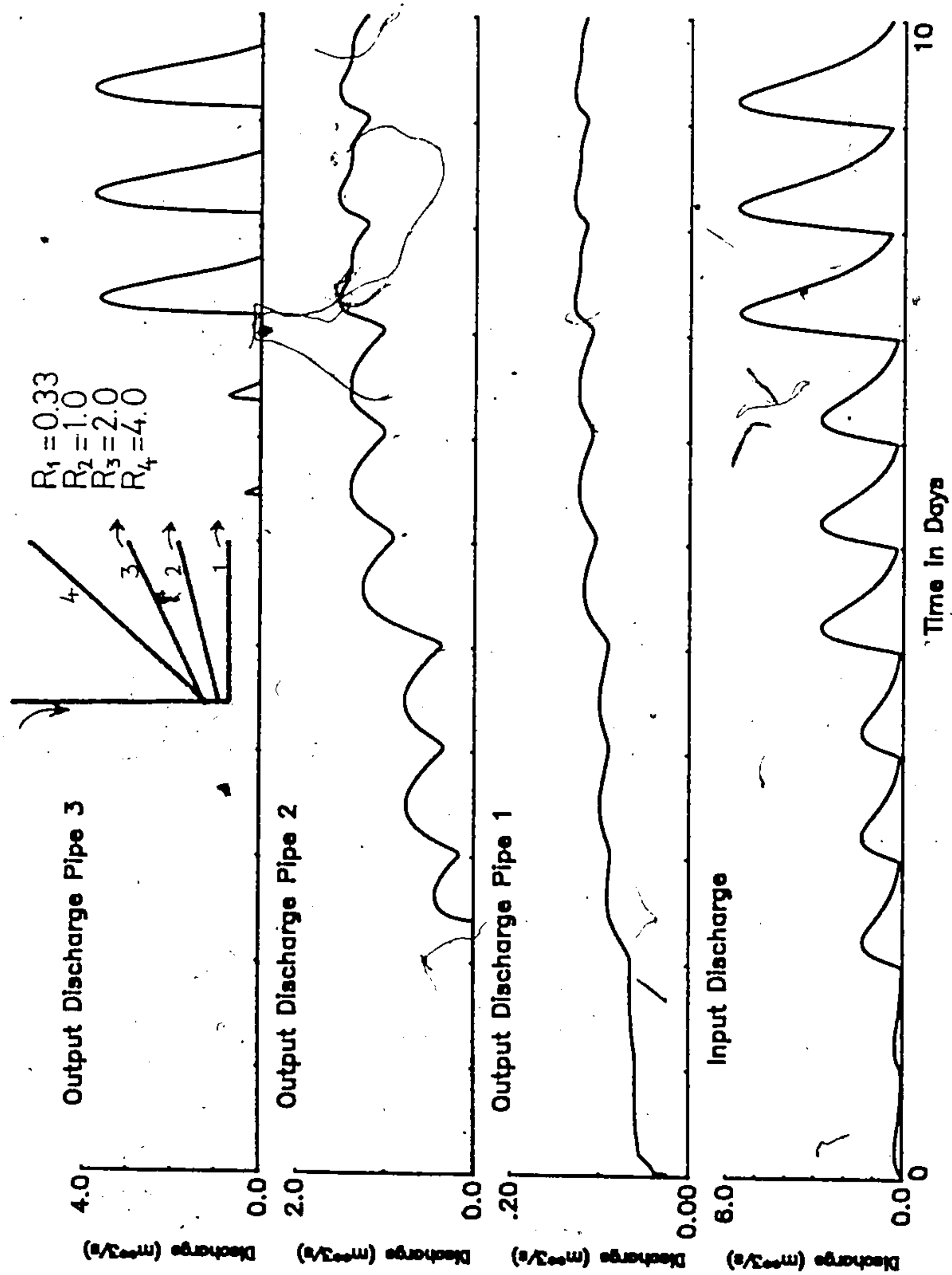
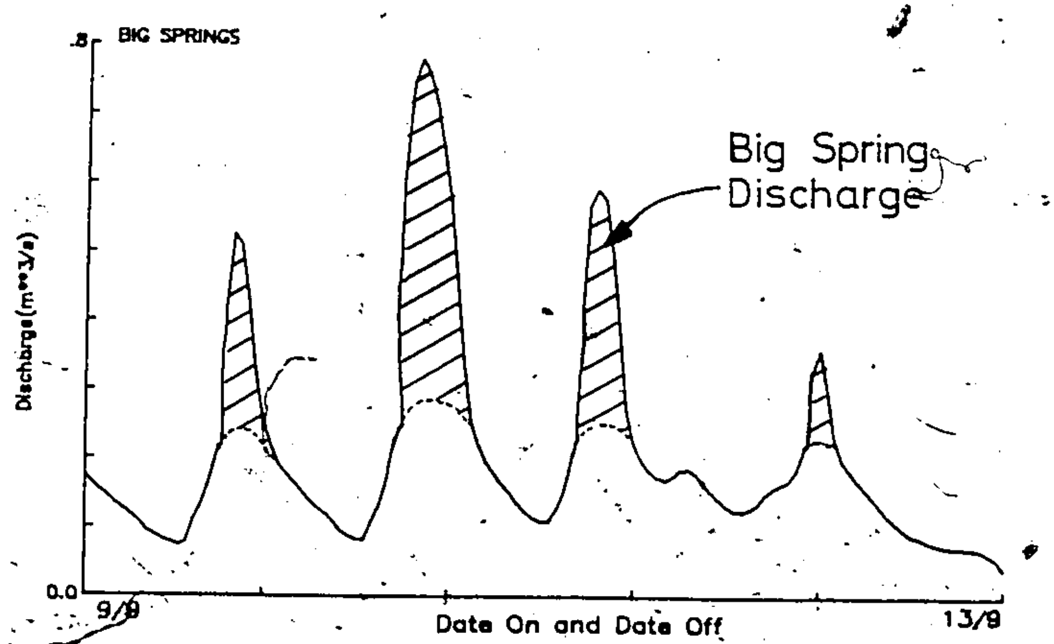


Figure 6.17. Run 8: Multi-day simulation with a four-conduit aquifer and a stepwise increase in the magnitude of the input hydrograph



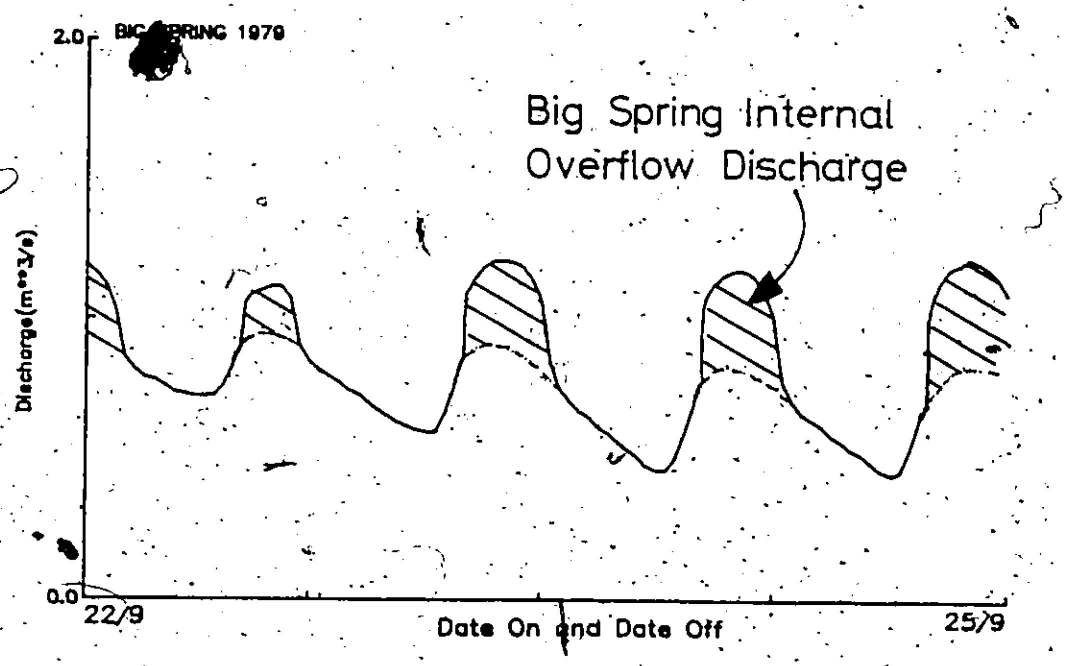


Figure 6.18. Selected portions of the Big Spring hydrograph when overflow-underflow events are occurring

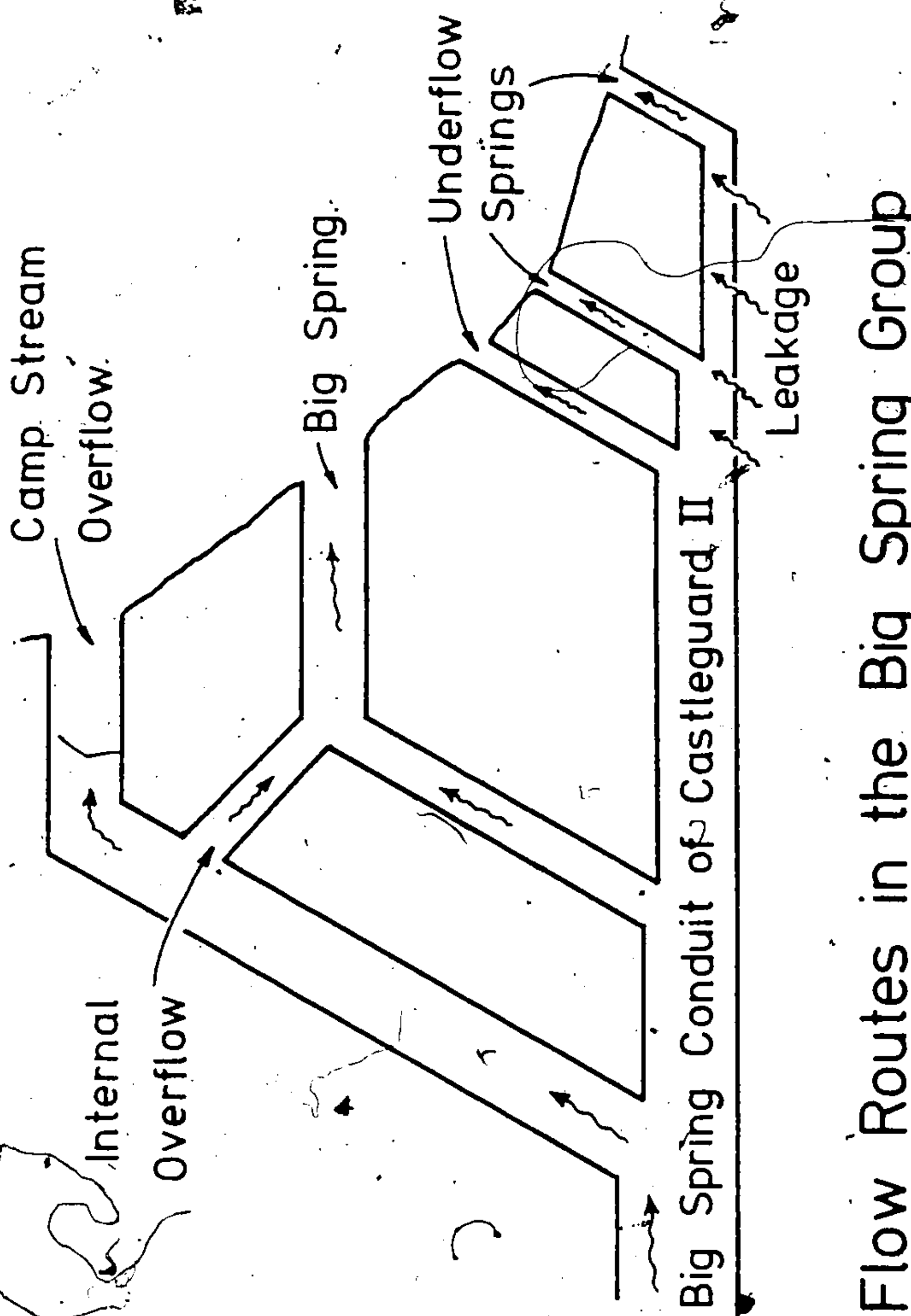


Figure 6.19. Internal conduit network of the Big Spring inferred from hydrographs in Figure 6.18

