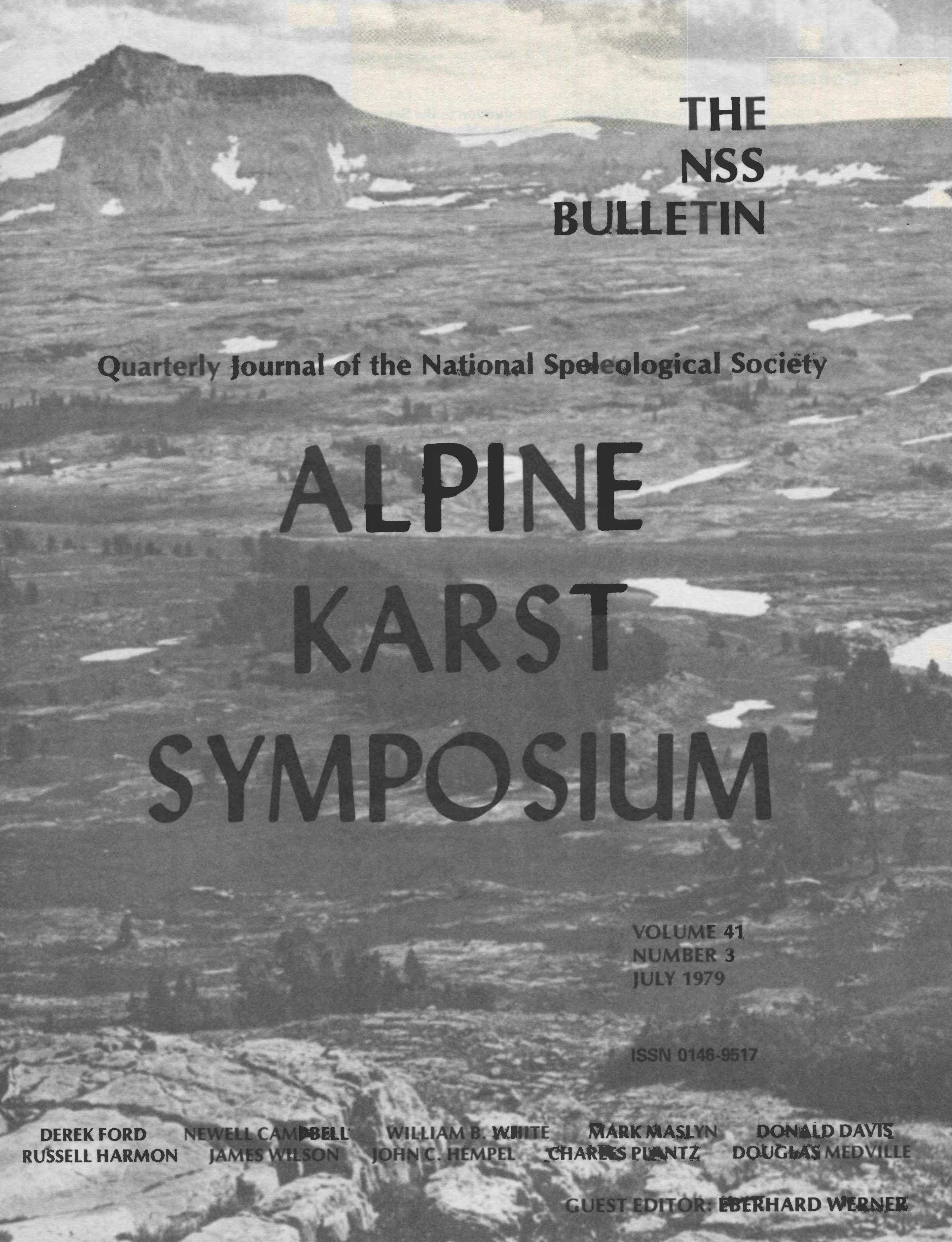

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Contents

Cover Photo: A portion of the Tosi Creek Basin in the Gros Ventre Range of Wyoming. This is a large (3 x 5 km) structural surface developed on the Middle Mississippian Madison Group limestones. Except for the mountains in the background, the entire area pictured is at nearly the same stratigraphic horizon. (Eberhard Werner)

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ALPINE KARST IN THE ROCKY MOUNTAINS —INTRODUCTION TO THE SYMPOSIUM

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KARST LANDFORMS are developed in many parts of the Rocky Mountains at high elevations. These alpine karst areas are distinct from low-altitude karsts of the same latitudes. Variations in chemical and physical factors create different solutional landforms, and the higher hydraulic gradients (due to greater topographic relief) cause different geohydrologic patterns. The landscapes have a different appearance because of the permanent snowfields and poorly developed soil and vegetative cover.

A problem encountered by most investigators has been defining the term *alpine karst*. A working definition has been largely developed by concensus among the various investigators. If we look at the various terrains, we see that an alpine karst is one which is at a high altitude, normally at or above the tree line, and where all surface water is frozen for an extended period every year.

Several characteristics of these terrains differentiate them from low altitude karst terrains. Because of the high topographic relief, most alpine karsts have very high hydraulic gradients. This implies high flow velocities and short residence times for the conduit waters. Thus, the water has less time for solution, and larger volumes of water are required to produce the same void space as much smaller volumes would create in areas of lesser topographic relief. Alpine areas in general, and alpine karsts in particular, tend to have thin or no soil cover. Relatively recent alpine glaciers stripped most of the areas of soil and loose rock, and the relatively severe weather tends to slow soil production by retarding chemical and biological weathering. Developed on the bare carbonates are a variety of solution features, mostly forms of karren, which are not commonly seen in low-altitude karsts. Solution is usually fairly evenly distributed over the entire area. Only where water is channelled into significant streams before the carbonate is reached will there be solution concentrated enough to form sizable caves.

Alpine karsts in North America are found primarily within the Rocky Mountains. A few areas do exist in the Sierra Nevada of California, the Northern Cascade Range of Washington, and in the mountains of northeastern Canada; however, since virtually all scientific work to date has been in the Rocky Mountains, this symposium was restricted to that area. The accompanying map (Fig. 1) shows those areas known to have alpine karst.

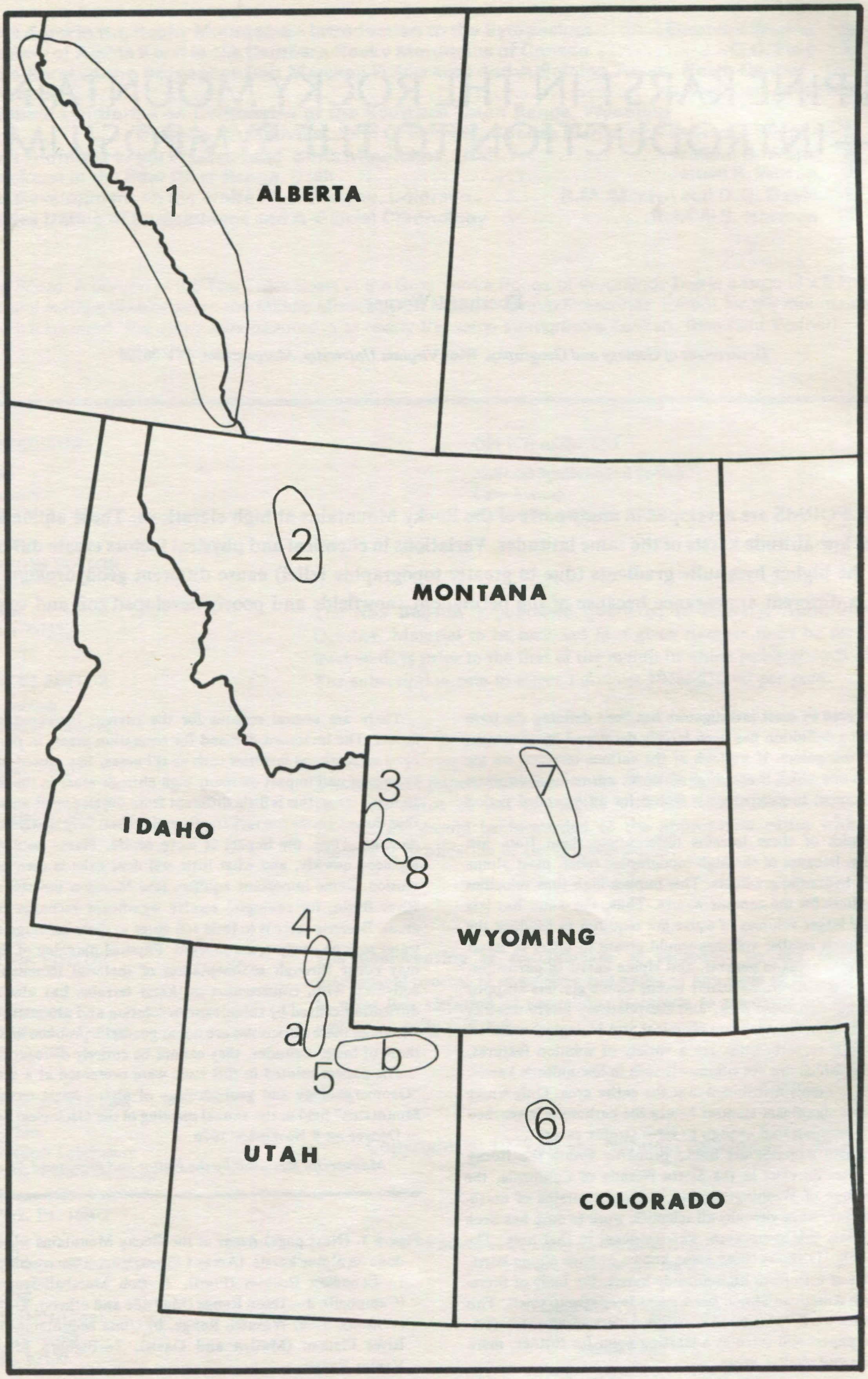
Compared with that of European high-altitude karsts, the body of literature existing for North American alpine karst areas is relatively small. This Symposium summarizes work so far done in North American alpine karsts. It is hoped that these papers will serve as a starting point for further, more detailed work in these and similar areas.

There are several reasons for the current increased interest in alpine karsts. The increased demand for recreation areas, in particular for developed recreational facilities such as ski areas, has caused a great increase in the use of and impact on many high altitude areas in the Rocky Mountains. In many ways, this is little different from development anywhere else, except that these terrains are very fragile and recover very slowly from environmental damage. Thus, the impact is more severe. Plants once disturbed are not replaced quickly, and what little soil does exist is then easily removed by erosion. Some important aquifers (the Madison limestones of the Powder River Basin, for example) receive significant recharge from alpine karst areas. Because there is so little soil cover in these recharge areas, the ground water may be easily contaminated. Physical plugging of the recharge paths may occur through sedimentation of material loosened by construction activities. Also, construction on karst terrains has always been beset by difficulties caused by subterranean solution and attendant surface collapse. Although these difficulties are not as general a problem in alpine karsts as in those of lower altitudes, they cannot be entirely disregarded.

The papers printed in this issue were presented at a symposium entitled "Geomorphology and geohydrology of alpine karst terrains of the Rocky Mountains" held at the annual meeting of the Geological Society of America in Denver on 8 November 1976.

Manuscript Received by the Editor and accepted 1 September 1978.

Figure 1. (Next page) Areas of the Rocky Mountains where work has been done on alpine karsts. (Areas 1 through 6 are discussed in this issue.) Area 1—Canadian Rockies (Ford), 2—Bob Marshall-Scapegoat Wilderness (Campbell), 3—Teton Range (Medville and others), 4—Bear River Range (Wilson), 5—a) Wasatch Range, b) Uinta Mountains (White), 6—White River Plateau (Maslyn and Davis), 7—Bighorn Mountains, 8—Gros Ventre Range.


ALBERTA
1
MONTANA
2
IDAHO
3
7
8
4
WYOMING
a
b
5
6
UTAH
COLORADO

Ford, DC (1979)—A Review of Alpine Karst in the Southern Rocky Mountains of Canada: *NSS Bulletin* 41:53-65.

A REVIEW OF ALPINE KARST IN THE SOUTHERN ROCKY MOUNTAINS OF CANADA

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ABSTRACT

There is extensive karst development in Paleozoic carbonate and sulphate strata of the Front and Main ranges, southern Rocky Mountains of Canada. Its characteristics are summarized under six headings: ***1) Lithology and structure.*** *Karst is best developed in massively bedded middle Cambrian, upper Devonian, and Mississippian limestones. Parakarst occurs in thinner or more mixed limestone and dolomite units. Gypsum is karsted wherever it crops out. Because of high hydraulic gradient, aquicludes of shale, etc., may be breached at favourable locations.*

2) Spatial and typological patterns of carbonate solution. *Waters are of the bicarbonate type; measured* $CaCO_3$ *hardness ranges from 10 to 275 mgl. Six classes are differentiated by hardness and by saturation characteristics — tundra and glacial meltwater, surface water below the treeline, karst springs, seepage waters in caves, regional rivers, and "depleted waters" from high-altitude sub-ice/ -firn sites. Sub-glacial calcite precipitates are associated with the "depleted" waters, which are the one uniquely "alpine" type.*

3) Groundwater systems and caves. *Holokarst, fluviokarst, and no karst drainage are found in similar lithologic and topographic settings. Topographic divides (including the Continental Divide) are frequently breached by groundwater drainage. Groundwater systems have no consistent relationship to geologic structure. The most common type of spring is young and hangs above the adjacent valley floor. Mature (cave) springs at base level are rare. Many spring sites have been aggraded by drift. Caves are predominantly either of the deep-phreatic type (due to steep rock dip) or of the invasion vadose type. Shafts are common but are usually obstructed.*

4) Karst groundwater systems and glacial hydrology. *Karstic circulation may be suppressed or accelerated beneath temperate glaciers. Features indicating suppression (full-glacial conditions) are minor phreatic solution of host rock, major re-solution of speleothems, and partial to complete filling with clastic debris. Features of acceleration (early- and late-glacial conditions) are rapid re-excavation, re-routing (via new invasion caves), and/or abandonment of caves.*

5) Types of surface karst landforms. *There are six ideal categories of alpine karst forms: a)* ***postglacial*** *forms show no inherited glacial morphology, although their distribution may be determined by glacial factors; b)* ***subglacial*** *karst forms may be erosional (shafts) or depositional (precipitates); c)* ***karstiglacial*** *forms are karstic adaptations of glacial closed depressions in bedrock or in drift; d)* ***glaciokarstic*** *forms are karst features modified by glacial scour or deposition; e)* ***mixed*** *forms display mingled glacial and karstic action from several different times; and f)* ***preglacial*** *forms are limited to abandoned and fragmented caves. In practice, it is difficult to discriminate categories 2 through 5, but Canadian examples are illustrated.*

6) Altitudinal zonation of surface karst. *Successively higher doline, karren, and frost shatter (periglacial) zones have been described from the Alps and other mountains. In the Canadian Rockies, a boreal zone, a tundra zone subjected to Wisconsinan glaciation, and a tundra zone that escaped Wisconsinan glaciation are recognized. Karst forms are limited in the boreal zone because of the depth of soil required to support forest. The glaciated tundra displays the greatest frequency and variety of karst; there is no sub-zonation of dolines and karren. Frost shatter dominates the unglaciated tundra. Solutional weathering seems often to be a necessary prerequisite to frost action on glacially scoured rock; therefore, care must be taken in attributing frost degradation of karst to climatic deterioration.*

INTRODUCTION

THIS PAPER summarizes work in the Rocky Mountains of Canada since 1965, by a team centered at McMaster University. Rather than to describe one by one the highly varied localities we have studied, it aims to offer broad conclusions under a number of systematic headings. It is hoped that this approach will be more useful for workers seeking comparisons or contrasts with their own findings in U.S. alpine terrains. Like other broad reviews, this paper contains simplifications and generalizations for which exceptions can be found in the Canadian alpine karsts as well as elsewhere.

The area considered is the southern half of the Canadian Rocky Mountain system, from the U.S. border to the latitude of Edmonton (54°N). It is approximately 560 km in length, 80 to 130 km in breadth. Summit elevations generally range from 2100 to 3300 m asl. Trunk-valley floors are at 1100 to 1500 m, giving a local relief that varies from 600 to 2000 m. Carbonate strata are the most abundant mountain-forming rocks.

Access is poor. Only three paved roads cross the region, although a fourth winds through much of it. Many valleys and most of the high country are without trails of any kind. We have seen, at most, some 10% of this large area on the ground and have studied, perhaps, a further 10% with some intensity on air photographs. We have found conventional (black and white) photography to be of comparatively little help in detecting new karst, especially caves, but it has been the basic tool in elaborating findings once some ground truth is established.

The mountain geomorphology is dominated by the forms of alpine glacial erosion, *i.e.*, cirque, arête, and horn assemblages, U-shaped valleys, hanging valleys, etc. In this respect, one important, initial contrast may be drawn with a majority of U.S. alpine terrains. The Canadian Rockies were thoroughly inundated with flowing, temperate-glacier ice during the last (late Wisconsinan) ice age, which peaked about 20,000 years B.P. All cirques, benches, and valleys were occupied by ice, which drained west to fill the Rocky Mountain Trench or east onto the Prairies, where it was often confluent with the continental (Laurentide) ice sheet draining from the northeast. Except in the northern ranges of Montana and Idaho, ice cover was not so extensive in the U.S. Trunk glaciers there were shorter and debouched onto ice-free valleys or basins, as they did in the well-studied alpine karsts of the Alps, Caucasus, and Pyrenees. At certain stages of a glacial cycle, the presence of temperate (waterfilled) ice may accelerate karst development (page 59). In principle, the potential for such acceleration was greater at the U.S. sites than in Canada, because of the closer juxtaposition of ice-covered and ice-free ground, which favoured higher hydraulic gradients.

FEATURES OF THE GEOLOGY

Strata of Upper Proterozoic to Cretaceous age occur in the southern Canadian Rocky Mountains, but Palaeozoic rocks predominate. The mountains are a thrust belt (Douglas, 1970), stratal plates being pushed from the west along low angle thrust planes. The eastern ranges are the "Front Ranges," where mountain-forming strata of Upper Palaeozoic age are thrust repeatedly over weaker Mesozoic strata that now appear as strike-aligned, southeast-northwest trending vales. Structures are predominantly homoclinal, with dips generally greater than 15 to 20° west. In the 'Main Ranges,' Lower Palaeozoic and some Proterozoic strata are overthrust repeatedly upon the Upper Palaeozoics. Plates were thicker and more rigid than in the Front Ranges, so that structures are more complex. There are plateaus and much fold topography, in addition to homoclines.

Apart from the thrust planes, faulting is of little importance in Rocky Mountain topography. Except at very local scale within a cave, it does not appear to be of significance in the karst development. Such development is determined by lithology, dip, strike, and jointing.

The karsted formations are limited almost entirely to the Palaeozoic, of which a simplified type section is given in Figure 1. In the upper Proterozoic, the Appekunny formation is a dolomite that crops out at Waterton Glacier National Park, on the U.S. Border. Some local underground drainage has been reported there, but we have not investigated. No karst is known in the Mesozoic rocks; they are predominantly clastic, although some thin limestone units occur.

In the Palaeozoics, the Lower Cambrian is represented by the Gog Group, predominantly resistant sandstones and quartzites that are major mountain builders. But sandwiched within the northern part of the area is 100 m of massive, well-bedded, crystalline limestone, the Mural Formation. Two neighbouring caves are known in it, of which "Arctomys" at -522 m is the deepest system in Anglo-America. It is a simple river channel developed down a 30 to 35° stratal dip (Thompson, 1976).

The Middle Cambrian contains one of the principal karst sequences of the Main Ranges. It is also one of the most remarkable and unusual that is known to me anywhere. The Cathedral Formation is an exceptionally massive, crystalline limestone. Individual beds appear to exceed 5 m in thickness. Associated is a very low frequency of joints that are, however, of exceptional length and depth. Neptunian dikes may be over 300 m in depth. These characteristics determine the nature of the remarkable Castleguard Cave, where passages of one shape, size and gradient are maintained for great distances (Ford, 1971a, 1975; Ford, *et al.*, 1976). The Pika and Eldon formations are marked by rhythmic alternations of limestone and dolomite 1 to 5 cm thick. But, parting planes or joints penetrable by water are rare, so that, like the Cathedral, they function as massive rocks. The intervening Stephen Formation contains thick shales. It is often an aquiclude.

AGE			THICKNESS (METRES)		FORMATION	DESCRIPTION	CAVES
MISSISSIPPIAN	UPPER	CHESTERIAN	0-60	RUNDLE GROUP	ETHERINGTON	Marine lst., dolomite (some chert, siltstone, evap.)	Crowsnest caves Maligne system Disaster Point Plateau Mt. Chungo Cadomin
		MERAMECIAN	160-215		MOUNT HEAD		
	LOWER	OSAGEAN	0-150	LIVINGSTONE	TURNER VALLEY		
			60-120		SHUNDA		
			60-150		PEKISKO		
		KINDERHOOKIAN	180-240		BANFF	argillaceous lst.	
					EXSHAW	shale	
DEVONIAN	UPPER	FAMENNIAN	30-110	PALLISER	COSTIGAN	lst., dolomite	
			120-300		MORRO		
			100-110		ALEXO-SASSENACH	sandstone, dol.	
		FRASNIAN	150-270	FAIRHOLME GROUP	SOUTHESK	mostly dol., some lst.	
			150-335		CAIRN		
			75		FLUME		
	M	GIVETIAN	0-35		YAHATINDA	sandstone, shale	
SILURIAN			210		BEAVERFOOT	lst., cherty dolomite	Top of the World caves
ORDOVICIAN	UPPER	CINCINNATIAN	180		MOUNT WILSON		
	MIDDLE	CHAMPLAINIAN	210		OWEN CREEK	some lst., most cherty dol. & shale	
			180		SKOKI		
	LOWER	CANADIAN	440		OUTRAM		
			520		SURVEY PEAK		
CAMBRIAN	UPPER	TREMPEALEAUAN	150		MISTAYA	some lst., most dol., shale & siltstone	
		FRANCONIAN	210		BISON CREEK		
			365		LYELL		
		DRESBACHIAN	165		WATERFOWL		
	MIDDLE	ALBERTAN	230		ARCTOMYS	Shale	
			275		PIKA	lst., dolomite	
			335		ELDON		
			135		STEPHEN	mostly shale	
			365		CATHEDRAL	lst., dolomite	Castleguard Cave
			135		MOUNT WHYTE	lst., shale	
	LOWER	WAUCOBAN	2130	GOG GROUP		sandstone, shale	
					MURAL	lst.	Arctomys Cave
						sandstone, shale	

Figure 1. Generalised section of the Palaeozoic lithology in the southern Rocky Mountains of Canada, from Halladay and Mathewson (1971).

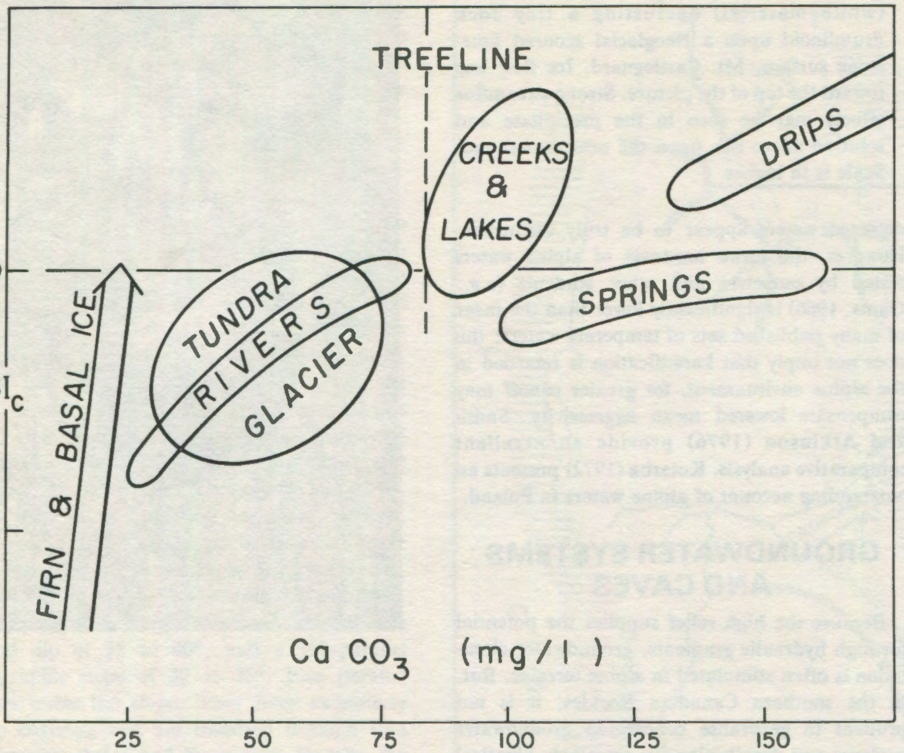

Figure 2. Plot of Ca^{2+} hardness (as $CaCO_3$) versus the Saturation Index with respect to calcite, to display differentation of alpine karst waters. Outlines are intended to be schematic and indicate areas into which the majority of the water groups will fall.