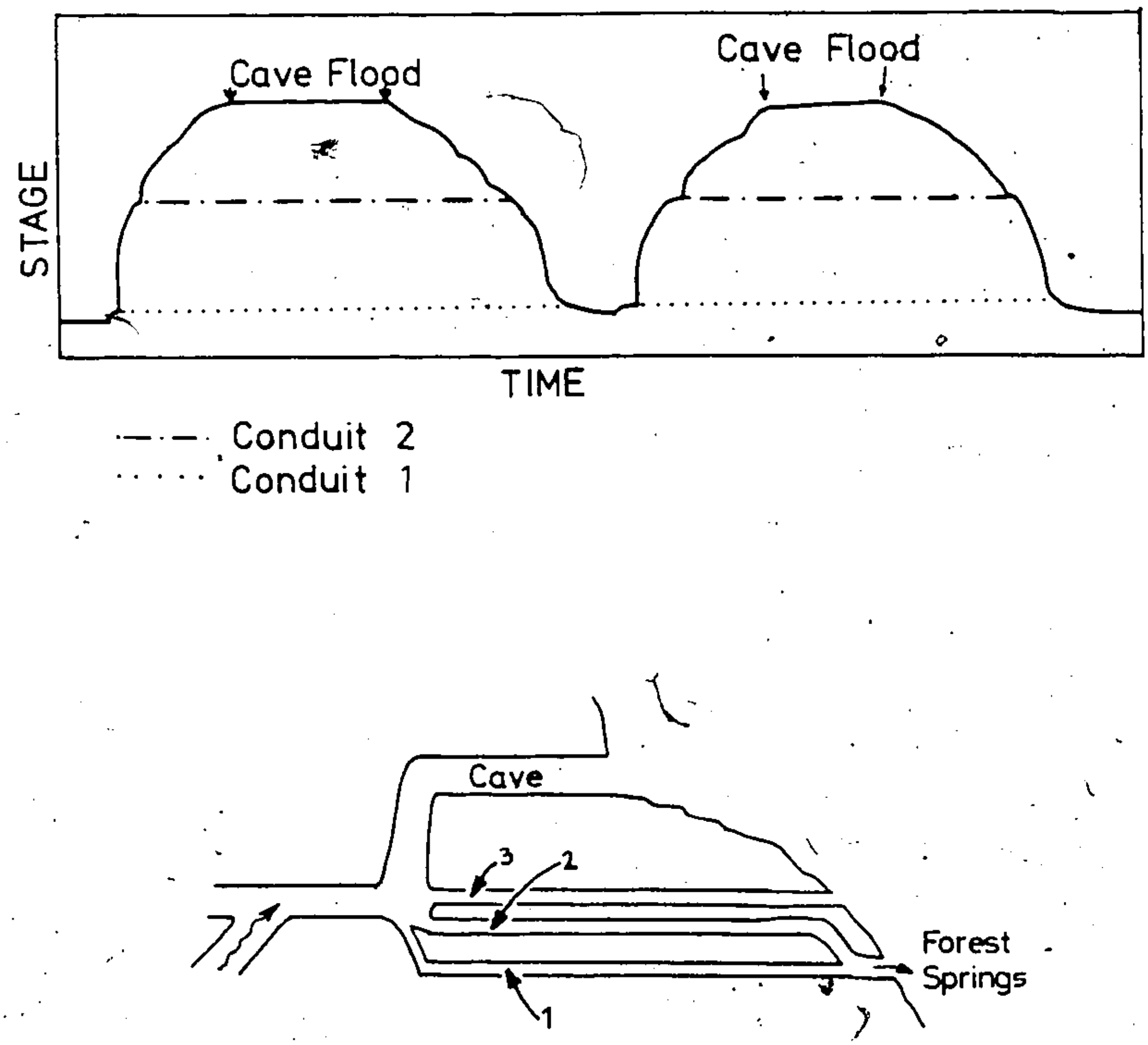


Figure 6.20. Segment of the stage record of the Forest Spring and the conduit network inferred from it

CHAPTER SEVEN

HYDROLOGY OF A GLACIERISED ALPINE KARST

7.1 GLACIER - GROUNDWATER INTERACTION

The loss of large quantities of glacial meltwater into karst has previously only been inferred on hydrological, hydrogeological and geomorphological grounds (e.g. Ford 1971a, Lauritzen 1981, Maire 1978). The dye trace from the Saskatchewan Glacier to the Big Spring is the first conclusive proof that such transfer is not only possible, but can be relatively rapid and complete. Observations in Castleguard Cave have suggested great hydrological activity in summer in the vadose zone beneath the Columbia Icefield, with streams sufficiently large to convey clasts at least 10 cm in length into the aquifer. In addition, the seepage water forming speleothems must also be glacigenic.

The discharge of the springs is characterised by diurnal fluctuations which are closely related to daily air temperature in the region. The daily discharge oscillations may represent proglacial flow superimposed on a more steady flux from the icefield. However, the steady chemical composition of the springs suggests that such mixing is probably insignificant and that the Icefield waters are themselves released in a diurnal cycle. The large volume of water contributed in such a way from the Icefield can only come from supraglacial melt, in which case a well developed, conduit drainage system is operating in the ice body itself, and the firn aquifer has relatively little moderating influence on discharge.

The relative clarity of the spring water suggests that the glacier bed component of the "ice-karst" aquifer is either short or fixed in position at the glacier bed (Collins 1979, Metcalf 1979), because more extensive contact with the glacier bed would allow waters to remove subglacially generated sediments. These channels exist all winter, because they then convey the perennial wind up and out of the cave. These qualities are most likely associated with "Nye Channels", probably following steps formed on major joints in the bedrock.

The cave wind is perennial, and under suitable thermal conditions is only stopped by complete flooding of the entrance passages (R.Baldwin pers.comm. 1974). Therefore, the subglacial channels are not only perennial, but vadose under all observed conditions, corresponding to the subglacial streams reported by Hodge (1979) and Engelhardt

(1978). Ford (1979, 1983a) hypothesised that an icecap superimposed on a karst aquifer could enhance karstification by increasing the hydraulic head. This is clearly not the case at present in the subglacial streams above the headward complex of Castleguard Cave. However, the high pressure subglacial film which has been shown to coexist with discrete conduits beneath certain glaciers will indeed enhance the general hydraulic head over large areas of the aquifer.

The ability of the subglacial film waters to enter the groundwater system is unproven. However, the drip waters forming speleothems in the cave beneath glacier ice are clearly of a different chemical character to the erosive shaft-forming waters, and may represent percolation of film water through bedrock fractures. Independent evidence for passage of these waters through bedrock was found on the lee faces of bedrock steps in recently exposed proglacial areas. Here, small (1 cm in length 2.5 cm diameter) "stalactite" forms are found on the rock face, overlying fine cracks in the host bedrock. These features are analogous to conventional subglacial precipitates formed by freezing of basal water in the lower pressure areas at the glacier bed, except that the water has penetrated through the bedrock rather than following the ice-rock interface (Figure 7.1). A cursory inspection of a thin section of a specimen stalactite revealed none of the clastic particles reported by Hallet (1979) from conventional precipitates, suggesting that these might have been filtered out in passage through the bedrock. Given that water is mobile in fissures at least locally beneath glaciers, then penetration to greater depths is possible, especially if a low pressure, air filled cavern exists beneath the glacier.

Most calcite speleothem deposition is the result of supersaturation of drip water caused by degassing of enhanced levels of carbon dioxide on entering the cave atmosphere. The high levels of carbon dioxide are usually attributed to biological activity in the regolith, but this can not be the case beneath the ice at Castleguard. Atkinson (1983), Dreybrodt (1982), Gascoyne and Nelson (1983) and C.C.Smart (1981) have each proposed contrasting hypotheses for the supersaturation of dripwaters in the subglacial parts of the cave. In the absence of sufficient relevant measurements, these ideas remain speculative. However, the mechanisms of both Atkinson and Dreybrodt imply supersaturation with respect to calcite within the percolation network. Gale and Reardon (1981) have shown such processes to gradually decrease the permeability of a grouted fracture, and this may render these processes inherently self-limiting in the natural state.

Atkinson (1979) measured partial pressures of carbon dioxide in drip waters in Castleguard Cave, and found them to be essentially the same as equivalent local atmospheric values. He used these data to reject a conventional degassing mechanism for precipitation. Nevertheless, such carbon dioxide levels are in excess of those reported for glacigenic waters (Ek 1964, 1965, Ford 1971a,b, Miotke 1974), and demand an independent mechanism for their generation. If the measured drip were very slow (making sampling impossible), then an atmospheric carbon dioxide level could be gained by back diffusion, in which case equilibration by degassing could equally well have reduced previously elevated carbon dioxide levels. Gascoyne and Nelson (1983) suggest that the oxidation of organic material as the source, citing carbon isotopic evidence. However, the presence of such organics has not been demonstrated and the process of oxidation also requires oxygen which is probably as depleted as carbon dioxide in glacigenic waters. In addition, Jouzel and Souchez (1982) have demonstrated that a remarkable range of isotopic fractionation is theoretically possible in the basal regelation film, making conventional interpretation of isotope ratio impossible.

The mechanism proposed by C.Smart (1981) is based on an observation by Hallet (1977), that the high ambient pressure at the glacier bed will cause high partial pressures of carbon dioxide in any bubbles in the ice. These bubbles are known to exist in basal ice (Kamb and LaChappelle 1968), but their carbon dioxide content and the bulk carbon dioxide-water ratio are unknown. Assuming a value of 1% for the carbon dioxide concentration in basal ice bubbles (Weiss et al.1972), and water equilibrated to this level, a model system is illustrated in Figure 7.2, in which massive supersaturation occurs when the percolation water encounters the cave atmosphere.

These hypotheses need testing by field measurements. However, apart from the logistical difficulties of working seven kilometres into Castleguard Cave, the seasonality of subglacial hydrology may invalidate any single set of measurements. One soda straw stalactite from the cave showed distinct rhythmic banding which may reflect seasonality in the hydrochemistry. Nevertheless, apart from these speculations it seems possible that the hydrology of subglacial karst groundwaters reflects the independence of subglacial conduit water and regelation film water.

7.2. SUBGLACIAL KARST DEVELOPMENT

The demonstration of hydrological continuity between ice and bedrock is necessary, but insufficient, evidence for subglacial karst initiation. While the percolation waters are possibly quite aggressive, the conduit waters have a remarkably low solution capacity. The low saturated hardness of valley spring water demonstrates that even atmospheric levels of carbon dioxide are not reached, despite much of the vertical range of the aquifer being apparently vadose. Clearly, the rapid flow-through time is insufficient for full equilibration, compared to the equally high altitude Terrace Mountain Aquifer in which water hardness reaches atmospheric equilibrium.

Glacial conduit water will reach bedrock with most of its limited solutional capacity still available. However, the spray and turbulence generated in the descent of vadose shafts and steep canyons will expend this potential very rapidly indeed (although this will also promote uptake of additional carbon dioxide from the air). Even the largest shafts as yet descended in Castleguard Cave constrict and become impassable at depth, suggesting a rapid exhaustion of solution capacity.

Palmer (1981a) has developed approximate growth rate equations for karst conduits. Discharge is important only in so far as the growing tube should always be water-filled, and beyond a certain maximum, growth rate is independent of discharge. However, the degree of undersaturation of the water is critical. As a result, Palmer (p.121) "suggests that the LENGTH OF TIME a passage carries water is more important than discharge in forming a large cave." Lauritzen (pers.comm.) has demonstrated that for meltwaters like those of the Big Spring, this time is longer than the duration of most glaciations. For example, assuming a minimal solution capacity of 1 mg/l for waters reaching the Castleguard II proto-conduit, ages ranging from 27 million to 2,700 million years are required for it to develop to one metre diameter. In addition, the majority of the solutional potential of the conduit waters will be expended on suspended sediments. Although these figures are crude, they do suggest that ice-free interglacial conditions are generally necessary for the caves to develop. That comparatively ice-free conditions might be attainable at Castleguard is suggested by the 6,000 year old flowstone on the south Benches; implying that the maximum altitude of soil and vegetation was much greater during the hypsithermal period.

A further point made by Palmer (1981a) is that initial conduit growth rates are relatively fast, but can be maintained only if water is available in quantities matching the tube capacity (e.g.in conventional karst landscapes, tubes growing beneath rivers are more likely to reach cave proportions). The regelation film has a limited volume of water available to maintain fracture discharge.

Furthermore, when a microfracture is enlarged beyond a certain limit, a pressure drop will be experienced at the entry point on the glacier bed. Rather than being a stimulative development, this will induce local freezing, concentrating solutes, and limiting continued hydrological activity and growth. Beneath glaciers, it is perhaps the regelation film waters which are responsible for cave initiation, while it is those few proto-caves encountered under suitable conditions by subglacial conduits which develop into subglacial shafts.

A remarkable feature of the five glacier ice plugs presently known in the cave, is the complete absence of finely ground detritus. This is reflected in the clarity of the spring waters, in which most of the turbidity is probably extraglacial in origin. Engelhardt (1978) has suggested that impermeable subsole drift is important in controlling water movement at the glacier bed, and Drake (1983) and Ford (1983a) have emphasised the importance of till in "shielding" karst surfaces from solvent activity. At Castleguard, the absence of sediment partly reflects an inherrent resistance of the local carbonate bedrock to glacial abrasion; most erosion is by "plucking". The high turbidity of the Castleguard River, however, demonstrates that the local bedrock can be glacially abraded, and that it is position within the glacier system which is possibly more important in controlling till generation.

7.3. EROSION RATES IN GLACIERISED KARST

No specific measurements were made to determine denudation rates at Castleguard. However, the low dissolved and suspended load of the spring water suggest remarkably low erosion rates in the karst catchment. Table 7.1 shows extreme estimates of net erosion rates, using crude guesses of maxima and minima for relevant variables. The rates are indeed remarkably low, (possibly due to incorrect extrapolation outside the period of field observation), and are at least three orders of magnitude less than published erosion rates for glacierised catchments (Embleton and King 1975), largely because of the extremely low suspended sediment loading.

Interpreting these data is difficult, because the flux of water-borne sediment from a glacier is not necessarily an estimate of total sediment generation, but depends also on the geography and migration rate of the subglacial conduit network transporting the debris (Collins 1979). Avulsion in subglacial drainage systems is often associated with massive sediment discharges (J.Shaw pers.comm.) suggesting that the subsole drift is mobile and may cause temporary channel closure at the glacier bed. Recognising the extensive bed area efficiently scoured by basal channels over a long period of time, the low erosion rates reported here probably reflect the relatively direct passage of supraglacial meltwater into bedrock, preventing any interaction with sediments generated at the glacier bed.

7.4. EFECTS OF SUBGLACIAL KARST ON GLACIERS

Hallet (1976b) demonstrated that the presence of solutes in basal regelation water would result in a significant retardation of basal sliding. The relatively high solubility of carbonate rocks mean that this effect is maximised in karst areas. Walder and Hallet (1979) identified active subglacial karst, but made no comment on its possible impact on the glacier. The effects of subglacial drainage may be conceived at both the conduit and the film scales. Lost conduit waters no longer release their frictional heat at the glacier bed, nor do they remove basal sediments. Leakage of the regelation water, may decrease basal water pressures, and the exchange of water between conduits and the film is less likely. Voluminous water storage at the glacier bed is not possible over a freely draining karst. Recent terrace gravels three metres above the normal level of the Castleguard River are therefore unlikely to represent jokulhlaups; rupture of avalanche debris dams is more probable.