However, it is breached above Castleguard Cave, so that groundwaters drain from the top of the Pika to springs close to the base of the Cathedral.

Between the Middle Cambrian and Upper Devonian, there are many limestone and dolomite units. They tend however, to be thin and to be separated by sandstone and shale sequences, so that karst development is of the 'parakarst' type (Birot, 1954), *i.e.*, karren, sporadic small sinkholes and short groundwater channels. Some small caves are known in these rocks at 'Top of the World' (Thompson, 1976).

The second great karst sequence of the Canadian Rockies extends from the Palliser Formation of the Upper Devonian to the Rundle Group (Livingstone, Etherington, and Mt. Head formations) of the Upper Mississippian. The Palliser is a classical karst rock, crystalline, massively bedded, and resistant; it is recognized everywhere in the region. The Rundle is similar in character and is the principal karst rock of the Front Range. In between, the Banff Formation is a limestone containing many continuous chert beds and much shale, and the Exshaw Formation is a carbonaceous shale. These two formations often combine as an effective aquiclude, so that separate karst systems develop on the Rundle above and on the Palliser below. But, they are breached by underground drainage at Crowsnest Pass and perhaps elsewhere. This permits 1500 m of stratigraphic penetration by groundwaters.

To summarize: well-bedded and massive limestones host the best karst, as in other areas. But, comparatively few outcrops of these best rocks appear to be karsted, except at the scale of karren; there has been effective competition from glacial, fluvial, and mass wasting processes operating at the surface. This is especially true where high stratal dip and topographic slope are concordant. We have found little significant karst upon dolomites, except where they may be drained into underlying limestones, as at Castleguard. However, cherty and shaly sequences or purely shale formations that, in other areas, might be expected to be complete aquicludes are frequently breached by groundwater in the mountains.

SPATIAL AND TYPOLOGICAL PATTERNS OF CARBONATE SOLUTION

Figure 2 is a plot of dissolved calcium carbonate (calcite) concentration versus the saturation index (SI_c) with respect to calcite. SI_c is calculated by the method of Langmuir (1971). The figure indicates where, with respect to these two variables, we expect to plot the majority of the different types of waters that we sample in the southern Canadian Rockies. The distribution is based upon two data sets:-

first, approximately 600 spring and summer (*i.e.*, meltwater season) samples of smaller waters of all types from limestone terrains. Included are lakes, creeks, karst springs, cave streams, seepages, and drips from stalactites, snowmelt and glacier melt, tundra bogs, etc. (Ford, 1971b and later collections). Samples come from above and below the treeline and range in discharge from <1 ml/min to 60 cubic metres/ second (cumecs).

second, three water years of record for the Athabasca and North Saskatchewan rivers—(Drake, 1973; Drake and Ford, 1974, 1976). These have two of the largest basins in the mountains. Each extends over both the Front and the Main ranges and is comprised of approximately 60% carbonate rock. Mean discharge exceeds 550 cumecs during the peak flow month (June or July).

The waters are predominantly of the bicarbonate type (*i.e.*, Ca^{++} and HCO_3 are the dominant ions in solution, with lesser Mg^{++}). There is an important sulfate component in the two big rivers (Drake and Ford, 1974). The waters fall into the following classes:-

1. Waters from all kinds of vegetated tundra, and from well aerated glacier melt streams. These are normally undersaturated with respect to calcite but approaching saturation at values of 60 to 80 mg/1. This is quite accordant with the equilibrium solubility of calcite, where the P_{CO2} (partial pressure of CO^2) = the mean global atmospheric value and water temperatures are close to O°C. There is much local variation on the tundra, waters being somewhat enriched above the illustrated range where vegetation is thickest. Interested readers should also see Woo and Marsh (1976) for an excellent account of small waters on a Canadian high arctic tundra.

At glacier snouts we find that the more turbid a melt stream is, the more nearly it is saturated with respect to calcite.

2. Waters of creeks and lakes below the treeline. These are saturated, or slightly supersaturated and de-gassing, at concentrations generally between 80 and 120 mg/1 $CaCO_3$. Their main sources are soil water and throughflow from forested, carbonate-rich tills, glaciofluvial, and colluvial deposits. Analysis indicates slight enrichment by soil CO_2. It must be emphasized that, even at these higher latitudes, the bulk of the mountain area lies below treeline and that the bulk of mountain solution takes place in forested detrital deposits—although spectacular solution landforms are rarely associated with it in the Canadian Rockies. Nicod (1972) reports similar findings from the Alps.

3. Major karst springs. Most of these are located at or below the treeline and are just saturated at 125 to 160 mg/1 $CaCO_3$. Their discharge incorporates much water from tundra sources, but the forest zone evidently dominates. Unlike Class Two waters (surface creeks and lakes), a trend toward increasing hardness in karst spring water may be discerned as the summer season advances. This trend is interrupted by flood pulses from rain storms or periods of exceptionally rapid snowmelt, however.

4. Groundwater seepage (drips and trickles) in caves. Some are from tundra, but the majority of our samples are of forest origin and appear to be

depositing calcite. The waters are supersaturated with respect to the cave environment in which they were collected and have ranged from 130 to 275 mg/l $CaCO_3$. These waters display a strong trend toward maximum hardness in August and September, although once again there is dilution by storm effects. It is clear that soil CO_2 dominates the chemistry of their aggressivity with respect to calcite.

5. Major rivers. In decreasing order of importance, their waters are derived from Classes two, one, three, and four above. As Figure 2 indicates, they tend toward an equilibrium concordant with global mean P_{CO2} and water temperatures of 0 to 10°C. But, there is a considerable 'tail' or range down toward lower values, explained by flood dilution and inputs from non-carbonate terrains that we do not sample specifically.

6. "Depleted waters", evolving toward saturation with respect to calcite at >25 mg/l, or only some 33% of the expected saturation value at 0°C, their usual temperature. These are high altitude trickling waters from the melt of firn (old, compacted snow) or small glaciers. They were taken close to their source, and so are little aerated. Included are waters taken from Castleguard Big Springs (Ford, 1971a), which debouch below treeline but are inferred from their thermal and chemical characteristics to be derived from the base of the Columbia Icefield, the largest extant glacier in the Rocky Mountains.

Ek (1966) has reported similarly depleted waters from the Alps. Depleted aggressivity may be attributed to a combination of the minor depletion of global atmospheric P_{CO2} that occurs with altitude (Drake, 1973), evolution under ideal closed system conditions because the melting ice or snowpack is saturated with water (phreatic!), and probable expulsion of CO_2 from aging ice and snow (Ek, 1966; Cogley, 1975).

Associated with the depleted waters are a peculiar suite of minor alpine karst landforms: sub-glacial calcite precipitates. The finest display yet known is found at the margins of glaciers receding across flat surfaces of Cathedral and Pika strata at Mt. Castleguard and nearby sites (Ford, Fuller and Drake, 1970). Hallet (1976) attributes such deposits to pressure melting and refreezing about obstacles at the glacier base. The morphology of many individuals supports this hypothesis (Figure 3a), but the calcite cover is so widespread and uniform at places around Castleguard that such a localised mechanism seems inadequate, (Figure 3b). "Sub-glacial speleothems" are transient karst phenomena: a few tens of years after the ice recedes, laying them bare, they are destroyed by rainwater or by the melt of fresh snow.*

The range and characteristics of most of the alpine karst waters described here can be duplicated in many non-alpine karsts. Only the depleted waters appear to be truly distinctive. However, the mean hardness of alpine waters tested by ourselves and other students (*e.g.*, Gams, 1966) is significantly lower than the mean of many published sets of temperate waters; this does not imply that karstification is retarded in the alpine environment, for greater runoff may compensate lowered mean aggressivity. Smith and Atkinson (1976) provide an excellent comparative analysis. Kotarba (1972) presents an outstanding account of alpine waters in Poland.

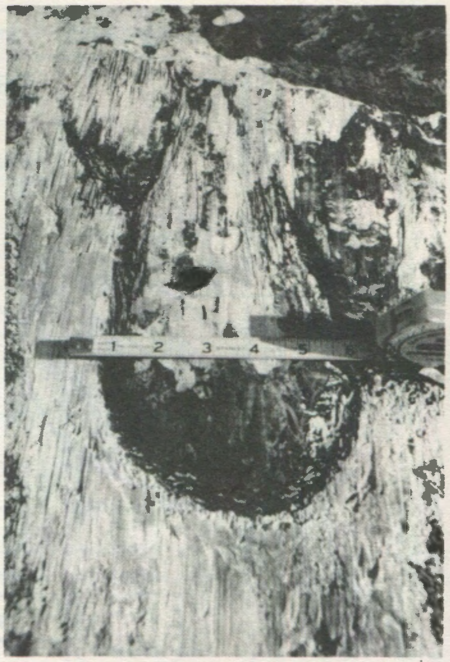

Figure 3a. (above) Sub-glacial calcite precipitate (white material) encrusting a tiny rock drumlinoid upon a Neoglacial scoured limestone surface, Mt. Castleguard. Ice flow was toward the top of the picture. Strong streamline effects may be seen in the precipitate and solution micro-rills upon the near (stoss) end. Scale is in inches.

GROUNDWATER SYSTEMS AND CAVES

Because the high relief supplies the potential for high hydraulic gradients, groundwater circulation is often stimulated in alpine terrains. But, in the southern Canadian Rockies, it is not prudent to generalise concerning groundwater patterns and distribution. In seemingly identical lithologic and topographic, etc., situations, one area may display holokarstic (entirely subterranean) drainage whilst another has entirely surficial drainage. This may often be taken as an indication that the holokarst has a history of karstic drainage development antedating the development of much of the particular modern topography or relief.

There are areas where the state is intermediate between holokarstic and wholly surficial drainage. One type is fluviokarst, where there is a perennial stream in the valley but underground drainage in the flanks. Another is where drainage is entirely underground, save at the peaks of exceptional melt or rainfloods. Certain mountain slopes in British Columbia (west of the Continental Divide) drain exclusively to Alberta (east of the Divide), except when it is raining hard. Topographic divides are sharply delineated in alpine areas, but they are not sure guides to the location of groundwater divides (as has been shown many times in the Alps and Pyrenees).

The orientation of groundwater systems, sink to spring, with reference to gological structure shows every variation, *e.g.*, down-dip, up-dip, along strike, aslant dip, and aslant strike. An excellent general illustration is given by the karst systems at Crowsnest Pass (Fig. 4). An aquiclude (Exshaw Formation) striking northwest-southeast is intersected by the east-west pass, dividing the karst unit into four blocks. Each was first drained

Figure 3b. (below) A broader view of a Neoglacial surface on Pika Formation strata, Mt. Castleguard. Ice flow was towards camera. Precipitate (white) covers 80 to 90% of surface.

* The author would very much like to hear from American workers who find sub-glacial calcite precipitates or "depleted waters" (or waters from similar firn and glacier sites that are not depleted).

along strike to the pass. The blocks were then dissected by deep, glacial cirque and valley cutting. Some east-face cirques are now drained underground to springs at the base of the riegel (cirque-front rock step); *i.e.*, groundwater drains up-dip. Others are drained on the surface. West-slope cirques and valleys are underdrained to the west, *i.e.*, down-dip groundwater drainage. Strike-aligned drainage above the aquiclude is preserved in the southwest block, where young cave systems have developed along the flank of a deep glacial valley. The valley discharges some 20% of its expected runoff, whilst an additional 100% appears in a spring perched on the flank at the mouth. The excess 20% represents groundwater abstracted across an anticline at the Continental Divide. Immediately north of the pass, glacial deepening of relief has permitted groundwaters to breach the aquiclude, and modern drainage is oriented aslant dip and strike.