Atkinson et al.(1983) and Ford et al.(1976) suggested that air temperatures in Castleguard Cave were lower than expected, because of a reduction in the geothermal heat flux to the glacier bed caused by heat abstraction by karst groundwater. The 2.2 degree Celsius temperature of the Big Spring was cited as evidence for this effect. However, these authors considered geothermal (conductive) heat flux as the only energy component affecting the water temperature. In fact, the temperature of the water emerging from the spring is the result of several factors: the temperature of the original water, sensible heat transfer from air, mixing with other waters, ground (geothermal) heat flux, radiative exchange, latent heat flux, and viscous dissipation. The underground flow route means that there is no direct solar heating, and probably little sensible heat transfer from the air, or net latent heat loss to evaporation. If the influent water is meltwater, an initial temperature of zero degrees Celsius may be assumed (although Griselin (1981) reported water temperatures of 1.5 degrees in a subglacial river). The kinetic energy at the spring is negligible, although producing a spectacular resurgence. If the altitude of influx is known, then the temperature increase which should result from viscous dissipation during descent to the spring can be calculated from:

$$Tout = Tin + ((Hin - Hout)\ d\ g\ /\ C)$$

where: Tout is temperature of output

Tin is temperature of input

Hin is elevation of input

Hout is the elevation of the output

d is the density of water

g is gravitational acceleration

C is the specific heat of water

The difference between this calculated value and the observed temperature of the spring is from geothermal sources, or mixing with different water, providing the assumptions made above are met. Observations of the behaviour of the spring temperature were made to assess the validity of the above discussion, and to estimate the actual geothermal heat scavenged by the subglacial groundwater flow.

Unfortunately, the temperature data were not as continuous as was hoped, but did show slight diurnal fluctuation of from 1.8 to 1.95 degrees Celsius with daily minima lagged by 3-4 hours after discharge peaks. In 1980, the water warmed to 2.9 degrees just prior to cessation of flow at the Big Spring.

If inflow elevation of 2,500 m is assumed for zero degree water, a temperature of 1.8 degrees Celsius is expected at the Big Spring, corresponding to measured daily minima. The diurnal periods of warmer water and the warming during recession can be interpreted in terms of either residence time (geothermal heat) or mixing with warmer, extraglacial water. In general, the conduit nature of the aquifer will make it a poor sink for geothermal heat, favouring the latter hypothesis.

Drake (1983 pers.comm.) developed a numerical model to test the assertions of Atkinson et al.and Ford et al. Initially, it was hypothesised that the isothermal glacier bed was affecting the ground temperature, but under these conditions, model cave temperatures remained higher than observed levels. Models including a downward flux of meltwater were successful in matching the observed data, but extension of the model to consider the three dimensional situation of the cave with reference to the Castleguard Valley was equally effective. The presently available data are insufficient to demonstrate a significant effect of subglacial waterflow on the ground thermal regime of temperate glaciers.

The corollary of karst drainage of high level glacier ice, is the possible discharge of karst aquifers into the bed of valley glaciers. In some respects, the reverse effects will occur, because the emergent water will constitute a point source of heat, basal pressure and a locus of fluvial erosion. It may be hypothesised that an unusually well defined basal conduit may develop in this situation, although it may close in winter conditions.

Some distance above the valley floor in the neoglacial benches bordering the South Castleguard Glacier, cave networks have developed at shallow depth, paralleling the valley wall. These often contain coarse glacial debris. Ford et al.(1983) suggested that proto-conduits emerged at this level at an early stage in the growth of Castleguard Cave. These fossil remnants have subsequently been exploited as karstic marginal stream channels during past periods of glacial advance.

7.5 GLACIATION AND THE CASTLEGUARD KARST AQUIFER

Castleguard has been widely quoted as a type example of a montane karst affected by glaciation (Ford 1971a, Ford 1983a, C.C.Smart 1983), and there is little purpose in re-iterating these discussions, except where further comment is considered necessary.

The recharge points of the aquifer, observed both at the surface and underground, are remarkably immature. This in part reflects the rapid evolution of the glacier ice surface, compared to the time-scale of the evolution of bedrock landscape. Supraglacial streams at the margin of a glacier will migrate as the glacier front shifts over time, preventing sinkhole evolution. Those subglacial shafts fed via crevasses may well become moribund as crevasses change location. As the bedrock step north of Castleguard Mountain is gradually eroded southwards, the associated crevasses will gradually shift, and new shafts will be cut further down the cave. The large moulin reported overlying the headward complex may constitute an exception to this observation, if it is able to maintain an adequate catchment in the glacier ice. The best developed shaft system would then be associated with this feature.

The major factor accounting for the continued immaturity of the sinks and springs is the limited solutional potential of the waters. It may be speculated that the well developed Castleguard I and II are therefore predominantly the product of interglacial solution.

The immaturity of Castleguard II, or alternatively, the blockage and destruction of the original outlet from Castleguard II, has been generally accepted as an inheritance from past glaciation. Here an alternative hypothesis is developed, broadly based on the present findings.

The structural dip at Castleguard is slightly greater than the net gradient of associated cave passages. The cave is therefore characterised by long dip passages interspersed with short comparatively steeply rising "lifting chimneys" (Ford and Ewers 1978). Figure 7.3 illustrates this form and also shows the approximate distribution of sediments throughout Castleguard I. Coarse sediments are commonly associated with the low points and lifting components of the cave. Schroeder and Ford (1983) have explained the cobbles at the 8 m shaft close to the entrance of the cave as a coarse, autochthonous residue, too massive to be lifted up the shaft by flood waters, but nevertheless well rounded by water agitation. A similar interpretation can be placed on other coarse deposits on Figure 7.3, except that at the base of the 24 m shaft they almost occlude the passage, and at the other two points the passage is completely blocked by sediment. The extensive fine sediments in the cave are rhythmic silt-clay laminates of about 1 cm thickness (though individual laminae reach 15 cm) and of consistent particle size distribution at all locations (Schroeder and Ford 1983). They occur throughout the cave, except for the one kilometre of passage from the entrance to the 24 m shaft. At least three phases of silt deposition have occurred, with sufficient time between events to allow induration of the sediment surface. Each depositional sequence was immediately preceded by a brief phase of erosion and deposition of coarse autochthonous gravel. The silts occupy vadose invasion trenches and so post-date the development of Castleguard II.

Schroeder and Ford identify the deposits as glacial flour injected into the cave from either end during glacial maxima. This hypothesis is unsatisfactory for several reasons. At glacial maxima, relatively little melt water would be generated above the headward complex, and it is also unlikely that crevasses or moulins would penetrate to bedrock here either. Furthermore, present day water from this source is remarkable for its lack of turbididty and would be unlikely to deposit the fairly massive laminae seen in the sediments. Injection from the valley to the top end of the cave would require ice at least 350 m higher than the cave entrance, ie. over 600 m of ice in the Castleguard Valley. Such injection would also include the entrance passage. The heterogeneous nature of fluvioglacial sediments is not seen in the cave deposits; they are characteristically distal deposits, remarkably undifferentiated throughout the length of the cave. The absence of a coarser facies, and any longitudinal fractionation argues against an external injection hypothesis.

An alternative interpretation rests on the observation that Castleguard II was active before the fine sediments were deposited. During interglacial conditions the major inlets to the cave may become proglacial, and streams might then carry a heavy, heterogeneous sediment load into the upper end of the cave. Transportation of this material within Castleguard II would be most critical at the base of any lifting chimney. The fracture zone marking the Castleguard Valley is a likely location for such a shaft, which would gradually be occluded by material carried down dip, but unable to be carried up the conduit (Figure 7.4a). Atkinson et al.(1983) have demonstrated the sensitivity of the hydraulic friction factor to local constrictions in conduits. As a result, the hydraulic grade line would rise in the upstream parts of the cave. In a single passage, this would create a "pressure conduit", but in this case water would rise up through the vadose invasion shafts and into Castleguard I. The relatively slow upward flow would also fractionate the sediments, preventing the coarse fraction from reaching the cave (Figure 7.4b). If the hydraulic grade line should rise above the top of the 24 m shaft, water would rapidly overflow to the cave entrance and Forest Spring (Fig.7.4c), without depositing further sediments in the entrance passages. The laminates are also found in "The Next Scene", an ancient passage which lies some 10 to 20 m above the main conduit in the central cave near the Grottoes (Fig.7.3). This demonstrates that the local piezometric surface was well above the main cave passage during the era of sediment deposition. The crucial point is to recognise that the controlling conditions were hydrodynamic, even though the water in Castleguard I was essentially static. Attempting to maintain hydrostatic conditions for sediment deposition leads to unrealistic head distributions.

The above hypothesis is highly speculative, and substantially more detailed information is needed on the cave sediments, and their relation to vadose invasion shafts. However, integration of Castleguard I and II into a conduit hierarchy brings the discussion back to the present. Contemporary sediment transport in the karst is negligible, but the association of turbid events with flow recession may represent similar processes occurring within the active aquifer. So far it has been assumed that it was "glacial deposition and collapse" which obstructed the outlet of Castleguard II, leading to the present day spring hierarchy. The internal constriction mechanism described above is more likely to have been interglacial, and subsequent glacial deposits have simply served to obscure the valley floor.

It is now a minor step to extend these ideas to the Castleguard III system, linking the Saskatchewan Glacier and the Upper Meadows to the Castleguard Valley. The marginal sinks in the Saskatchewan Glacier today convey fluvioglacial material into the karst, much as the sinks on the upper Meadows would have done when the glacier surface was higher in elevation. If these conduits also became blocked with coarse debris, overflow routes would develop and water could be stored in the porous medium filling the passage, providing long term storage able to sustain perennial flow. The Red Spring reservoir was shown to be relatively conservative, with an estimated volume of 0.20 million cubic metres (Section 4.3.3 (1)). If this were a single conduit stretching the 7 km from the Saskatchewan Glacier to the Red Spring, it would have an average radius of 3 m.[a] Such a "mature" route would presumably be very ancient, and could provide the speleogenetic "target" for the early development of Castleguard Cave hypothesised by Ford et al.(1983). This feature would have a very damped rainfall response, and could also retain dye, accounting for the poor recovery of the Meadows trace (Section 4.3.3 (3)). The long residence time of water in a choked conduit could allow comparatively hard water to evolve, and mixing of this water with dilute flood waters in a common outlet passage could generate the remarkable chemograph of the Red Spring (Figures 4.21, 4.22).

Small paragenetic passages developed above the sediments may account for the limited, but rapid dye and rainfall response. However, the cave floods (and associated rainfall response) are highly dynamic, and clearly represent a relatively open flow route, independent of any blocked conduit. The underflow leading to the valley aquifer may be either a partially blocked passage or an immature conduit. Evidence suggests the latter feature, because not only was dye transmitted rapidly and discretely to the Big Spring (Figure 4.20), but there was a close association between cave floods and Big Spring turbidity (Figure 4.16). Figure 7.5 is an attempt to summarise the nature of the Meadows Karst system; it is greatly stylised, emphasis having been placed on hydrological function more than on morphological reality. The underflow marked "A" connects to the same point on Figure 5.17, the model of the Valley Aquifer. Figure 4.27 provides more specific details concerning the hydological system.

Ford (1979) described a fining upwards fill sequence from Nakimu Cave (Glacier National Park, B.C.), which he attributed to advance of a glacier over the cave. In general, however, cave fill events need bear no absolute relation to a glacial chronology, depending largely upon the volume of clastic load entering the system. It is the geographical relationship between the karst aquifer and the glacier system which is critical in determining the interaction between them. At Castleguard, the catchment geomorphology and hydrology are such that sedimentation is more likely during an interglacial climatic optimum or glacial advance or retreat, rather than in times of more complete ice cover.

7.6. ORIGIN AND DEVELOPMENT OF THE CASTLEGUARD KARST

A transection valley like Castleguard Meadows defies simple explanation in either a fluvial or glacial landscape. Here, it is postulated that during an ice free era, the Meadows Valley extended southwards from Mt. Andromeda (Fig.7.6a). What is now the upper Saskatchewan Glacier constituted a western tributary valley (Figure 7.6a). Erosional breaching of the Stephen Formation at the site of the present Saskatchewan Glacier provided access to the Cathedral Formation, and a karst system developed beneath the Meadows, exiting through fractures on a similar breach in the Castleguard Valley. Although the Meadows link is today called Castleguard III, this is a rejuvenated and recent feature and the earliest element is best termed "Castleguard 0". To the northwest of Castleguard Mountain a valley was then cutting into the Cathedral Formation, developing Castleguard I which was directed towards the speleogenetic target of Castleguard 0 (Figure 7.6a). At some stage in this sequence glaciation started to disrupt the landscape. The preferential glacial development of valleys with a north easterly aspect led to the accelerated deepening of both the upper and lower Saskatchewan Cirques compared to the Andromeda cirque, leading to a breaching of the Meadows Valley by the Saskatchewan Glacier. The present day location of this hypothetical event has diagnostic massive buttresses on either side of the Valley (Figure 7.6b). Castleguard 0 was truncated and blocked by glacial debris, but Castleguard I continued to evolve onto the lowest downdip exit point as described by Ford et al.(1983). Subsequently, as the Castleguard Valley was deepened, Castleguard II developed, graded to the new valley floor.

There are three obvious questions regarding this sequence: (i) Why did Castleguard 0 develop north-south and not downdip as did Castleguard I and II? (ii) Why is there no evidence for the Castleguard 0 "target" in the present Cave? (iii) Why is there no Castleguard "1.5" known at an intermediate level between I and II?

(i) The Stephen Formation dips down the Castleguard Valley, and the Cathedral Formation would have been exposed earlier up-valley providing a potential discharge point slightly off dip with respect to the sinks (Figure 7.6a). In addition, the Meadows provided a route with minimum overburden compared to the downdip route beneath Terrace Mountain. The resulting slight expansion of fractures along the valley may have provided a preferred axis for speleogenesis. Certain early passages in the headward complex of the known cave have also developed obliquely to the dip. It may be that overburden is a speleogenetic control in the region and the initial elements of Castleguard I skirted around the Castleguard Mountain massif.

(ii) The original junction of Castleguard 0 and I may have been removed by enlargement of the Castleguard Valley. However, the exit of Castleguard 0 should still be found in the valley side, but may not be obvious because it is choked with debris and may be a lifting chimney. A possible site is one of the Forest Spring orifices which is alluviated and rises through allochthonous sediments.