An instance of the truly exceptional - indeed, improbable, groundwater systems that may occur in alpine terrains is found in Surprise Valley, Jasper National Park. The valley is strike-aligned. The north flank is in Fairholme rocks, has a stratal dip of 35 to 40°, and a sub-parallel topographic slope of 30 to 35°. Five parallel gullies incise the slope. They have excessively steep channels and are trenched through thin sandstones, shales and limestones. Despite the presence of stream channels to discharge runoff straight down the excessively steep slope, the waters of four gullies are abstracted into a limestone 4 m thick and conveyed along the strike (horizontally) to a spring in the fifth. There is surface overflow only in very wet weather. The groundwater system is above the altitude attained by Wisconsinan glacier surfaces, so that it cannot be attributed to recent glacier marginal flow: yet, it is of small dimensions, appearing to be young.

Summarizing, we find every relationship between topography and the situation and orientation of groundwater flow. Valleys may be underdrained to their mouths, especially where they are hanging valleys. Groundwaters may be oriented up-dip, down-dip or along strike to underdrain them. Groundwaters may drain across the axes of valleys. They may drain along the valley flanks, above or below the elevations of the valley floor. They may cross topographic divides.

Fig. 5 indicates the principal categories of karst springs. The most commonly observed type is young, the water being discharged from a crevice that is impenetrably small. It is also 'hanging' above the valley floor, although limestone may extend to and below the floor (Fig. 6). Often, the bedrock orifice is masked by kame, scree, or other slopefoot detritus. Young, hanging springs are unusual in extra-glacial terrains. Their predominance here reflects glacial entrenchment which disordered karst drainage.

The mature karst spring, where waters discharge from a well-formed cave that may be of enterable dimensions and which is located at or close to the elevation of a valley floor, is rare. In fact, there is only one unequivocal example known to us in the Canadian Rockies. This is Old Man Cave, on the north shore of Crowsnest Lake in Crowsnest Pass (Thompson, 1976). It is close to the stratigraphic base of the Palliser Formation and only 12 m above the lake surface. Other springs discharge below the surface. Old Man Cave is an attractive site, reached by a short walk from the highway through the pass.

The final type of spring is that which discharges below the valley floor, where that has been raised by glaciofluvial, etc., deposition. Such aggrada-

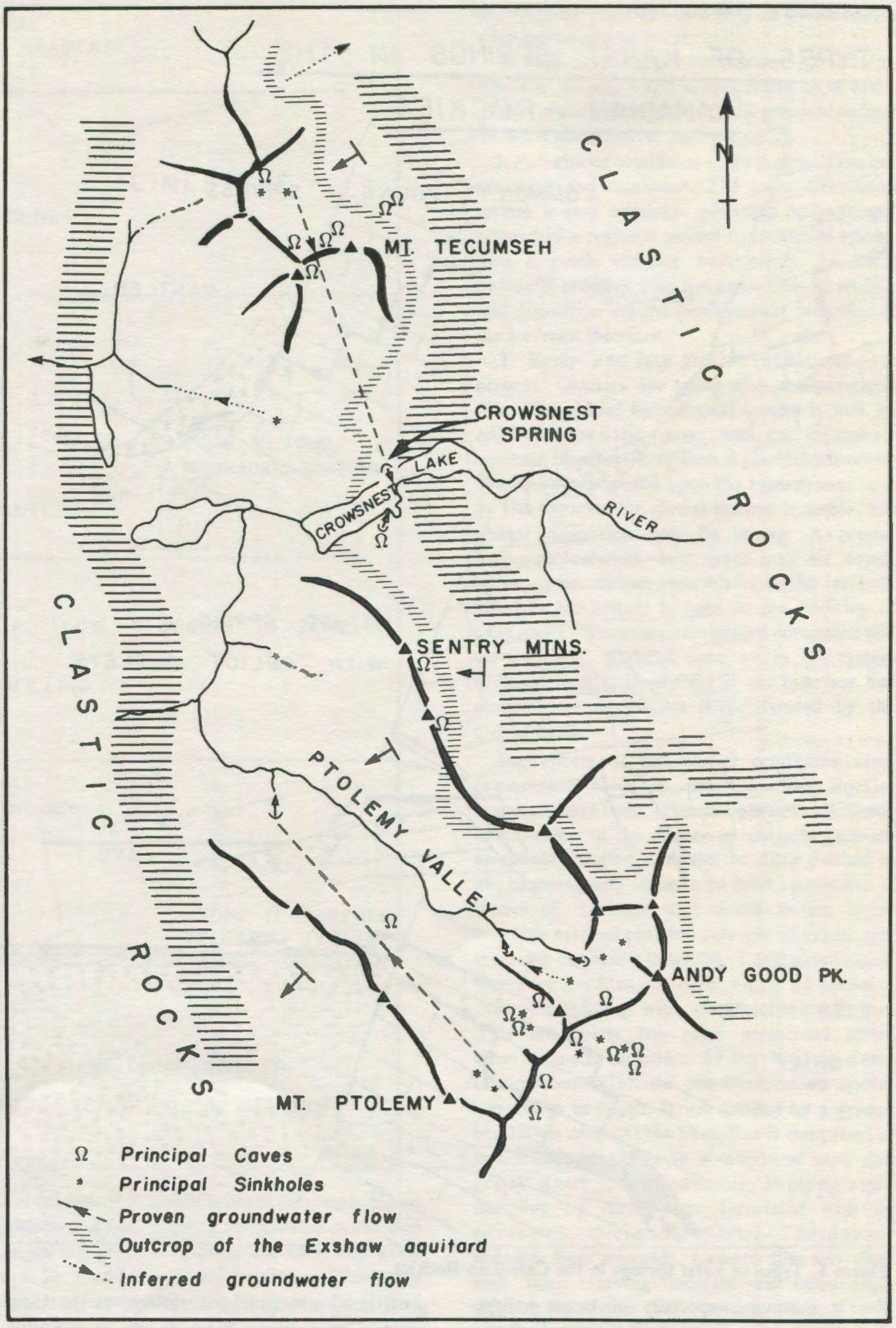

Figure 4. Karst drainage at Crowsnest Pass, Alberta-B.C. Strata dip westerly at 20 to 50°. Crowsnest Lake occupies the Pass and bisects the outcrop of the Exshaw aquiclude rocks, creating four potential zones of karst drainage. In the southwest zone (Ptolemy Valley), drainage is underground and northwest on the strike. Modern drainage in the southeast is largely surficial. In the northwest zone (Mt. Tecumseh), there is underground drainage down the dip, plus drainage to Crowsnest Spring that breaches the aquiclude. The northeast zone displays groundwater drainage against the dip.

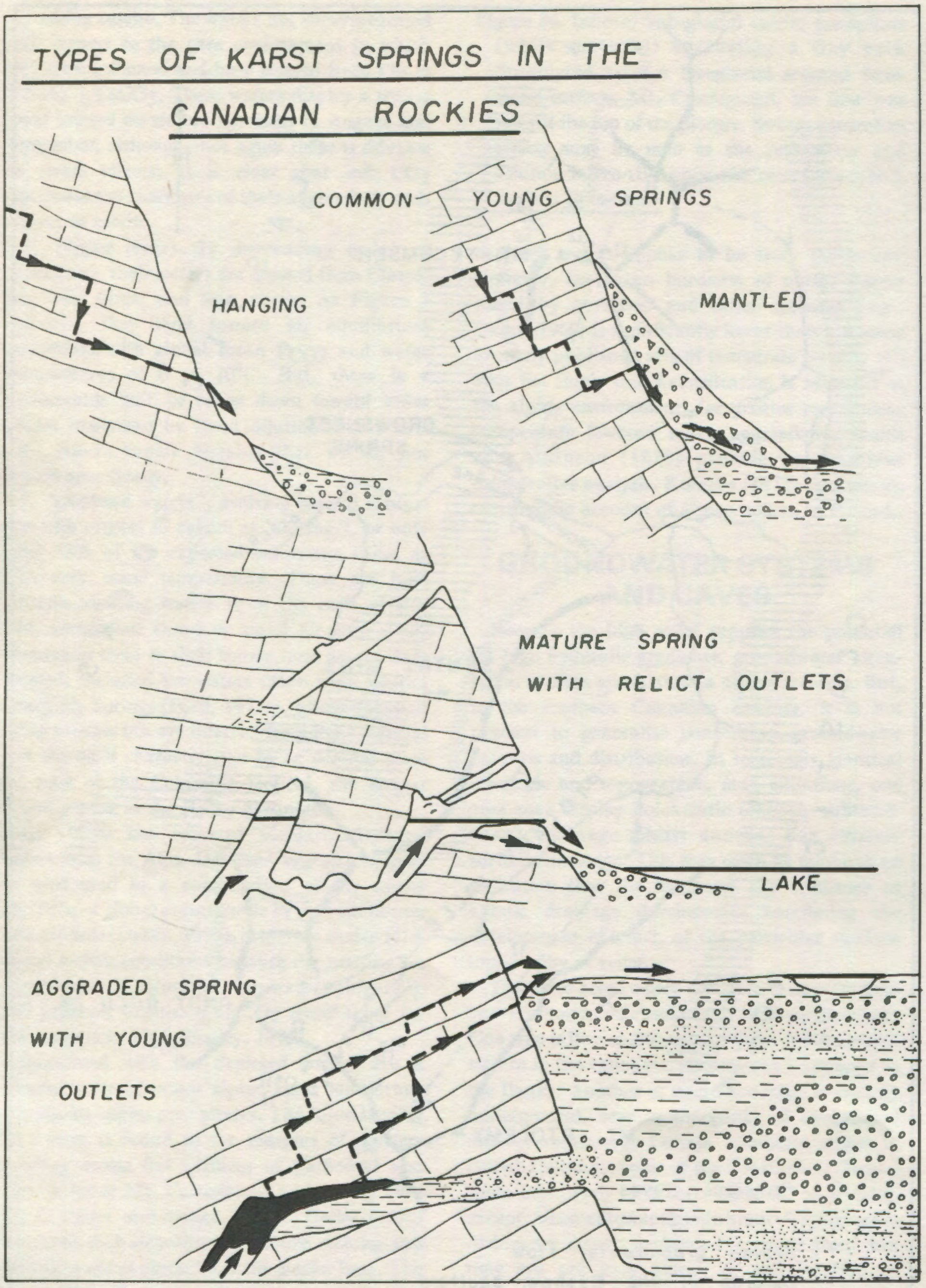

Figure 5. Types of karst springs in the Canadian Rockies.

Figure 6. One of the "Big Springs" in the Cathedral limestone, Castleguard River valley. The spring is young and 'hangs' as a waterfall 40 m above the aggraded valley floor. Waters debouch from a bedding plane 4 m behind the crest of the falls. Discharge has been gauged at > 14 cumecs upon occasion.

tion is common, especially in larger valleys. Probably, a great many small springs discharge directly into valley-gravel aquifers and so have no topographic expression. Larger springs may retain their identity, the water rising in ponds at the bedrock/fill contact. The great example is the springs of the Maligne River (maximum discharge > 60 cumecs), which rise at a canyon mouth and along the adjoining flank of the Athabasca River valley at Jasper (Brown, 1972). They are inferred to discharge upwards through the roof of a great buried cave that now lies over 100 m below the Athabasca.

A concluding point upon springs is that, in a great many extraglacial situations, as the karst matures groundwater conduits will be organised to discharge at the lowest outcrop of the aquifer rock, regardless of controls such as geological structure that might tend to direct it elsewhere and to closer outlets. This was true of the strike-aligned palaeokarst drainage to Crowsnest Pass (Fig. 4): the Pass is the lowest entrenchment into the limestone in that area. However, such sites are rarely exploited by modern karst drainage in the Canadian Rockies. This is an index not so much of youth in the karst as of repeated disordering of systems by glacial action.

Ford (1976) and Ford and Ewers (1978) distinguish six fundamental types of solution cave systems. Four of them are stages in a progression from a deep phreatic loop to an ideal watertable cave; two are vadose. Of these, the deep phreatic types, (Ford and Ewers, 1978, states 1 and 2); appear to be dominant in the southern Canadian Rockies, *i.e.*, caves are sloping phreatic galleries, up or down which water passed. This predominance of deep phreatic over the watertable type of cave is due primarily to steep stratal dips. The nearest thing to a watertable cave yet known is Castleguard Cave, a mixture of state 2 phreatic and "drawdown vadose" system elements (Ford and Ewers, 1978). These observations imply that most of the caves we have explored are relict, drained fragments of phreatic palaeokarst that have been dissected by glacial entrenchment. A few, such as Gargantua (Thompson, 1976) evidence multi-phase (multi-storey) genesis.

The other common type of cave is the 'invasion vadose cave', where comparatively young streams have invaded rock drained by older caves and have opened new routes through it. These are abundant, because each glaciation deranges previous stream inputs, infilling old sinkholes, etc. Many of them are of late glacial or post-glacial age and quickly become impassably small. Abundant is the vertical shaft type of entrance: the truism that "deeper the shaft the less likely you are to find accessible cave at the bottom of it" is nowhere more correct than in alpine regions, where vigorous frost action throws down the walls to infill the floors. The great example is the Sornin Plateau above the Pierre St. Martin system, in the Pyrenees. More than 2000 shafts have been bottomed there: three give access

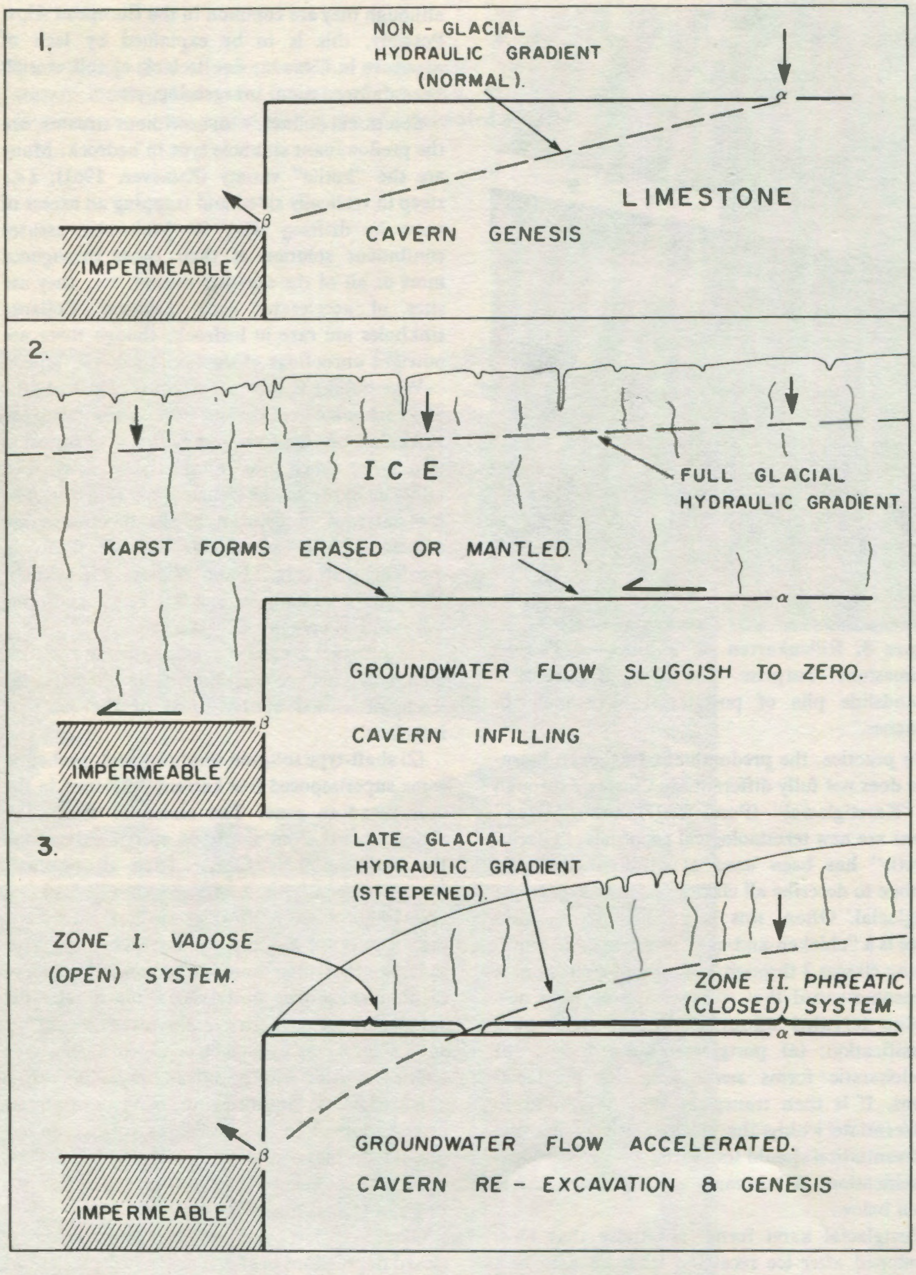

Figure 7. A model of the three differing glacio-hydrological conditions that may occur in temperate alpine karsts. 1. Full interglacial conditions; glacier ice is absent and conventional karst water circulation prevails. 2. Full glacial conditions; glacier ice submerges topographic relief on the karst and karst water circulation is adjunct to glacial water circulation. 3. Early and late glacial conditions; Glacier ice enhances karst relief and glacial water circulation is adjunct to karst circulation. There may be distinct vadose and phreatic zones at the ice/rock interface.

to the cave, the remainder end impenetrable to man. As at Sornin, we keep trying!

KARST GROUNDWATER SYSTEMS AND GLACIAL HYDROLOGY

Relationships between karst groundwater and water circulating in present or past glaciers have received surprisingly little attention in the extensive European literature upon alpine karst. Alpine terrains are the birthplaces of glaciers and the sites where they linger longest during interglacial times. In the southern Canadian Rockies, as in the U.S. cordillera and the European Alps, it may be presumed that glaciers were temperate even during fully glacial climates, *i.e.*, ice temperatures permitted the circulation of water except in an uppermost thin layer that was cold in winter. Ford, *et al.* (1976) give an extended discussion of evidence to this effect from the Castleguard Cave — Columbia Icefield area.

Fig. 7 presents a model of the three different hydrological situations that may be envisioned in a temperate alpine karst:

1. Full interglacial conditions—ice absent. There is 'normal' karst development, as in extra-glacial regions, with an hydraulic gradient defined between sinkpoint '*α*' and spring '*β*'.

2. Full-glacial conditions — ice present. Glaciers are deep and confluent. The karst circulation system is now a minor, probably insignificant, adjunct of a regional glacial hydrological system with a much elevated water-table: the karst hydraulic gradient does not exist. Glacial erosion and deposition are the predominant processes at the ice/rock interface.

3. Early- and late-glacial conditions—ice present. Glaciers are small and comparatively thin. The glacial hydrological system is now an adjunct to the karst system: rock and ice share a common piezometric surface. A glacial hydrostatic head is superimposed upon the karst system, *α* to *β*. The capacity for glacial erosion is feeble, but glacial deposition may be strong. A propos carbonate solution, two zones may be noted: (a) an outer, vadose zone where at the ice/rock interface the system is open to the addition of CO_2 at 0°C, favouring accelerated corrosion; and (b) an inner, phreatic zone where the system is closed to additional CO_2 at the interface but groundwater circulation is accelerated by the glacial head.

The effects of full glacial conditions upon pre-existing caves in the Canadian Rockies appear to have been: (a) minor phreatic solutional modification of the vadose or phreatic bedrock morphology (often it cannot be distinguished at all, implying very sluggish to inert circulation of waters of 'depleted' and closed system type): (b) more marked phreatic solution of calcite speleothems in some cases (*e.g.*, Castleguard Cave; Cave; see Ford, *et al.*, 1976, Fig.7) (c) partial or complete infilling with glacial clastic detritus. This is usually the most prominant effect, there is much variation. At the Nakimu Caves (Thompson, 1976), the pre-Wisconsinan system was 140 m in depth. It was infilled by a graded bed 120 m deep. At the base, this is comprised of boulders and cobbles in a matrix of sand and gravel. There is little structure, implying rapid dumping by fast waters associated with an advancing, thickening glacier. The deposit becomes finer upwards. Layered silts and clays with some varving indicate an increasingly sluggish circulation that eventually became inert, a sub-glacial lake. At Castleguard Cave, there was partial infilling (<15m) with laminated silts and clays including some varves, indicating a sluggish circulation throughout full glacial conditions.

The effects of the early and (especially) late glacial conditions, No. 3, can be radically different. Pre-existing systems, as assumed in Fig. 7, may be subjected to rapid re-excavation plus re-routing of groundwaters, and there is frequent deepening. Alternatively, the system may be partly or wholly bypassed if much infilled, being replaced by new invasion vadose caves or by surface streams. Nakimu Caves are the pre-eminent example of rapid action. At the close of the Wisconsinan glaciation, much of the deep, graded bed there was cleared catastrophically; calculated local water velocities exceeded 6 m/sec in some *horizontal* passages. The hydrodynamic situation is best likened to that of the flushing of a giant lavatory, as diurnal meltwater floods poured into the system via the vadose glacial zone. A series of new galleries were cut beneath the old cave, doubling its depth in the space of a few thousand years.

At the Castleguard karst and elsewhere, there was evidently rapid sub-glacial development of new shaft inlets in the vadose glacial zone. Fresh shafts are found in such hydrologically unlikely places as the crests of roches moutonnees (glacially sculptured bedrock hills), where there is negligible catchment today.

Summarizing, a pre-existing system may be choked or masked and, in effect, erased by glaciation, or its partial infilling may be compensated by periods of accelerated erosion. Which occurs appears to be a function of the ratio:

$$\frac{\text{duration} \times \text{glacial erosion (deposition) power in Condition 2}}{\text{duration} \times \text{hydraulic erosion (corrosion) power in Condition 3}}$$

In the lowland glacial situation that prevailed over most of Canada east of the cordillera, conditions always favoured Condition 2 and karst erasure, because the low bedrock relief provided few situations where receding glaciers would linger at the top of topographic steps. There are great steps and sites for lingering, such as cirques, in alpine terrains. In the Introduction, it was emphasized that glacial inundation was much less in U.S. alpine terrains. This is to imply that Condition 2 hydrology should be less well developed there and Condition 3 proportionally more significant than in Canadian alpine terrains.

TYPES OF SURFACE KARST LANDFORMS

It has been implied here that caves in alpine terrains do not display any erosional features that are uniquely attributable to alpine glaciation, although certain system elements, such as invasion vadose caves, are more common there than in extra-glacial regions. At the surface, it is clear that matters will be more complex because flowing ice impinges directly upon the rock.

In theory, an ideal and comprehensive classification of surface karst forms developed in the context of repeated alpine glaciations would recognise the following fundamental distinctions:-

1. Postglacial karst landforms.
2. Subglacial karst landforms.
3. Karstiglacial forms — karstic adaptations of glacial forms.
4. Glaciokarstic forms — glacial adaptations of karst forms.
5. Mixed forms — karstiglacial + glaciokarstic components combined as a consequence of repeated glacial and interglacial events.
6. Preglacial karst landforms.

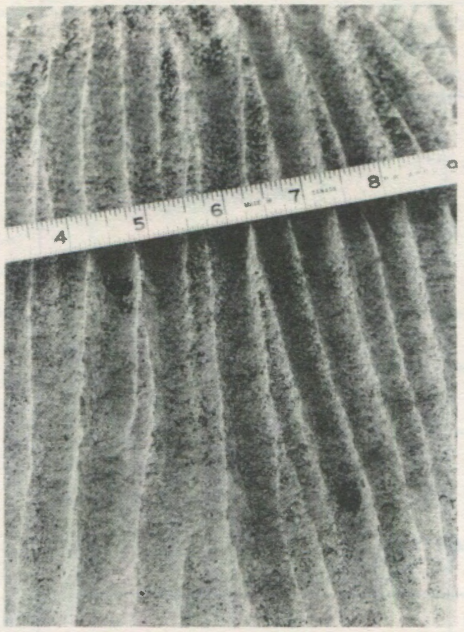

Figure 8. Rillenkarren on a block of Palliser limestone, Surprise. The block is part of a landslide pile of postglacial age. Scale in inches.