(iii) Castleguard "1.5" may yet be discovered, but alternatively a sustained glacial epoch could have deepened the valley by 300 m, during which time the cave did not evolve significantly. Furthermore, the fractures of the Castleguard Valley would have been a favoured location for a phreatic lifting chimney, and the spring orifice may have lain at a much greater elevation than the lowest point in the Castleguard II conduit.

The present day active systems II and III represent a response to internal sedimentation and paragenetic evolution of alternative flow routes. The Col karst on the upper Meadows is a disorganised group of relic proglacial sinkholes feeding into the newly evolving conduit system of Castleguard III.

VARIABLE	(UNITS)	MINIMUM	MAXIMUM
CATCHMENT AREA	(km2)	30	128
SEDIMENT CONCENTRATION	(Kg/m3)	0.02	0.035
TOTAL ANNUAL DISCHARGE	(m3/a)	23,000,000	40,000,000
TOTAL ANNUAL LOAD	(Kg/a)	460,000	1,400,000
EROSION RATE	(Kg/m2/a)	0.0036	0.0467
EROSION RATE	(mm/Ka)	1.4	19.

NOTES: DENSITY=2500 Kg/m3

Table 7.1: Estimation of approximate erosion rates for the Valley Spring Catchment.

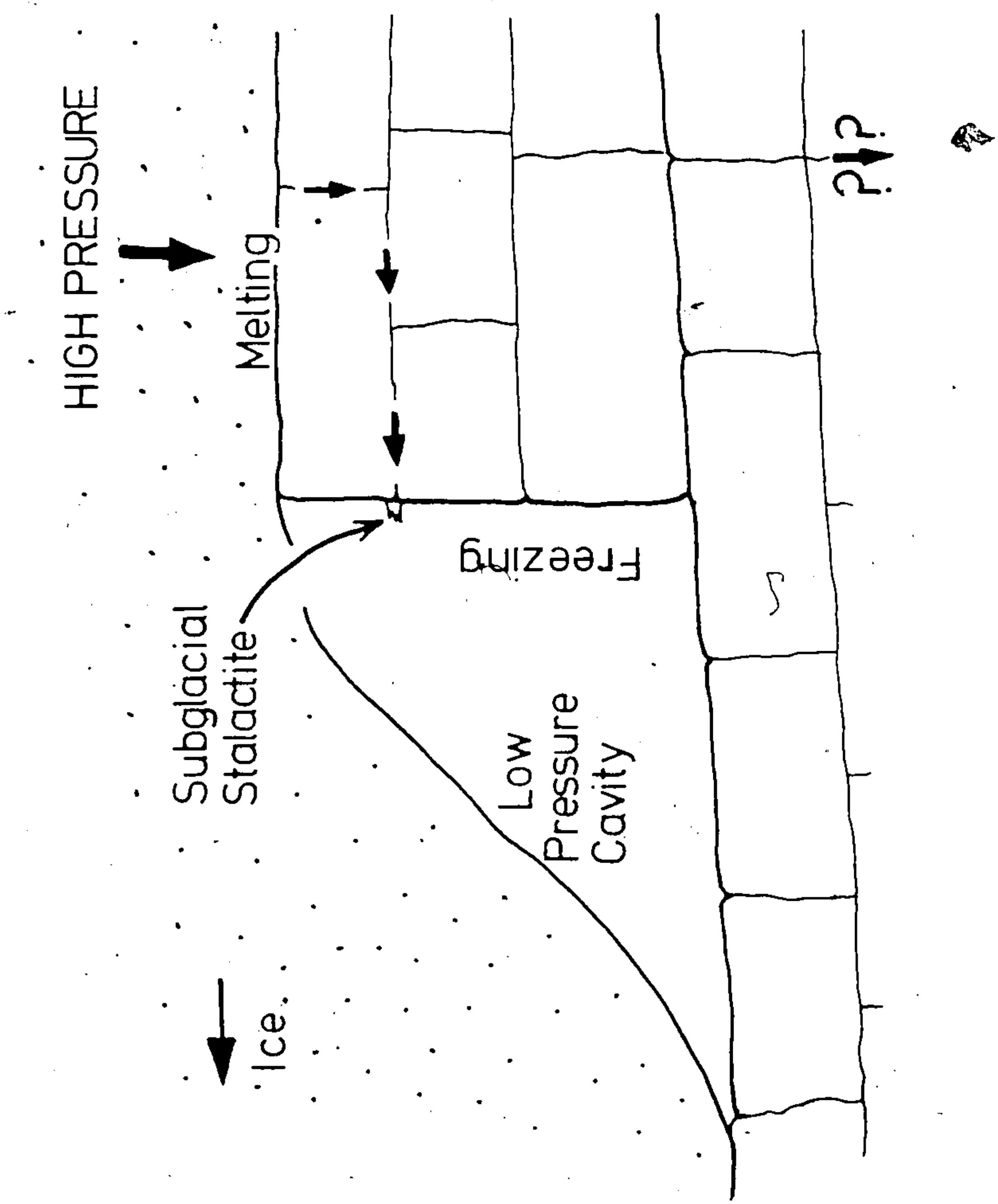


Figure 7.1. Formation of subglacial "stalactites"

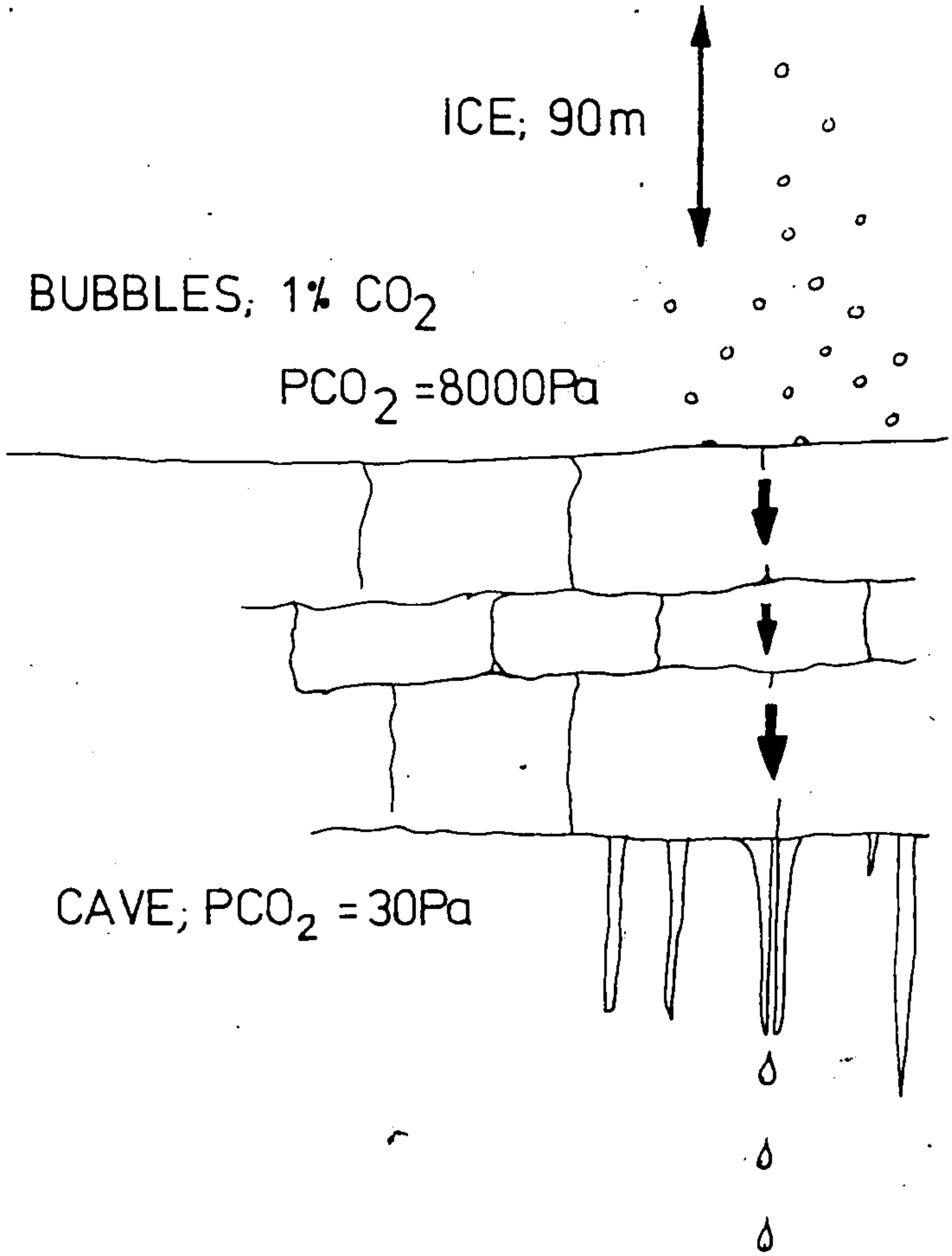


Figure 7.2. A model for subglacial speleothem growth by degassing of carbon dioxide

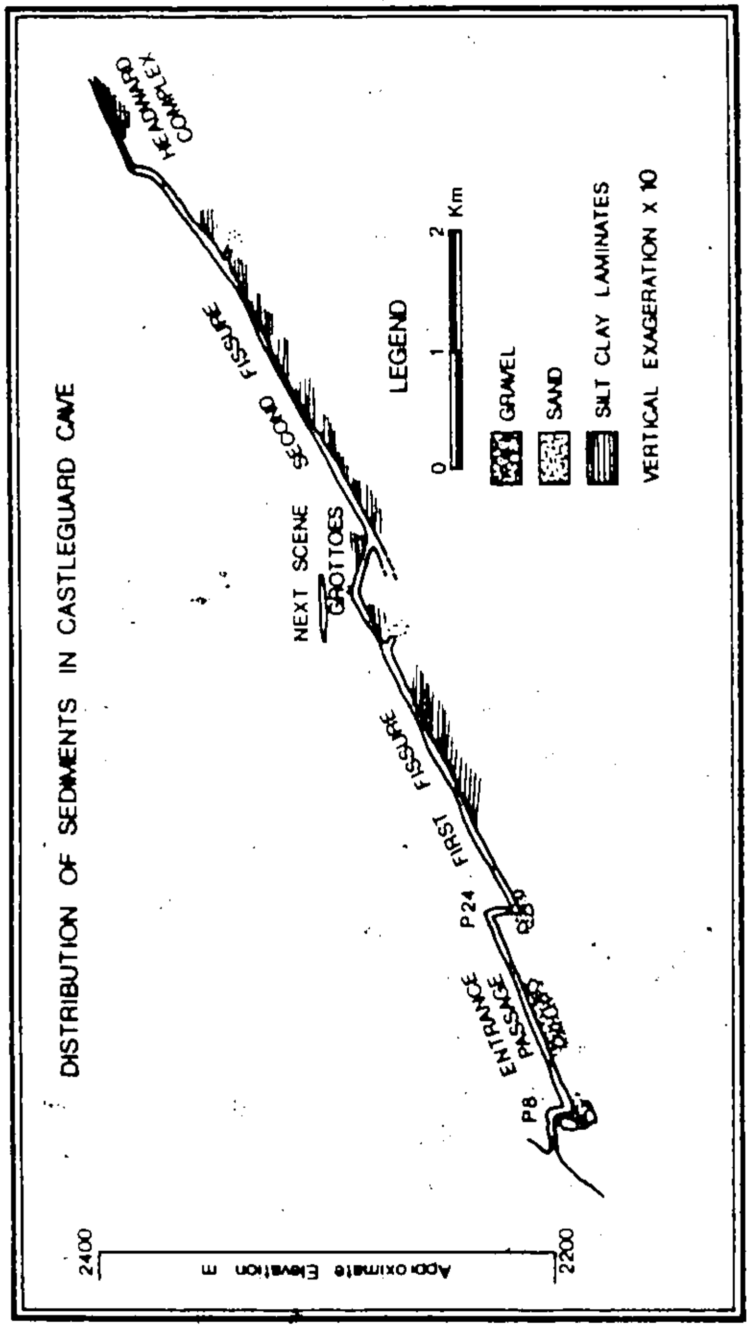


Figure 7.3. Distribution of sediments in Castleguard Cave (partly after Schroeder and Ford)

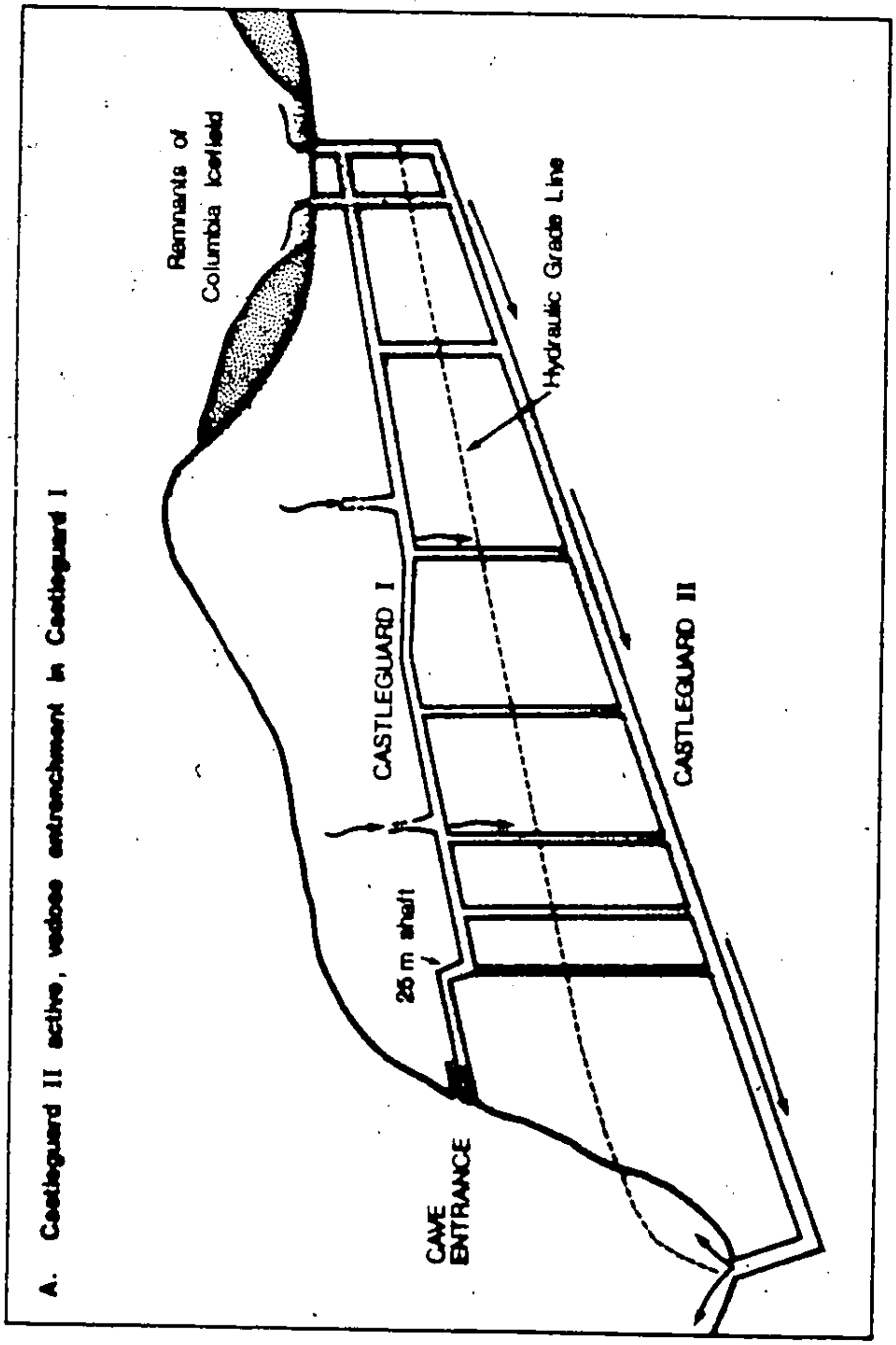


Figure 7.4a. Hypothetical model accounting for the silt-clay laminates distributed throughout much of Castleguard Cave

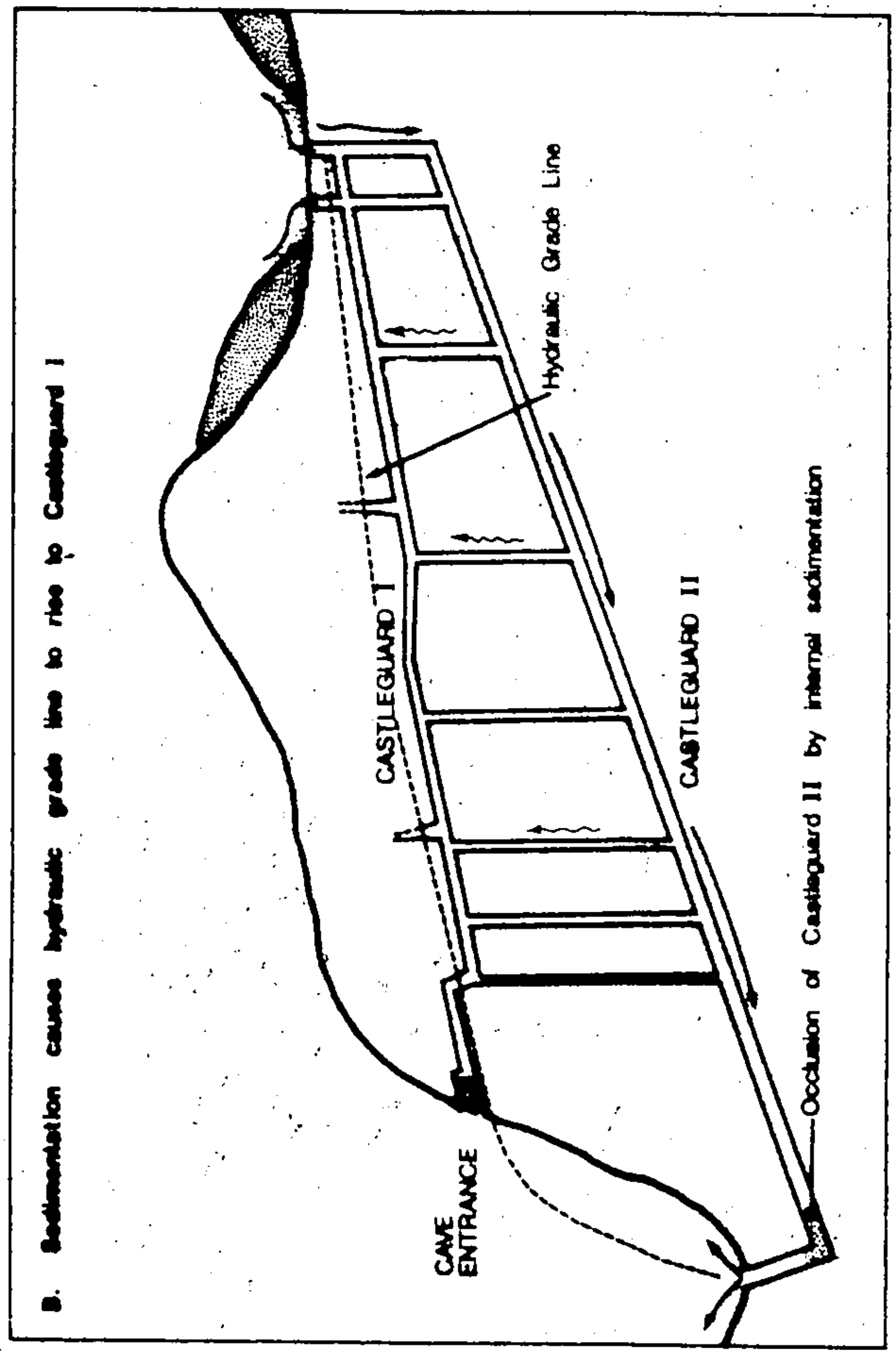


Figure 7.4b. Hypothetical model accounting for the silt-clay laminates distributed throughout much of Castleguard Cave

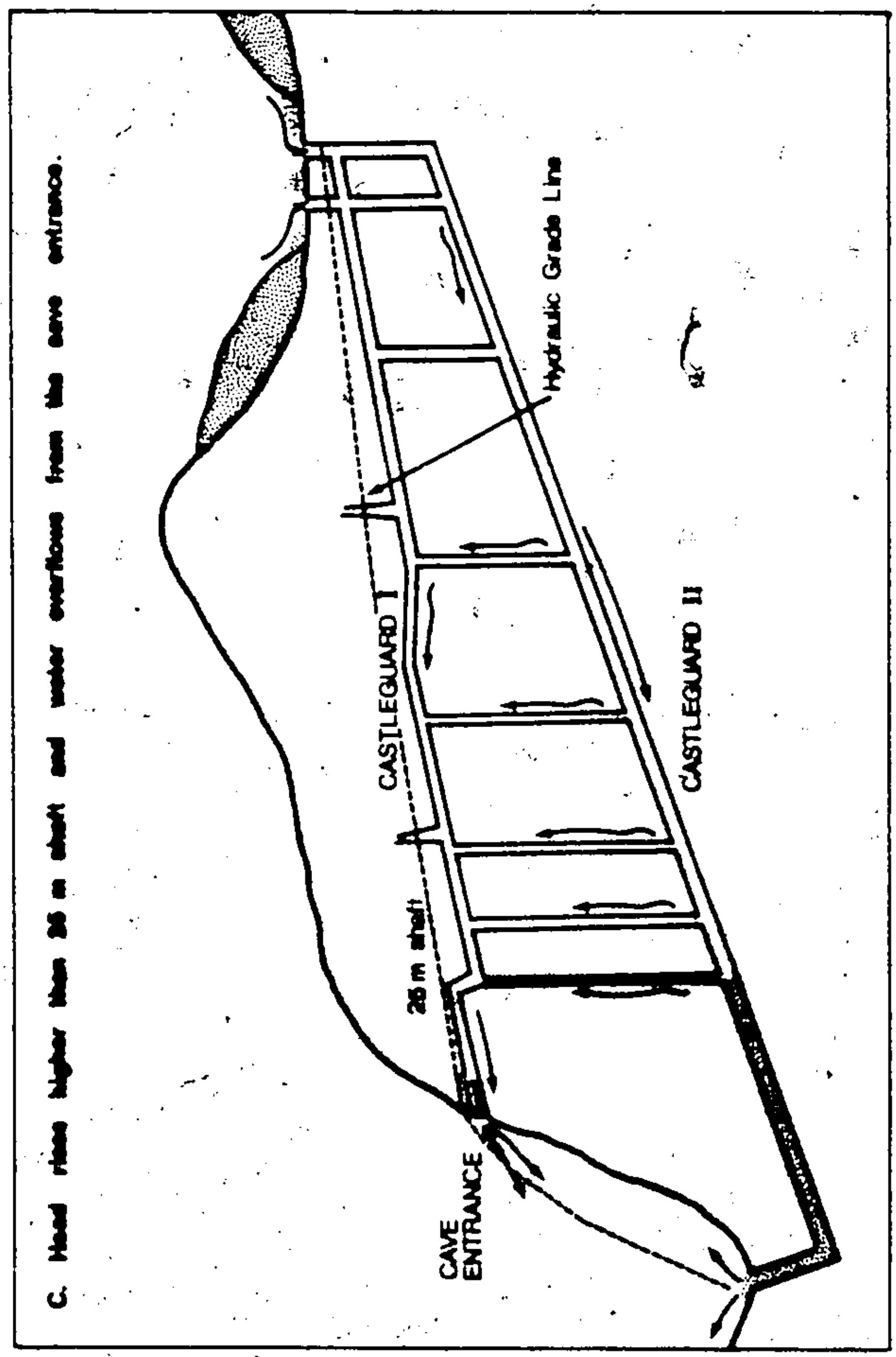


Figure 7.4c. Hypothetical model accounting for the silt-clay laminates distributed throughout much of Castleguard Cave

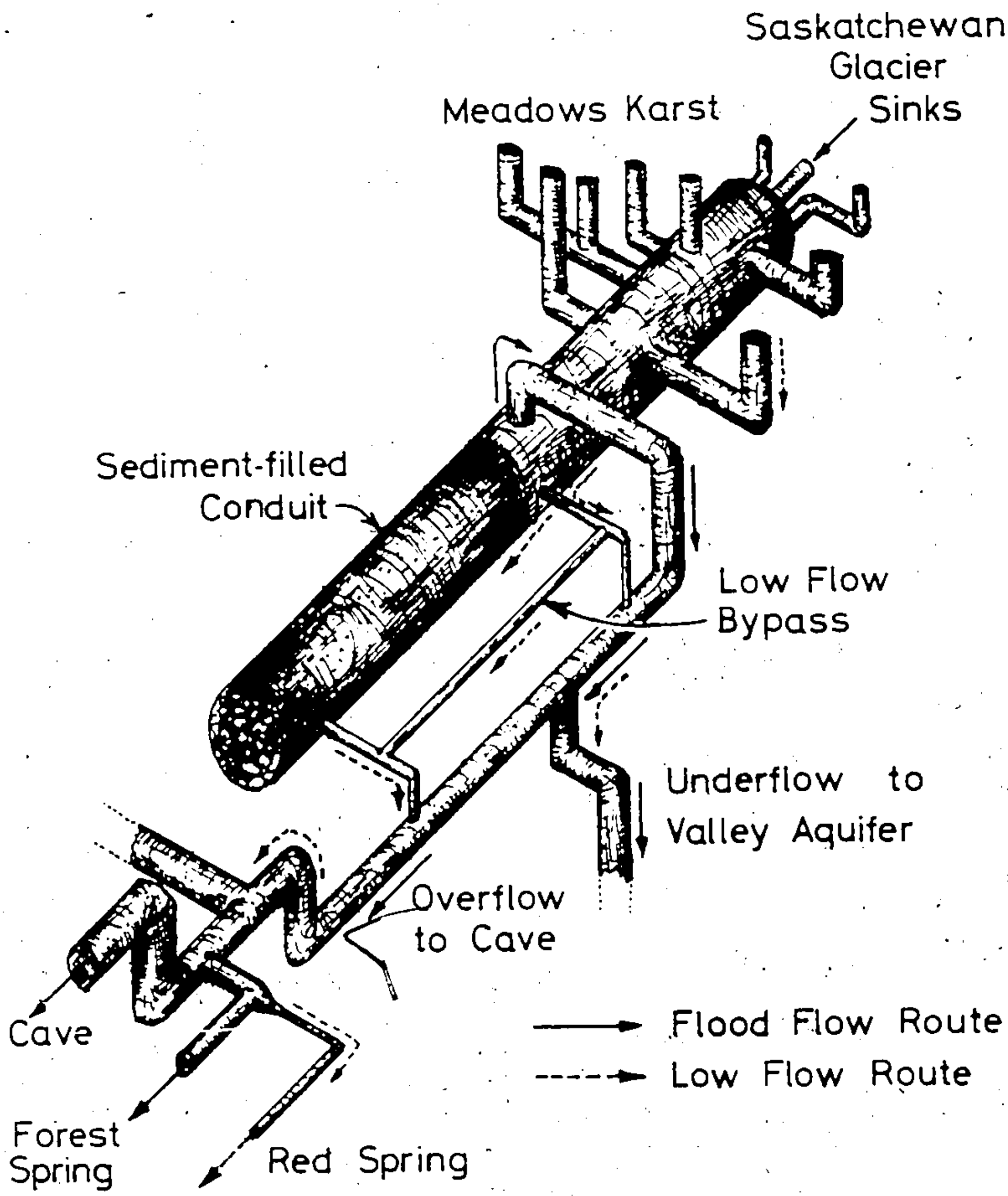


Figure 7.5. Tentative functional model of the Meadows Karst with internal clastic alluviation