In practice, the predominant European literature does not fully differentiate Classes 2 through 5. "Karstiglacial" (Ford, 1977) and "Mixed" forms are new terminological proposals. "Glaciokarstic" has been used at different times in Europe to describe all classes of forms except the preglacial. Often, this is good sense, because there is a "chicken-and-egg" problem in discriminating classes 3 through 5. To avoid confusion, it is recommended that workers approaching new alpine terrains adopt a simple, three-part classification: (a) postglacial karst forms, (b) glaciokarstic forms *sensu lato*, (c) preglacial forms. If it then transpires that it is valid to differentiate within the glaciokarstic group, the differentiation should follow the lines of the ideal classification. A summary of characteristics is given below.

Postglacial karst forms are those that have developed after ice recession from an area and where *the geometrical form itself has no glacial morphologic component. Distribution* of the forms *may* be guided by glacial features. As an example, small solution pits are often aligned along glacial striae on limestone, dolomite, etc. Pit morphology is the same as that of adjacent, non-aligned examples.

In the Canadian Rockies, postglacial karst forms are the most abundant class. As in other alpine karsts (*e.g.*, Rathjens, 1951; Haserodt, 1965; Miotke, 1968) karren and dolines are the principal types. Karren (solution pits, groovings and channels — rarely > 10 metres in greatest dimension) are ubiquitous. 'Free' karren (Bögli, 1961) are the most common. These are developed upon bare rock and have angular rims (*e.g.*, Fig. 8). Few of the more rounded forms associated with development under soil cover are seen, although they are common in the European Alps. Possibly, this is to be explained by lack of exposure in Canada, due to lack of soil erosion from deforestation, overgrazing, etc.

Solutional dolines, with or without streams, are the predominant sinkhole type in bedrock. Many are the "kotlic" variety (Kunaver, 1961); *i.e.*, steep to vertically sided and trapping an excess of snow by drifting (Fig. 9). Meltwater assures continuous solution at their bases throughout most or all of the summer season, *i.e.*, they are sites of accelerated local solution. Collapse sinkholes are rare in bedrock; though there are outright unroofings of the 'karst window' type.

Where there is substantial cover of glacial drift on carbonate rocks, we see some collapse sinkholes, but the form more often is of suffosion (piping) of debris into an underlying cavity, with minor collapse in the debris walls (Fig 10). The few outcrops of gypsum in the mountains are associated with spectacular collapse forms in overlying drift (Fig. 11 and Wigley, *et al.*, 1973).

As noted already, subglacial karst landforms fall into two categories:

(1) subglacial calcite precipitates and associated, tiny pressure solution rills developed in the inner or closed system zone of the ice/rock interface.

(2) shaft-type solution sinkholes and, probably, some superimposed free karren, developed in the outer, vadose zone. The sinkholes cannot be differentiated from shafts of postglacial age on the basis of their forms. Their distribution, however, may show an emphatic lack of association with postglacial surface hydrology; *e.g.*, shafts on the tops of bedrock hills. This indicates that often there will be great difficulty in distinguishing true postglacial from accelerated subglacial karst activity, for development of subglacial forms cannot be confined to sites that lack association with postglacial hydrology.

Karstiglacial landforms are features of glacial origin adapted to karstic drainage; *i.e.*, underground drainage through soluble bedrock. The karst activity may or may not have modified the original glacial form.

Glacial erosion processes scour topographically closed depressions in all rocks. Upon deglaciation, if upon non-karst rocks, these usually fill with water and become lakes or ponds drained at the surface. Canada displays uncounted millions, ranging in scale from < 1 metre in diameter to Great Bear Lake. If upon karst rocks, these will be partly or wholly drained underground where there is a sufficient hydraulic gradient, if the depression is not sealed by deposition of clay grout, etc. There are many examples in the Canadian Rockies. Perhaps the most common is the over-deepened cirque drained underground. I have counted more than 100 examples. Some are drained entirely, water exiting through a postglacial sinkhole in the floor or seeping away through featureless detrital fill. More often, the feature retains a lake, though this does not rise to its rim and overspill (Fig. 12). Sometimes there is a melt season lake, drained dry in winter. There

Figure 9. (right) Small solution doline of the 'kotlič' variety developed in steeply dipping beds of Livingstone limestone, Ptolemy Valley. The lingering snow plug makes this a site of accelerated solution.

Figure 10. (left) A shallow suffosion depression developed in a 5 m mantle of carbonate-rich till resting upon Palliser limestone, Banff National Park.

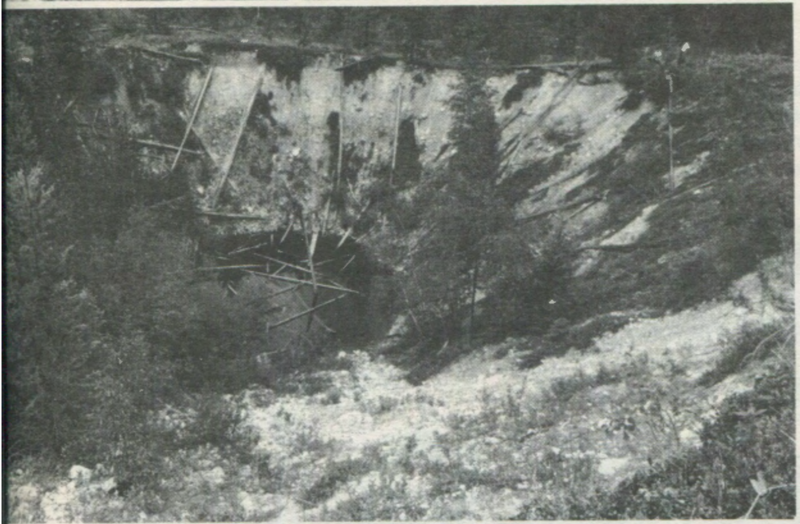

Figure 11. (left) A collapse doline in outwash deposits that bury gypsum to a depth of about 30 m, Canal Flats, B.C. Figures at right for scale.

Figure 12. (right) A large, overdeepened cirque in Rundle limestone, near Nigel Pass, Alberta. The cirque lake (or tarn) is depressed 5 m below the bedrock rim and never overflows. It is drained underground to young karst springs 4km distant. This is an example of a karstiglacial landform.

are also paternoster lakes and moraine-dammed lakes drained karstically in these various ways.

Smaller scours, down to the scale of the smallest true karst doline, may also be adopted. At the 1976 alpine karst symposium in Denver, several speakers from the U.S.A. showed pictures of such features but apparently believed them to be entirely karstic in origin. Care must be taken to distinguish them, or the magnitude of true karst development in an area may be misrepresented.

The vagaries of glacial and proglacial clastic deposition also create innumerable closed depressions of all shapes and sizes. In our context, there are four types:

(1) depressions which, because of high clay content or lack of hydraulic gradient in the drift, cannot drain underground and so fill up as ponds; (2) depressions where the above conditions permit drainage underground through the drift, with no modification of the underground medium; (3) depressions like type 2 but in drift containing a significant quantity of soluble clasts, so that there is subterranean solution and often partial piping and piping phenomena in the depression, plus some collapse; (4) depressions like type 2 or 3, but drained into karst bedrock.

Only the Type 4 feature is a karstiglacial landform. Types 2 and 3 can be described as 'pseudokarstic'. But clearly there will be every difficulty in differentiating between them at most field sites. All 4 types are found on drift-covered limestone in the Canadian Rockies.

Glaciokarstic landforms *sensu stricto* are original karst features modified by subsequent glacial erosion or deposition. Because alpine glaciers are normally efficient scouring agents and karst rocks are comparatively soft, only the larger forms can be expected to survive a full glaciation and then in a quite highly modified form. There is immediately a 'chicken-and-egg' problem in distinguishing them from karstiglacial forms in bedrock, and in many instances the problem cannot be resolved.

The clearest examples of glaciokarstic form in the Rocky Mountains are found in areas of Neoglacial ice recession. The Neoglacial advance began within the last one thousand years, and there has been general recession since 1900 A.D. or thereabouts. Advances were small and the glaciers feeble, so that many karst forms that had grown in their path during preceding postglacial times were able to survive. Two unequivocal examples from the Castleguard Mountain area are showing in figures 13 and 14. The meanderkarren of Figure 13 is a rarity: even Neoglacial ice appears to have been capable of erasing most karren, and I know of none in the mountains that survived full (late Wisconsinan) glaciation. However, they are found in lowland Canada where Wisconsinan glaciers were less abrasive.

Larger sinkholes survive full glacial conditions, and we may guess that many smaller ones exist, concealed beneath drift or indistinguishable from the karstiglacial class. Figure 15 shows a large

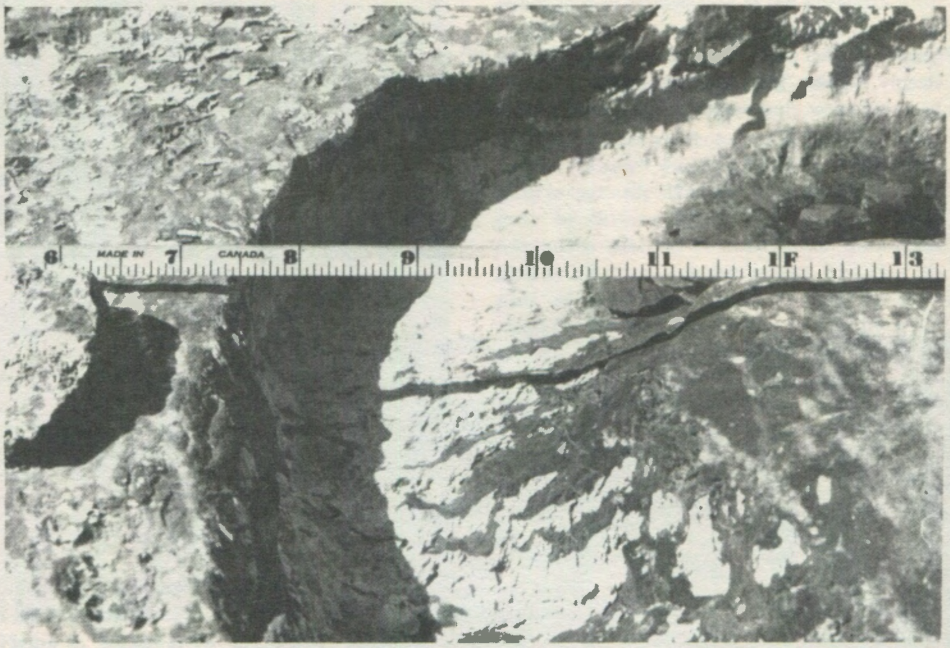

Figure 13. (above) A glaciokarstic landform. This meanderkarren in Eldon limestone was overriden by Neoglacial ice flowing from left to right. The right-hand karren wall is rounded down by ice abrasion. Sub-glacial calcite precipitate is seen in the trough. Scale in inches.

Figure 14. (below) The ideal glaciokarstic landform! This solution doline in Cathedral limestone at Mt. Castleguard was overridden by a Neoglacial glacier flowing from left to right. The upstream (left-hand) doline wall has been plucked and steepened by glacial quarrying. The downstream wall has been rounded down by glacial abrasion and bears an encrustation of sub-glacial precipitate. Connoisseurs will appreciate that this rare landform is the morphologic converse of the common 'roches moutonnees' of glaciated terrains.

Figure 15. (above) Surprise II, a closed depression occuped by a deep lake, a large detrital fan of post-glacial age, and (in foreground) an ablating glacier. It is over 90 m from the lake surface to the depression rim at right. The depression is centered on Exshaw strata, with the top of the Palliser formation to the right and Banff/Rundle rocks composing the cliff to the left. This depression is of glaciokarstic or "mixed" origin. It is drained through the Palliser rocks.

Figure 16. (below) The Medicine Lake depression, sink of Maligne River, Jasper National Park. The depression is viewed from the N.W. (downstream) end, at the close of winter conditions. Waters have receded 15 m below the high water mark indicated by the treeline and sink into the boulders in the foreground. The depression is 6 km in length and 2 km wide. At the close of the Wisconsinan glaciation, it was filled with glaciofluvial deposits that have since been removed through cave channels. It is probably of 'mixed' glacial and karstic origin.

closed depression drained into limestone bedrock where the depth/width ratio appears much greater than the common range for standard glacier scour depressions. The simplest explanation is that it is a big karst feature, adapted by ice during the Wisconsinan. It may be of the 'mixed' class.

The 'chicken and egg' problem is at its worst with mixed forms, where repeated glacial action may have been interspersed with sub-glacial, inter-glacial, and postglacial karst action. Only some very large landforms can be tentatively assigned to the class of mixed forms.

The outstanding Canadian example is the site of Medicine Lake (Figure 16), located in a strike-aligned, hanging glacial valley in Jasper National Park. The lake occupies a closed depression 6 by 2 km in size, with an average depth of 15 m. It overflows for perhaps 10 days once every two years. At other times, the entire river sinks and resurges at Maligne springs, 16 km distant (Brown, 1972). At > 60 cumecs maximum discharge, it is probably the largest resurgence yet measured in an alpine terrain. The lake is reduced to a small, deep pond over a sinkpoint in winter.

The present lake basin is excavated almost entirely in drift, the lost material having been removed via cave channels during postglacial times. The drift, in turn, appears to occupy a closed basin in bedrock. From the form and location, the bedrock basin is unlikely to be a simple glacial scour hole. It is an ice-modified, large karst feature that, in turn, may well have adapted a lesser glacial scour!

Cirques tend to be particularly abundant and well-formed in limestone alpine terrains. Fels (1929) was the first to suggest that many may be glacial adaptations of pre-glacial sinkholes. Given the history of repeated glacials and interglacials in alpine areas, a mixed development is suggested. In extraglacial limestone mountain country of the Pyrenees and Greece, I have seen many large closed karst depressions with assymetric rims that would become ideal prefabricated cirques if glacier ice were able to occupy them.

In the Canadian Rockies, the only preglacial karst landforms that may have survived without specific glacial modification of their morphology are caves. These are mere fragments and at high elevations. Glacial action has dissected the original systems, but, within themselves; the remnants may display no modification attributable to subglacial or glaciofluvial action. However, the systems may be interglacial, rather than preglacial, in age.

ALTITUDINAL ZONATION OF SURFACE KARST

European studies of alpine karst distribution often stress altitudinal zonation. In the Limestone Alps of Austria, Bauer (1961) places a modern upper limit of karstification at about 1800 m asl, the limit of karren development. Higher areas display little or no karst, because frost action is dominant there. In a mountainous region of New

Guinea, Jennings and Bik (1962) have described successively higher zones of doline formation, karren formation, and periglacial action. Bauer sees the effect of the warmest period of postglacial times in Europe (the 'Boreal Optimum' of 6000 to 4000 years ago) in the migration of the karren zone 300 to 400 m above the present limit. These karren are now degraded by frost attack.

In the Rocky Mountains of Canada, the following zones can be recognised:

(1) Boreal forest.

(2) Glaciated tundra — modern tundra that was covered by Wisconsinan glacial ice.

(3) extraglacial tundra — modern tundra not covered by Wisconsinan ice = frost zone.

Almost all of the modern forest zone was also inundated by Wisconsinan ice. Dense growth of forest requires a substantial depth of unconsolidated detritus with soil. The postglacial period has not been long enough for the development of sufficient overburden from decomposition of carbonate rocks left bare by glaciation. Therefore, the forest is restricted to areas of drift cover (glacial, fluvial, or colluvial): the treeline itself is very irregular because of this factor. The karst features are those developed in drift (postglacial and karstiglacial suffosion and collapse dolines), plus springs and caves. Where there is bare rock below the treeline, this displays the same forms as in the glaciated tundra zone, save that the more vigorous growth of lichens may reduce the amount of small karren development.

In the Alps, Pyrenees, and Jura mountains of Europe, Nicod (1972) recognises a specific "nival forest zone" characterised by large bedrock dolines that in some instances are so deep that there is strong temperature inversion and, consequently, tundra in their bases. There is no equivalent zone in the Canadian Rockies. It appears that the European zone has developed in valleys not reached by glaciers (or, at least, by glaciers of the last glaciation). Similar zones may be found in parts of the U.S. cordillera because, as noted, they were not as thoroughly inundated by ice as were the Canadian mountains.

The glaciated tundra is the zone of greatest frequency of surface karst features, especially of the postglacial type. In some localities, one must suspect that this is only because there is no forest to conceal them. But in general, it reflects the fact that the amount and extent of detrital cover is less than in the forest zone. The areas of bare, ice-scoured rock are greater.

There is no altitudinal differentation between dolines and karren in the zone. Distribution and typology of these features reflects local lithology, hydrology, etc. At the Castleguard site, there is some weak evidence of altitudinal migration of karst features, equivalent to Bauer's Boreal Optimum migration in the Alps. Dolines at higher altitude and close to cliffs in the shaly Stephen formation are partly infilled by solifluctual debris, and their bedrock rims are broken down by frost shatter. Those 100 to 200 m lower have not suffered damage to the same extent.

Considerable caution must be exercised before frost damage and degradation of surface karst features is ascribed to some putative climatic deterioration. From many Canadian examples, it is apparent that the contest between frost and karst processes when impinging upon freshly deglaciated rock is not controlled simply by climatic parameters such as temperature. If it is massive in character, the scoured rock may be too 'tight' for quickly effective frost penetration. Although the climate may be one that favours frost features rather than karst, there is an initial period during which aqueous solution opens up the rock, and normal karren and small sinkhole forms develop. Once sufficiently opened, frost shattering becomes predominant and the essential precursor karst forms are degraded or obliterated. Even then, karst groundwater circulation may continue beneath the veneer of shattered rock (Fig. 17). From discussions with J.D. Ives (Institute of Arctic and Alpine Research, Boulder, Colorado), it has become apparent that the extensive surfaces of frost-shattered blocks termed 'felsenmeere' can develop more rapidly on massive limestone than upon massive insoluble rocks. This is because of the 'karst precursor effect' described here. Its results should not be confused with effects of climatic change.

Figure 17. "Frost Pot", a joint-aligned shaft sinkhole within a well developed felsenmeere on Pika rocks, Mt. Castleguard. The surface is in the glaciated tundra zone, but escaped Neoglacial ice scour: a Neological terminal moraine is seen at upper centre of the picture. The shaft becomes impassably narrow at -30m.

Where limestones of the glaciated tundra zone are thin-bedded, pressure release upon deglaciation springs their fissures open and, like other thin-bedded strata, they are attacked rapidly by frost. Karren and other surface karst development may be entirely inhibited.

The extraglacial tundra zone is that exposed to frost action throughout the classical Wisconsinan glaciation, as well as during postglacial times. There has been sufficient time for the obliteration of most precursor karst. Apparently active karst features of karren scale are found here at all altitudes, especially where there is a southerly aspect, but they are few and sporadic.

It would be quite misleading to publish altitudinal boundaries between this and the glaciated tundra zone, except at very local scales. Position of the boundary was determined by ice thickness at the Wisconsinan maximum. Normally the extraglacial zone lies above the glaciated zone, but this is not always so. There is a great deal of interfingering, and the extraglacial zone sometimes reaches down below the treeline.

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ALPINE KARST OF THE SCAPEGOAT-BOB MARSHALL WILDERNESS AND ADJOINING AREAS, NORTH-CENTRAL MONTANA

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ABSTRACT

Alpine karst of the Scapegoat-Bob Marshall Wilderness is found in three main localities: Sawtooth Range thrust belt, Lewis Overthrust, and Silvertip syncline. Karst occurs on ridge tops and high plateaus elevated by thrust faulting or folding and is limited to rocks of Cambrian, Devonian, and Mississippian ages. In general, the flatter the limestone, the more extensive the karst development. Synclines tend to have cave systems located along their axes. Steeply dipping rocks may have moderately well developed surface karst, but are less likely to contain caves. Any soil or vegetative cover on the limestone severely reduces the formation of karst and restricts it to above timberline locations. Cave systems are usually found at the base of the carbonate section and are rarely connected to surface pits.

Both meltwater and springs, where measured, are undersaturated with respect to calcite. Most of the karst and caves is considered to be no older than the last glacial stage.

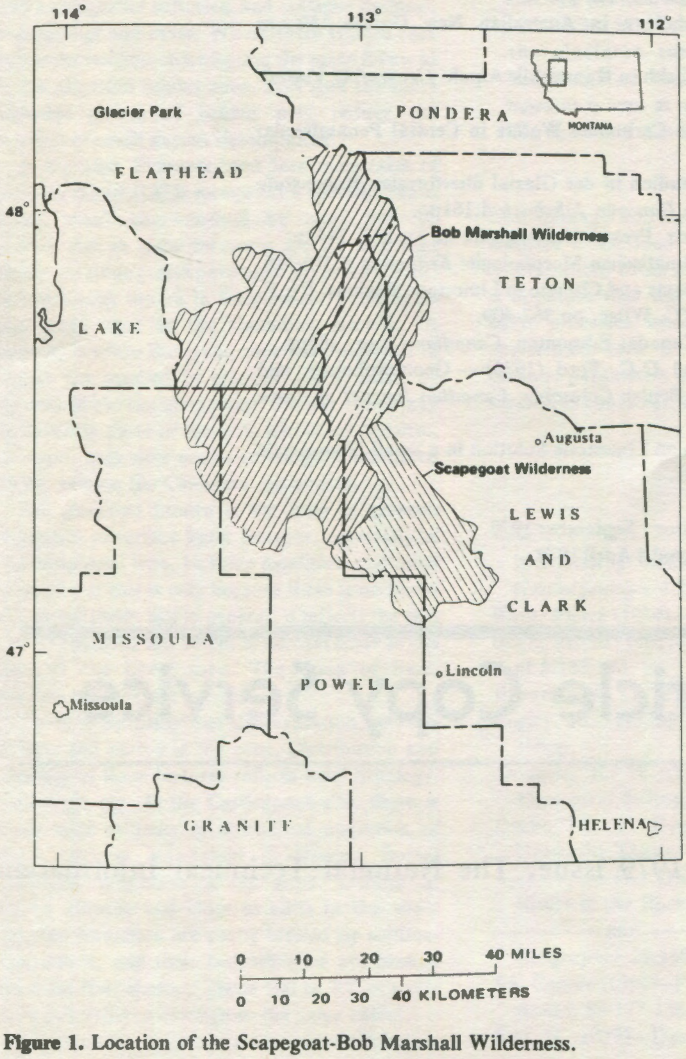

Figure 1. Location of the Scapegoat-Bob Marshall Wilderness.

INTRODUCTION

MONTANA HAS a number of mountain ranges that contain alpine karst. The Little Belt, Pryor, Big Snowy, Beaverhead, and Little Rocky Mountains all have isolated patches of high altitude karst, but the most important area is found in the Lewis and Clark Range and Sawtooth Range in north-central Montana. Most of this karst lies within the Scapegoat-Bob Marshall Wilderness, but some sites occur east of the wilderness boundary.

The Scapegoat-Bob Marshall Wilderness lies about 25 km west of Agusta, 30 km south of Glacier National Park, and 60 km northwest of Missoula. The two adjoining wildernesses encompass an area 120 km by 80 km and are accessible only by horseback or on foot (Fig. 1). Many alpine karst sites are difficult to reach because they are at elevations above 2000 m and as much as 60 km from the nearest road. Some areas, seen from the air, have not been explored on the ground.

The average precipitation in the wilderness is 125 cm per year, derived mainly from winter snowfall. Strong winds create large snowbanks that remain throughout the year. The harsh climate and late snow cover impede exploration and study of the karst.

GEOLOGY

Regional Setting

The Scapegoat-Bob Marshall and surrounding mountains are composed of long, north-south trending ridges and valleys created by folding and thrust faulting. Most of the area is part of the Disturbed Belt, a zone of intense overthrusting along the eastern edge of the Rocky Mountains (Mudge, 1970). The rocks are mainly Precambrian through Mesozoic in age, are sedimentary in origin, and are often aligned so that resistant carbonates form the ridges while softer sediments underlie the valleys. Numerous field studies have been conducted in recent years; for more information see Mudge, *et al.* (1974), Mudge (1970, 1972), and Johns (1970).

Structure

Alpine karst is related to three structural features: thrusts of the Sawtooth Range, the Lewis Overthrust and related folds, and the Silvertip syncline (Fig. 2). The Sawtooth Range thrust belt consists of numerous north-south thrust sheets located along the eastern boundary of the wilderness. Cambrian and Mississippian carbonates have been elevated by thrusting and occupy the tops of the ridges; karst is developed here in long, narrow belts. The host rock changes from Mississippian to Cambrian as one moves from east to west across the range. One large thrust within the belt, the Hoadley (Straight Creek) Thrust, has elevated a large block of Cambrian rock to a height of 2500 m. This 30 km^2 block, known as the Scapegoat Plateau (Fig. 3), is extensively karsted.