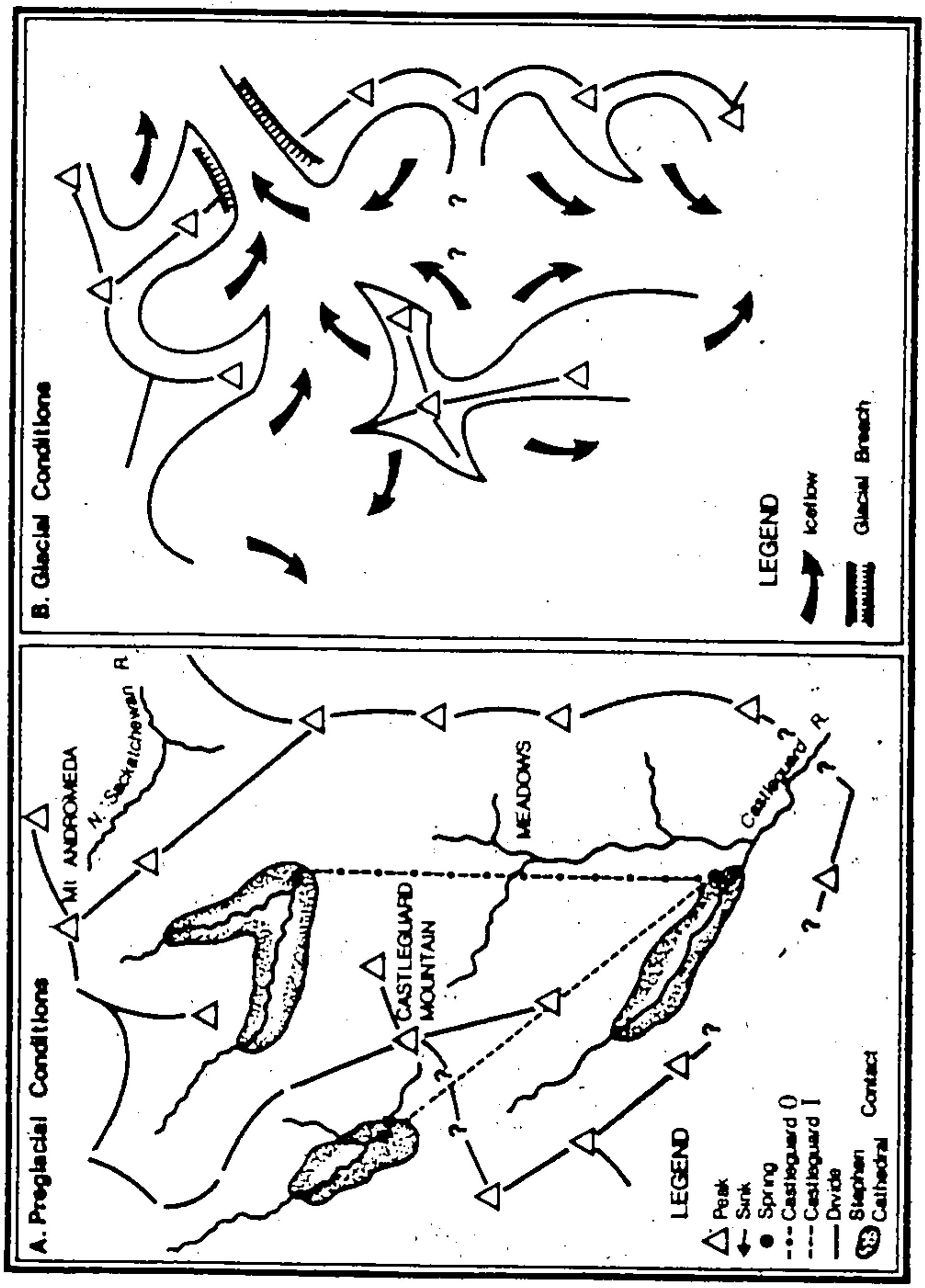


Figure 7.6. Hypothetical model for the evolution of the Castleguard Karst

CHAPTER EIGHT

CONCLUSIONS AND RECOMMENDATIONS

8.1 CONCLUSIONS

8.1.1. THE CASTLEGUARD KARST

1. The main aquifer of the karst lies beneath Castleguard Mountain and Valley, and is a conduit end-member of the "sponge"-conduit continuum of styles of karst aquifer. It is believed to consist of a trunk conduit which integrates scattered recharge shafts, and discharges through numerous constricted outlets.

2. The catchment area of the Valley Aquifer extends well into the central Columbia Icefield which supplies surface meltwaters to the aquifer during the ablation season. Small cirque glaciers around Castleguard Mountain drain into the karst subglacially and proglacially. The upper Saskatchewan Glacier drains into the Meadows karst, but is lost to the valley aquifer except when recharge exceeds a certain limit.

3. There is a semi-autonomous aquifer lying beneath the

Castleguard Meadows and draining largely through the cave springs hanging 250 m above the floor of the Castleguard Valley. It is a "hybrid" aquifer; under low summer flow, discharge is sustained by diffuse groundwater flow, with a minor conduit component. During days of great ablation the springs discharge massive floods draining through a mature, otherwise inactive conduit.

4. The catchment for the Meadows Aquifer depends on discharge conditions. Under low flow small spring-fed streams (plus a small ablation and precipitation component) recharge the porous medium and the minor conduit aquifer. Much of the Meadows recharge is lost through a constricted underflow conduit to the valley aquifer. However, when ablation rates are high, recharge from the Saskatchewan Glacier, or the Meadows can exceed the capacity of the underflow and floods are routed through a well-developed conduit to the cave springs.

5. During flood events, limited capacity of the lower cave springs causes water to build up in the cave, where it may emerge as "Cave Floods". These events reach up to 5 cubic metres per second and are strongly diurnal, reaching peak flow around midnight and minimum flow in the early afternoon. Rainstorms occurring during periods of cave flooding supplement recharge and may induce cave floods during expected minima. The largest floods were observed

when intense melting of the seasonal snowpack in the upper Meadows took place.

6. An independent karst aquifer also exists beneath the Terrace Mountain Range in the Eldon-Pika Formation. Little is known of this aquifer except that it exhibits conduit flow, an overflow-underflow hierarchy, and has longer groundwater residence times than the valley aquifer.

7. The hydrological regime of the region is dominated by ablation, especially when the central Icefield is active. Rain events are comparatively brief and volumetrically insignificant, although when superimposed on diurnal maxima, rainstorms can induce very high, short term discharges.

8. Combination of the massive lithology and the conduit aquifer creates conditions closely akin to an urban storm sewer system. In common with such systems, the present aquifer is more successful in draining its catchment than in providing a long-term storage medium. Consequently, the aquifer is remarkably ineffective as a natural reservoir; providing little moderation of the intensely seasonal runoff regime.

9. The evolution of the Castleguard karst has been complicated by glaciation. A sequence of development may be hypothesised:

(i) "Castleguard 0" developed in a fluvial landscape along the line of the present Meadows. This provided a speleogenetic target for Castleguard I which developed from streams sinking to the northwest of Castleguard Mountain.

(ii) Glaciation disrupted cave development; Castleguard 0 became blocked, Castleguard Valley was deeply incised.

(iii) Castleguard II developed beneath Castleguard I, graded to the valley floor. Castleguard III started to evolve along the lines of Castleguard 0.

(iv) Internal sedimentation (plus spring occlusion or conduit collapse) restricted flow in Castleguard II leading to (paragenetic) development of the numerous present day springs, and the extensive silt-clay laminates seen in Castleguard I. Castleguard "2.5", the Meadows underflow developed, linking the two independent aquifers. (v) Although the valley aquifer is at present hydrologically very active, solution rates are very low, retarding the rate of growth and development of the system. However, the high-level abandoned springs suggest that the system may be gradually increasing in capacity as lower conduits are enlarged. The Artesian Spring is at present the lowest outlet and has the most sustained flow and will become the dominant spring providing glaciation does not disrupt the aquifer once more.

8.1.2 GLACIERS AND KARST

1. Glacier-to-groundwater discharge occurs at both a conduit and diffuse scale, matching the glacial regelation water to percolation water, and surface meltwater with karst conduit water. The modest aggressiveness of bulk meltwater limits the effective penetration distance of subglacial conduit waters, especially in vadose shafts. The chemistry of subglacial percolation water is complex, but may allow rapid development of micro-fissures at the glacier bed.

2. The general interaction between karst and glaciation is complex, but as Ford (1983a, p154) says, "destructive and inhibitive effects predominate .... Nevertheless, ... general prediction of the impact of past glaciers upon karst in a given area is not feasible without detailed site studies." The present work differs from Ford in a few points, however.

(i) Glacier ice superimposed above a karst aquifer does not necessarily increase overall hydraulic head. Subglacial conditions at Castleguard appear to be vadose with reference to conduit waters. Regelation water, on the other hand is at high pressure, and may initiate extensive, but small scale karstification. Two processes limit substantial development: (a) the regelation film has insufficient local volume to develop larger conduits, and (b) development of an open route from the glacier bed will lower local basal pressures causing freezing and pressure isolation of the entry point.

(ii) The association between the glacial cycle and processes acting on the karst is not necessarily straightforward and depends intimately upon the geography of the karst in relation to the zonation of the given ice body. The configuration at Castleguard would place the main inlets to the valley aquifer beneath the dry snow (accumulation) zone during a peak of glaciation. Under present circumstances the inlets are active, but carry little or no sediment. More characteristically, it is proglacial streams which carry a heavy sediment load. Thus the maximum rates of karstification are associated with interglacials, provided the groundwater system remains active. Although continuing to function hydrologically beneath ice cover, the karst is little developed, and full glaciation would completely prevent any significant karst evolution. This tentative cycle is presented in Table 8.1.

8.1.3. KARST HYDROLOGY

1. The present work has had some success in resolving the

internal structure of a karst aquifer. This has been helped by the nature of the Castleguard system: a discrete conduit network responding to regular daily pulses of discharge. This allowed clearer identification of spring behaviour than would have been possible under a conventional rainfall-runoff regime. The other principal factor is the data density. Very few previous studies have employed an hourly sampling period, which permitted such ready recognition of the nonlinearity of spring behaviour. The rapid flushing of dye through the system also allowed a remarkable number of traces in such a short field season. Once the voluminous field data were organised, plotted and inspected, the behaviour of the karst aquifer was so overtly non-linear that a complete conceptual re-evaluation was unavoidable.

2. The conduit aquifer has certain distinctive and diagnostic properties. The relationship between recharge and discharge, and between flow at various springs is rather non-linear. Contrasts in spring behaviour reflect the geometry of the associated conduit and conduit network. Spring discharge is strongly affected by conduit radius, and is comparatively insensitive to head. However, under conditions of varying flow, lower-lying springs exhibit less variation than a similar higher level spring. Short-term discharge behaviour reflects spring position

within a vertical "overflow-underflow" hierarchy. As a result of these observations, considerable care has to be observed in interpreting spring hydrographs and in applying both pulse-train analysis and recession curve analysis to conduit aquifers. These techniques are by no means redundant, but their application demands careful measurements and interpretation is a little less straightforward. Collective gauging of spring groups may obscure the important diagnostic properties of the conduit aquifer. In this respect, direct analogue records are more informative than coarser digitised data.

3. Presently available fluorometric techniques allow very high resolution, field based water tracing at moderate cost. Most karst water tracing is concerned with point-to-point applications for which these methods are excessively expensive, complex and time consuming. However, where information concerning the nature of the flow route is required, considerable insight can be gained. Unfortunately, analytical fluorometry has advanced faster than karst hydrology, and an imbalance has resulted between the technological and conceptual levels in water tracing.

4. A procedure for the qualitative interpretation of tracer breakthrough curves has been developed. A high level of understanding of the tracer, sampling and analytical procedures, and of the field site are necessary

before any interpretation is made. Processes affecting tracer behaviour depend on all of these factors, and alternatives should be explicitly rejected before a particular interpretation is adopted. A broad appreciation of the experimental environment is therefore necessary, demanding a systematic approach to the problem. Such a methodology has been developed here, and demands consideration of:

(i) The probable structure of the flow route.
(ii) Effects associated with the tracer.
(iii) Effects resulting from the sampling regime.
(iv) Network effects.
(v) Hydraulic effects.
(vii) Interaction with storage media.

5. The "dimensionless" recovery diagram plots the dye recovery from each spring over the proportion of total discharge flowing from that spring against total system discharge. In studies where more than one spring is positive, this figure provides a clear statement of the relationship between springs.

8.2. RETROSPECTIVE AND RECOMMENDATIONS

1. Castleguard proved a stimulating and unusual field location for the reasons outlined above. However, the

discussion throughout this text has failed to emphasise the extreme complexity of the aquifer, in both the spatial distribution of inlets and outlets, and the rapid oscillations of discharge. All significant inlets are inaccessible, and most outlets difficult to instrument, and unsteady flow probably predominates in much of the aquifer. An expansion of the instrumental network in a subsequent study would undoubtedly increase understanding, but the marginal gain would probably not be justified, given the unavoidable indeterminacy of the system.

2. The high frequency of sampling for both discharge and tracer measurements remains inadequate. The dye sampling network was generally rather skeletal and the automatic samplers somewhat unreliable, and frequently undermined the considerable investment of money and labour necessary for each trace. A better balance of expenditure would have seen a superior sampling network. In situations where breakthrough curves change very rapidly, or where contamination is a problem, continuous-flow fluorometry is an excellent sampling procedure. At present, unfortunately, only one dye can be sampled for at one spring, nor does the method provide any opportunity for sample treatment, and care should be taken with tracers sensitive to temperature or pH changes. A commitment was made to replication of traces at Castleguard, at the cost

of ignoring 98% of the springs in the valley (although, perhaps only 10-40% of the total discharge was thus ignored). The variability of the results justified this decision, but wider sampling might have led to a more complex, and less absolute model of the valley aquifer.

3. The abandonment of the water chemistry program in 1980 was necessary to ensure that the expanded hydrometric and tracer program was properly executed. However, the carbonate and isotopic hydrochemistry of the valley aquifer remains a promising study, but the complexity of the system, and the peculiar difficulties of the environment would demand a thorough analytical approach. The Meadows karst aquifer, and the Terrace Mountain aquifer are both discrete entities, and perhaps more amenable to study than would be the entire system.

4. The present dissertation has only succeded in gaining a superficial understanding of the glacier-groundwater relations in a karst aquifer. The successful trace from the Saskatchewan Glacier to the Big Spring has opened up a new horizon in this respect. Quantitative tracing from the Columbia Icefield would provide a far clearer understanding of the hydrology. However, the necessary expansion of the sampling network would necessitate a massive increase in logistical support.

6. The emphasis in much of the present work has been on resolving the physical nature of the "black box" karst aquifer. An alternative approach might have attempted a stochastic evaluation of the system. The non-linearity and lack of stationarity of the present system and the data prevent the straightforward adoption of this approach. Although an improvement might be gained from measuring total system discharge, such techniques are probably best developed for "simpler" cases such as that of a single glacier.

6. The numerical model of a conduit aquifer developed here is both unrealistically simple, and conceptually naive. The expansion of the model to realistic situations, while still remaining faithful to physical proceses, is unlikely. The immediate application will be in developing interpretative methods for pulse train and recession analysis in conduit aquifers.

GLACIAL STAGE	HYDROLOGICAL PROCESSES	SOLUTIONAL PROCESSES	CLASTIC PROCESSES	KARST PROCESSES
Pre-glacial	Active	Active	Moderate (increasing)	Rapid
Early Glacial	Active	Limited	Inactive	Slow
Full Glacial	Inactive	Inactive	Locally Active	Moribund
Late Glacial	Active	Limited	Inactive	Slow
Post-glacial	Active/ Inhibited	Active	Very Active	Fast/ Inhibited

Table 8.1. Hypothetical status of geomorphic and hydrological processes in the Castleguard karst through a glacial cycle.

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APPENDIX A

INSTRUMENTATION AND TECHNIQUES

A.1. CHEMICAL AND ISOTOPIC ANALYSES

A.1.1. DEUTERIUM

Water samples were collected in 30ml glass bottles or 25ml plastic liquid scintillation phials. Bottles were dry before collection, but were rinsed if sufficient sample was available. Rainwater samples were collected from beneath vacuum pump oil in standard 5" rain gauges. Sample bottles were stored under water in the field to prevent evaporation.

Dr. R.Krouse, University of Calgary kindly gave access to to a mass spectrometer. Analysis was by direct influx of water vapour from a drop of sample on a hot plate. Vapour was reduced over heated uranium metal and entered directly into the mass spectrometer. A semi-automatic cycle of 15 seconds per determination allowed rapid replication. Analysis was repeated until stable values were obtained, in order to overcome memory effects. The final values are the average of five to nine determinations. Samples were alternated with a known standard. An empirical precision of better than plus or minus one per mil was obtained.

A.1.2. TRITIUM

Tritium samples were of about 120ml in two-valved, tightly sealed steel bottles, previously filled with dry nitrogen. Samples were taken after flushing with 50-100 volumes of water, either by gravity flow or siphoning. Samples were stored cool, with water on both sides of the sealing valves

Dr.B.Clarke of McMaster University generously provided analytical facilities and assistance. Analysis was indirect, inferring the tritium concentration from the quantity of Helium-3 accumulated in a degassed, frozen water sample over a known period of time (Clarke et al 1976).

A.1.3. CARBONATE WATER CHEMISTRY

Water samples of 250ml were collected in polyethylene bottles previously rinsed in sample water. All analyses were performed within two days of sampling. The methods employed were quite conventional colorimetric titration (e.g.Gascoyne 1975). Results are the average of at least two titrations, replication continuing until some consistency was obtained. Problems of precision were encountered in some water samples. This was inferred to stem from sub atmospheric partial pressures of carbon dioxide and the low dissolved load. Samples exposed to the air for more than a few hours were found to contain increased dissolved carbonate in proportion to exposure time. Presumably this additional carbonate came from colloidal and fine suspended load which were present in most samples. This also prevented clear end point determination in some cases. In ordinary circumstances, these effects are small, but in meltwaters low in dissolved load, they constituted a significant error.

Conductivity was measured with a YSI-33 Salinity Conductivity Temperature meter. Precision was low in the dilute waters and some interference from suspended sediment noted. A continuous conductivity record was obtained using a Rustrak Model 288 and an operational amplifier (courtesy of Mr.R.Bowen, McMaster University). The recording system was unfortunately difficult to calibrate, having marked nonlinearity. In addition, the power supply of the instrument was not self-adjusting, and this resulted in marked instrument drift over time. For this reason the conductivity data are expressed in arbitrary units.

A.2. METEOROLOGY

Temperature was measured in degrees Celsius on a Cassella Hygrothermograph with an 8-day clock, cross calibrated with a maximum-minimum thermometer. In 1979, instrument shelters were improvised from packing crates and boulders and gave consistent, noise-free data. In 1980, louvered screened boxes about 1m above ground level were used. These suffered from attacks by rodents and tended to act as snow traps in foul weather. At these times the clockwork drives were prone to failure. The periods when snow (and occasional hoar frost) lay on and around the bimetallic sensor are marked by prolonged isothermal periods in the temperature series.

Data were obtained by point estimation to the nearest 0.5 degree celsius. The resulting series was markedly stepped during periods of very gradual change in temperature. This was improved by smoothing with a single pass of a Hann filter. The undesired result of this smoothing was loss of power in some of the real, high frequency, high amplitude events.

Rainfall was recorded by A.E.S. standard tipping bucket raingauge (now obsolete). Every 0.01" of rainfall is recorded as an impulse on an event recorder. There were occasional problems with the solenoids on the event recorder, which were solved by strengthening the return spring and replacing the alkaline battery for the solenoid every six weeks. The mercoid sensor switch needed replacing on one machine. The record was calibrated using the total accumulated rainfall from a nearby 5" collecting raingauge. This, combined with immediate inspection of the chart provided the best check on machine malfunction, which was usually only intermittent and difficult to discern.

Data were obtained as hourly totals, corrected by period totals (5"gauge/tipping bucket), and converted to millimetres.

A.3. HYDROLOGY

Given the experimental design for the hydrological program, suitable gauging sites had to be found. Sites of low turbulence, stable section, and suitable stage-discharge relation were preferred. In mountain streams weirs are difficult to instal, and stable sites hard to find. Considerable effort was put into physically modifying streams to provide good control sections. This involved using boulders to stabilise the section, and constructing weirs and obstacles upstream to reduce the kinetic energy of flow through a section. On larger rivers this was not usually possible.

Stilling wells were dug in as deeply as possible and anchored with steel girders and cables, fixed with rock bolts or pitons. Boulders were used to armour the base of the well, and gravel to damp oscillations. A general concern was with spring floods and possible glacier floods (for which fresh gravels 6m above stream level provided some evidence on the Castleguard River). Although installation usually took a number of hours, observation of behaviour over longer periods often resulted in subsequent modification. Turbid streams often deposited sediment around the base of the well, damping its response. This was overcome by filling the well with water to flush away the blockage. An indication of recorder response was also gained in this manner.

Leopold and Stevens Type F and A71 and Ott XX stage recorders were used. The more versatile and light weight Type F was found the most suitable in remote areas, with the additional advantage of furnishing charts frequently for field digitisation and inspection. The corollary of this is, however, that recorders needed to be visited rather frequently, and precisely. The continuous strip chart recorders could conceivably run out between visits. Instruments were mixed imperial and metric, which limited interchangability.

Charts ran from 4-30 days. Whenever a recorder was visited, the following were recorded on the chart: depth to water in the well, time, date, location and comment.

Charts were digitised by hand. The type F charts were field digitised, the longer charts later. The transition from chart to chart was noted and accumulated to concatenate the chart data. The reading precision was about 0.01" and 0.2mm, or more or less depending on the need for precision. The relationship between the chart data and stage data (datum depth-depth to water) was inspected by chart and entire series, to check for smooth concatenation. Anomalies were ommitted, corrected or averaged depending on the character of the "error". Charts with an inexplicably evolving chart-stage relationship were corrected by a least squares fit. Time errors were corected either by digitising with a transparent mask with a correct time grid, by hand interpolation of highly varable data, or by linear interpolation where it was appropriate. The chart data were then converted to stage by:

$$S = A + B \quad C$$

Where: S=Stage in inches or mm

A= an empirical constant

B= Stage recorder gear ratio

C= Concatentated chart data in inches or mm

Both the chart and stage data were plotted and inspected for digitising and data entry errors.

Considerable effort was expended to obtain more than 90 discharge gaugings of the streams in the area. Of these 50 were by dilution gauging and 40 by current meter. The current meter was a standard Price AA or a Pygmy meter mounted on a wading rod, maintained as recommended by Smoot and Novak (1968). Measurements were made using the single 0.6 or the 0.8, 0.2 depth method (Church and Kellerhals 1970). The great mobility of the method meant that it was used at the remoter sites, or where several gaugings were necessary in sequence. The major problem with the method was imprecision in the highly turbulent streams. In addition, the counting clicks were often inaudible, and wading in the larger rivers was impossible even with complete wet suit and safety line. The flashy nature of the streams meant that significant changes in discharge often occurred during a single gauging, although this also meant that a single day on site could furnish a relatively complete rating curve.

Dilution gauging followed the slug injection technique described by Church and Kellerhals (1970). Dye was drawn into special dye pipettes, drained into a glass beaker from which the dye was injected and rinsed into mid-channel. The injector did not sample, but retired downstream to rinse the equipment. The cloud was inspected for quality of mixing. Sampling was at up to 15s intervals around peak concentration where error is most likely. The tail samples were up to 5 minutes apart, and up to 30 minutes after the first arrival time. Mixing lengths were often remarkably short in cascading reaches, but intolerably long in larger rivers such as the lower Castleguard. Two or three background samples were collected before injection, because tracer tests were sometimes going on at the same time.

Standards were made up from river water and a stock concentrate. One standard was usually sufficient, although multiple ones were used on the turbid Castleguard River. In highly concentrated samples, results were checked by quantitative dilution in case of nonlinearity. Discharge was calculated by dividing the mass of dye injected by the area under the time concentration curve, using appropriate units. A program for integration was written based on that of Church (1974), although a crude rectangular and residual triangle integration technique used in the field gave comparable results. The major potential source of error was in differences in the temperature of the standard and the samples. This was corrected in the computer program using the exponents recommended by P.L.Smart and Laidlaw (1977, Fig.A.3).

The rating curve was assumed to be of the form

$$Q = A\ (S-B)\ C$$

Where Q is discharge in cubic metres per second.

S is stage in units of length (inches or mm)

A, B, C are empicical constants.

The constants were found by optimisation. However, difficulties were encountered because unacceptably high values of C and low values of A were obtained. This was attributed to the roughness characteristics of the typical mountain stream, in which boulders broke surface at low flow, but were submerged at higher flows. The curve for the Cave Stream had to be broken up into component parts. This might have been preferable on other sections, but insufficient data were available. This problem is also a constraint on the extrpolation of the rating curve. Brief high discharges could not be gauged, and the rating curve probably over estimates the discharge. Figure A.1 is an example of a rating curve.

The Big Spring was difficult to gauge, because it could not be metered at high flow, the available mixing length was too short, and there were numerous inflows and an outflow along this length. Furthermore the stage in the plunge pool of the spring did not register increases in discharge beyond a certain level. The resulting rating curve is unsatisfactory, consisting of two straight segments, linked by an arbitrary circular arc tangent to both lines (Fig. A.2). The finite discharge at zero stage represents the continuance of flow in other springs. The change in slope occurs when the overflow to the Big Spring (the Camp Stream) starts to operate. The upper segment of the curve probably underestimates discharge at high flows quite substantially. The absolute value of the data is in doubt, but the relative variation is representative of reality.

The Castleguard River was very unstable in 1979, so that the data from that year were not used. The Meadows Stream (I) section was altered (and improved) by floods in the spring of 1980, but this was not noticed until re-rating was impossible. This accounts for the contrast between the two discharge series in 1979 and 1980.

The discharge data were stored as single data files for each site and each year. All time series data were stored as:

Title or caption

Data units

Hour on

Hour off

Data in groups of 12

The time was initialised at 1.00 h on the 13 of July and data reference in software was by hour from this time.

Coversion to and from date routines were written, but in general reference was to day number with 1377 as #1.

A set of interactive Fortran routines were written for plotting, transforming and selection of arbitrary sequences of data from any time period and any series.

A.4. DYE TRACING

The fluorescent dyes selected as tracers were Rhodamine WT, Fluorescein, and Lissamine FF (P.L.Smart and Laidlaw 1977). Dyes were decanted into preweighed polyethylene bottles and weighed before entering the field. Injection was directly into the stream sink. Wind drifting of powdered dye was initially a problem, but premixing in a bucket overcame this. After injection the area was thoroughly washed down, and bottles, clothing and personnel washed as necesary.

Various automatic water samplers were used as available: ISCO, Sigmamotor, and Northants Engineering. The Sigmamotor proved the most robust and reliable when operated from an external battery. The vacuum driven Northants was more independent, but had a fixed 2 hour sampling interval. Samples [illegible] taken at up to 5 per hour using a multiplex facility. [illegible]er frequency discrete single samples could not be obtained from remote sites. The sampler bottles were thoroughly rinsed in de-ionised water after each major trace, and during the trace using deionised, water passed through a column of activated charcoal, so that the water was dye free.

Five hundred 30ml sampling bottles were numbered and transported in rack boxes. They were pre-rinsed in clean water, and the caps soaked in water over night. Despite these precautions some contamination occurred, probably as a result of using the same bottles for tracing as for dilution gauging in which high dye concentrations were encountered. It is recommended that solid caps be used in future, and that a different set of bottles be used for each purpose. Bakalowicz (pers comm.) recommends the use of a hydrogen peroxide solution for cleaning, because it will oxidise any residual dye.

Samples were stored in cold dark conditions and analysed at night. This is because the fluoresence of Rhodamine WT is highly temperature dependent (Fig.A.3), and a marked improvement in detectability can be gained. Analytical cuvettes were rinsed twice in sample and wiped dry before analysis. Sampling sequence was preserved to limit the effect of serial contamination, and reduce the time spent in reajusting range on the fluorometer. Sample temperature was measured to the nearest 0.25 degrees Celsius, and recorded with fluorescence.

The fluorometer was a Turner Designs Model 10 Series Fluorometer, powered by 12 volt battery or smoothed 110v AC from a gasoline driven alternator. The machine proved robust and highly sensitive. Resolution of Rhodamine WT was to 0.01ppb, a significance level obtainable only because the Castleguard water had negligible background. (Higher backgrounds were found for green dyes, up to 1.0ppb.) Adjustments were available for blanking out of background levels.

Standards were prepared from 1ppm stock solution, using river water for dilutions, and clean water for traces. Maximum sensitivity was used in traces so that background variations could be observed. They were corrected later numerically. Background samples provided instrument zero in dilutions.

A single standard was sufficient for calibration in the linear range. A span adjustment allowed direct reading in concentration. The temperature of the standard was recorded. The instrument was calibrated before and after each batch of samples, although this was found to be unnecessary.

The filters employed are those recommended by Smart and Laidlaw (1977), with light sources suggested by the manufacturer. Changing filters was a somewhat tedious procedure compared to that on the better known Turner 111 Fluorometer.

The instrument was also operated in flow-through mode, driving a Rustrak 388 Recorder. A record of fluoresence and instrument range was obtained. In this context the automatic ranging of the instrument was indispensible. The total power consumption was considerably greater in flow through mode, and no control over temperature was possible. The latter problem was overcome by passing the incoming water pipe through a coil condensor in a cold spring-fed creek immediately before passing it through the instrument. Calibration was managed by a special pipe and plug, which allowed rinsing and multiple check calibrations without using excessive volumes of standard.

The quality of data provided by continuous analysis is believed to be unsurpassed, and it is strongly recommended for high resolution results in relatively fast flow systems. The main advantage is the production of noise free data, and the avoidance of any suspicion of aliasing.

As a nephelometer, the instrument was excellent, although conversion of the instrument was unnecesarily awkward. A colloidal silica solution "Ludox-HS" provided a stable, but arbitrary standard. Difficulties in the interpretation of nephelometric data meant that no attempt was made to convert readings to turbidity (see Chapter 4).

Dye concentration data were entered into data files, corrected for temperature effects (only necessary for Rhodamine WT), and standardised to concentrations equivalent to a 1000g injection. Where discharge data were available, the mass breakthrough curve was also calculated.

A set of interactive 2-D and 3-D Fortran interactive plotting routines were developed for monochrome and colour presentation of the data.

The tracer data were analysed using a modified version of a program written by R.Kellerhals (in Church 1974) which uses a third order integration routine with a smoothing function. In addition, a dispersion term D/UL was estimated following Brady and Johnson (1981) (where D is the dispersion coeffecient, U is average tracer velocity, and L is path length). No assumptions concerning the flow path are made in this definition, but it is not directly comparable between different tracer links. An exponential extrapolation to infinity was included in the calculations where appropriate. Dye recovery at each spring was calculated using measured or approxinmated discharge.

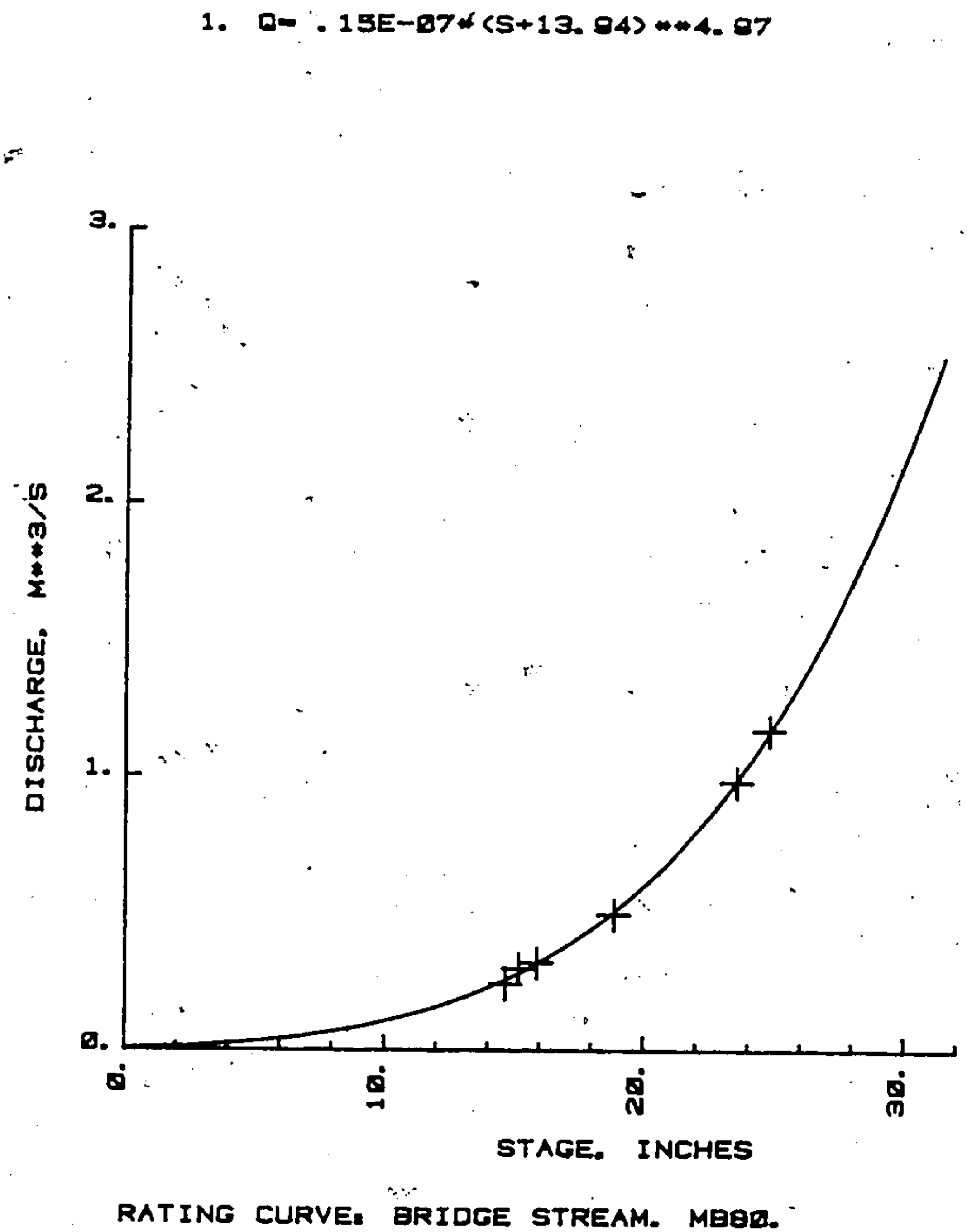

Figure A.1. Discharge rating curve for the Meadows Creek (#2) stage recorder

1. Q= .60E-01*(S+ 4.34)**1.00

2. Q=10.75-(87.37-(S-18.59)**2)**0.5

3. Q= .28E+00*(S+-14.9)**1.00

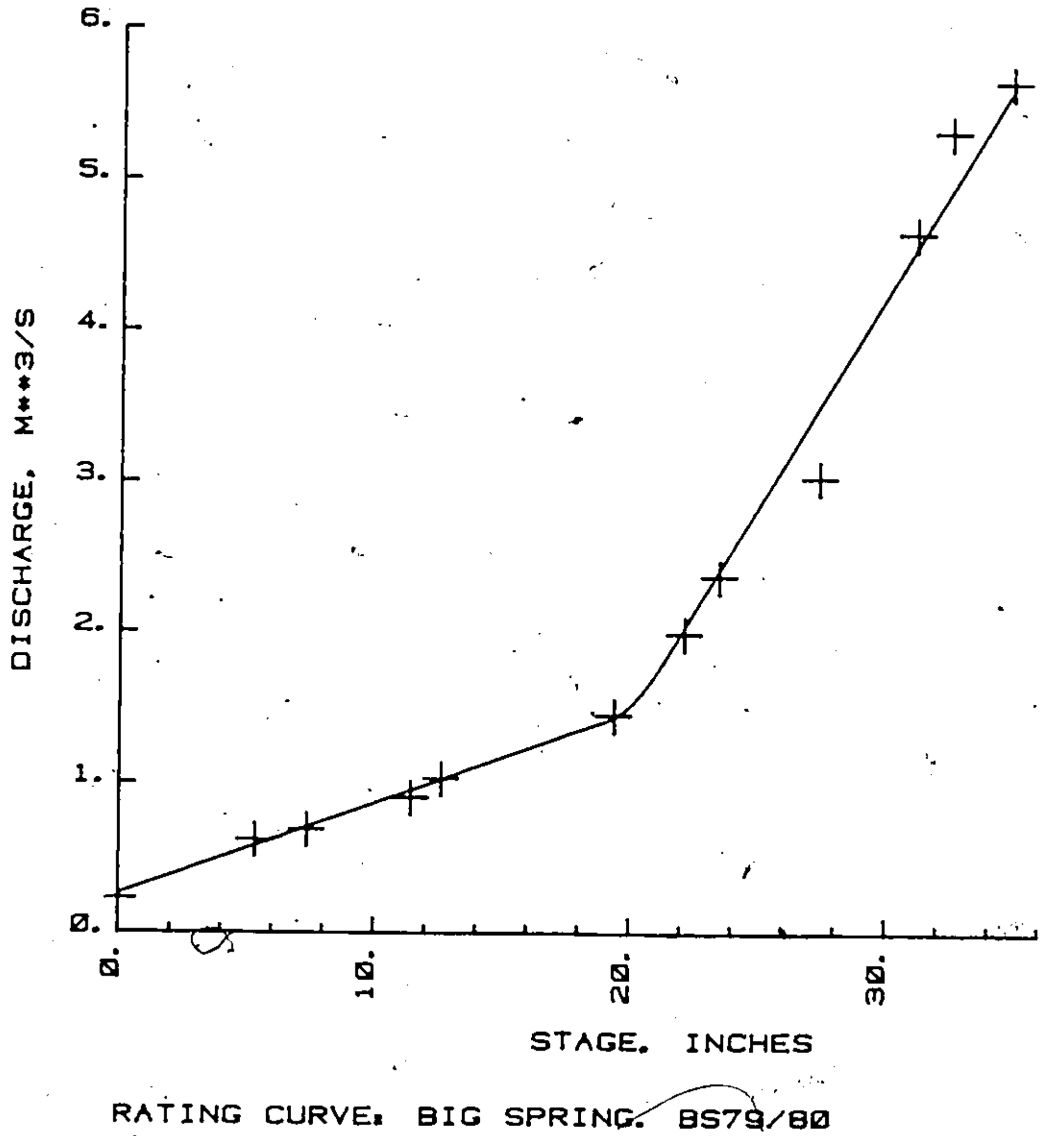

Figure A.2. Discharge rating curve for the Big Spring

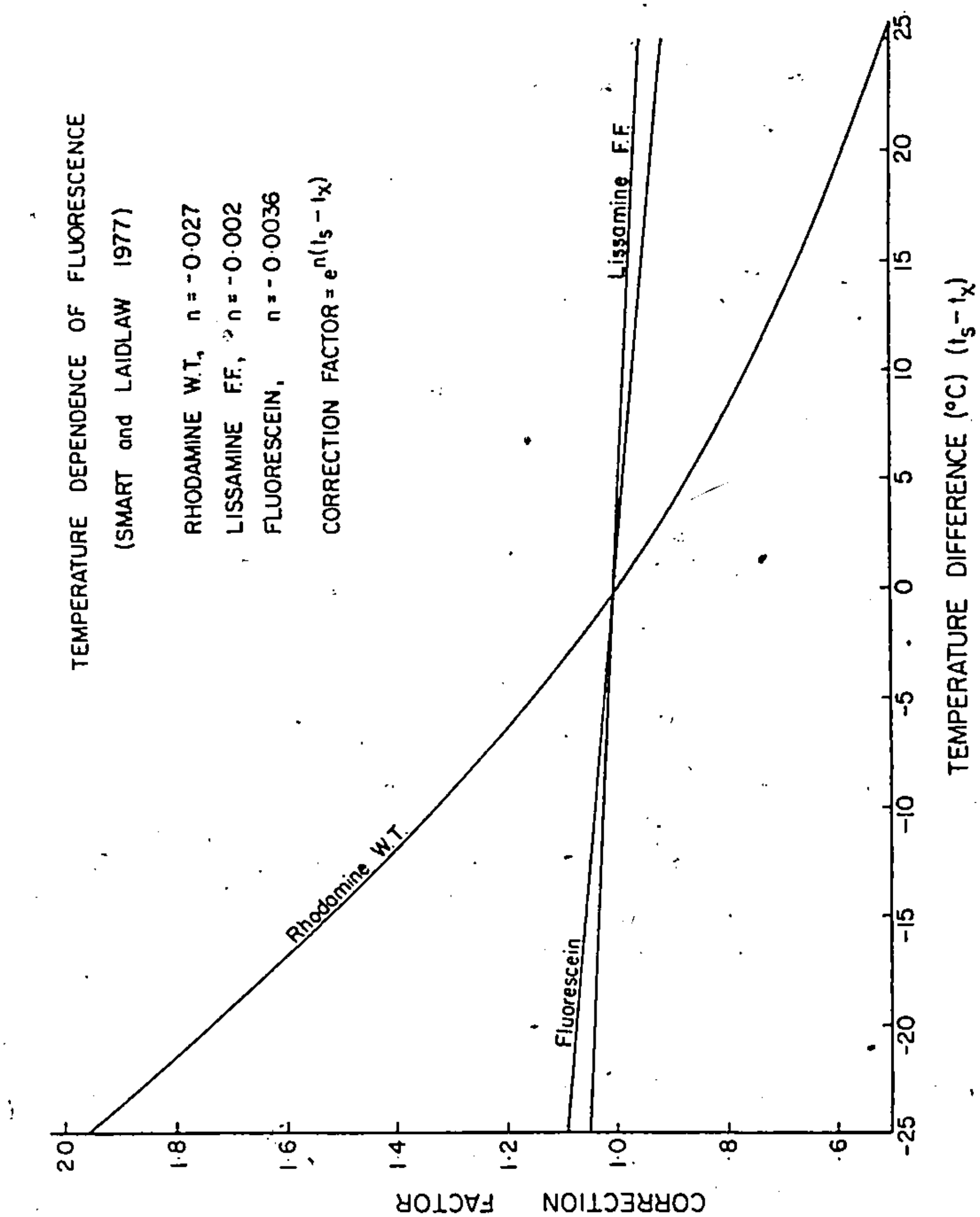

Figure A.3. The temperature dependence of fluorescence for certain dye tracers