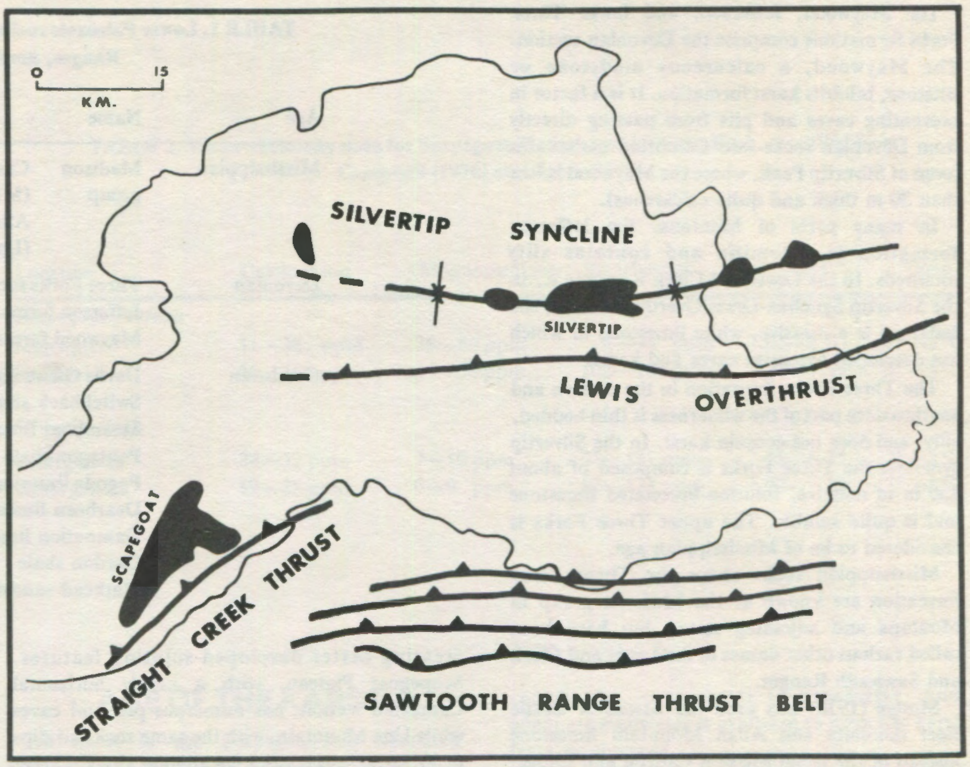

Figure 2. Major structural components that contain karst in the Scapegoat-Bob Marshall Wilderness Important karst areas are blackened.

Figure 3. Part of the Scapegoat Plateau, looking west across Green Fork Cirque. The rock is of Cambrian age. Photo by Roger Knobel.

West of the Sawtooth thrust belt is the Lewis Overthrust, an important and well known fault that has a displacement of as much as 40 km. Displacement decreases to the south, until the fault becomes a large asymmetrical fold belt near the southern edge of the wilderness (Earhart, 1975, personal communication). Important karst areas formed along the Lewis Overthrust are at Pentagon Mountain, the Chinese Wall, and the Flathead Alps. The rocks associated with karst are Cambrian, Devonian, and Mississippian carbonates.

West of the Lewis Overthrust is a long north-south trending fold, known as the Silvertip syncline. The fold is composed of lower Paleozoic rocks that have dips ranging from 30° to 90° along both limbs. Karst is formed in Cambrian rocks near Gunsight Peak and Una Mountain. At Silvertip Peak, a 15 km^2 karst is formed on Devonian and Mississippian rocks that occupy the center of the syncline. the northern part of the Silvertip karst has been studied by Ayers (1976) and by Palmer and Palmer (1978).

Stratigraphy

Rocks which contain karst are limited to Cambrian, Devonian, and Mississippian carbonates. Table 1 shows a stratigraphic section for these rocks.

Cambrian rocks associated with karst and caves are the Damnation, Dearborn, Pagoda, and Steamboat limestones. The Devils Glenn dolomite and underlying Switchback shale are often removed from high areas by erosion. The Pentagon shale crops out only in the Pentagon Peak area and thins rapidly in all directions so that, in many places, a continuous sequence of Cambrian limestones, as much as 400 m thick, is available to the development of caves and karst.

The Maywood, Jefferson, and lower Three Forks formations comprise the Devonian section. The Maywood, a calcareous mudstone or siltstone, inhibits karst formation. It is a factor in preventing caves and pits from passing directly from Devonian rocks into Cambrian carbonates (even at Silvertip Peak, where the Maywood is less than 30 m thick and quite calcareous).

In many parts of Montana, the Jefferson formation is dolomitic and contains silty interbeds. In the Lewis and Clark Range (*e.g.*, in the Silvertip Syncline-Lewis Overthrust area), the Jefferson is a massive, white limestone in which are developed extensive caves and karst.

The Three Forks formation in the eastern and southeastern part of the wilderness is thin-bedded, silty, and does not contain karst. In the Silvertip syncline, the Three Forks is composed of about 150 m of massive, solution-brecciated limestone and is quite soluble. The upper Three Forks is considered to be of Mississippian age.

Mississippian rocks above the Three Forks formation are known as the Madison group in Montana and adjoining states, but have been called various other names in the Lewis and Clark and Sawtooth Ranges.

Mudge (1970) has used the names of Castle Reef dolomite and Allan Mountain limestone instead of the usual Mission Canyon and Lodgepole formations.

TABLE 1. Lower Paleozoic rocks in the Lewis and Clark and Sawtooth Ranges, north central Montana.

Age	Name		Maximum Thickness
Mississippian	Madison group	Castle Reef dolomite (Mission Canyon fm)	540 m
		Allan Mountain limestone (Lodgepole fm)	
Devonian	Three Forks formation		175 m
	Jefferson formation		290 m
	Maywood formation		90 m
Cambrian	Devils Glenn dolomite		
	Switchback shale		
	Steamboat limestone		
	Pentagon shale		
	Pagoda limestone		580 m
	Dearborn limestone		
	Damnation limestone		
	Gordon shale		
	Flathead sandstone		

Distribution

The thicknesses of all carbonates are fairly uniform over the entire area. Erosion and uplift control the location of the karst. Karst surfaces are established on Devonian and Mississippian rocks on the higher peaks and ridges; where erosion has removed them, similar solution features are formed on underlying Cambrian rocks.

FACTORS INFLUENCING KARST FORMATION

A number of alpine karst areas are found scattered throughout the wilderness and, although each is distinctive, many generalizations can be made about the physical nature of the karst:

1. Areas where faulting has steeply tilted the limestone tend to have isolated solution features located only along the tops of ridges and peaks. Tilted carbonates have such rapid runoff that only small scale solution features form. In steeply dipping limestone, few caves are found. In places where the limestone dips to the west, wind blows the slopes clear of snow and little snow pack remains to form caves. The faulted ridges of the Sawtooth Range and parts of the Lewis Overthrust are examples.

2. Limestone that dips 10° or less is more likely to form karst. Water tends to move along vertical joints rather than down bedding planes (Silvertip Peak is an exception—see discussion below), creating better developed solution features. Scapegoat Plateau, with a nearly horizontal Cambrian section, has numerous pits and caves while Una Mountain, with the same rock but dips of 20° to 30°, has only a few shallow sinks. Larger snowbanks tend to accumulate on flat-lying rocks and provide more water for solution. The protected east side of Scapegoat contains many sinks while the west cirque is blown clear of snow and has few solution features.

3. Synclinal areas tend to contain more cave systems than faulted regions. One reason may be that water drains down dip from two directions rather than one. Water then travels along the fold axis, dissolving out cave passages. Silvertip and parts of Scapegoat are examples of this.

4. Surface cover on the limestone severely reduces the development of karst. Not only do thin layers of argillite or mudstone prevent downward movement of water, but soil only a few centimeters thick greatly slows down karst development. This is one reason why significant karst is found only at or above timberline. Patches of grass or trees contain few solution features, while nearby barren rock has many pits and sinks. The increased CO_2 content of the soil apparently does not increase solution enough to overcome the sealing action of the soil.

SURFACE FEATURES

Karsted surfaces in the Lewis and Clark Range and Sawtooth Range contain many kinds of features. Karren, including all of the varieties described by Pluhar and Ford (1970, can be seen wherever massive carbonates are exposed. Thin-bedded limestones tend to be broken by frost action, and the karren are poorly developed or destroyed. Joints control most of the solution on flat surfaces. There is little surface drainage. Water seldom runs more than a few feet before sinking. Joint orientation has been measured at only two locations. Campbell (1973) found major joint trends at N20E, N35W, and N65W on Scapegoat and some minor jointing in a due west direction. At Silvertip Peak, Ayers (1976) found that most joints ran N45W or due north, with some minor jointing at N40E.

Larger surface solution features consist of sinkholes and pits as great as 50 m in diameter and 100 m deep. Some have been plugged by frost wedging, but many become narrow at a shallow depth. All sinks are formed along joints or at joint intersections. The larger ones are always associated with nearby snowbanks (Fig. 4). Snowbanks provide large volumes of water with a high CO_2 content. Few large pits are found on steep slopes and, if water does sink there, it apparently flows along bedding planes without causing much solution. One exception is the Silvertip Peak karst, where cave passages are forming near the surface in steeply dipping, soluble limestone layers and are associated with normal faulting (Palmer and Palmer, 1978). Elsewhere, few cave passages are found near the surface but, instead, occur far below the bottoms of the deepest pits.

CAVE SYSTEMS

Meteoric water often sinks along joints for great vertical distances before encountering insoluble beds. At Scapegoat and other areas, this distance can exceed 300 m. When an insoluble bed is encountered, water flows laterally to the canyon walls, emerging as springs. Insoluble beds can be thin, shaley layers, such as those found in the Damnation limestone, but are more often thick units, such as the Gordon shale, or rock high in magnesium, as is the Devils Glenn dolomite. Thin, insoluble layers tend to be breached by erosion by fast moving water.

Large cave systems usually form at the base of thick carbonates. Underground streams collect

Figure 4. Large snowbank on the Scapegoat Plateau, with associated large pits.

TABLE 2. Water chemistry data for Scapegoat Plateau and Silvertip Peak, modified from Campbell (1976) and Ayers (1976).

Location	Calcium ion	Magnesium ion	Bicarbonate	pH	Temp.
Scapegoat Plateau					
Meltwater	11—38 ppm	26—86 ppm	4—94 ppm	7.07—8.07	1°—2°C
Springs	66—108 ppm	34—58 ppm	105—168 ppm	7.22—7.99	3°—6°C
Silvertip Peak					
Insurgences	24—32 ppm	7—10 ppm	33—141 ppm	7.85—8.25	2°—3°C
Resurgences	19—28 ppm	6—9 ppm	82—123 ppm	7.90—8.20	2°—4°C

water that seeps downward along vertical joints and carry it through the cave system. Cave passages are developed either along joints, in very soluble beds, or along synclinal fold axes. Faulting seems to have no effect on passage orientation.

Cave passages formed along joints are narrow and canyon-like, while passages formed in soluble beds or along synclinal axes are often elliptical tubes. Passages are mostly horizontal, but include occassional overflow bypasses and short pits joining soluble layers. The development of cave levels is probably more the result of downcutting from one soluble layer to another rather than of drops in the water table or changes in surface stream base level. Large cave systems with lengths exceeding 10,000 m and depths approaching 400 m have been found.

The caves contain few speleothems. A few stalactites and stalagmites have been found on Silvertip (Ayers, 1976), and ice occurs in other caves, but the low temperatures (1° — 6°C) retard the formation of speleothems.

WATER CHEMISTRY

There has been little hydrologic study done in the wilderness. Scapegoat waters were analyzed for a single day by Campbell (1976), and Ayers (1976) sampled water on Silvertip Peak over a 26-day period. The data is summarized in Table 2. Both studies showed undersaturation with respect to calcite at both sinks and resurgences. Meltwater was only slightly more unsaturated than springwater. Not enough data is available to generalize water chemistry for the whole area.

AGE OF THE KARST

All karst in the Scapegoat-Bob Marshall Wilderness appears to be quite young. Although Ayers (1976, p. 43) considers part of the Silvertip Peak karst to be older, other areas probably have formed during or since the last glaciation. Glacial meltwater may have initiated formation of the karst, but active solution is continuing now. The largest solution features are found near modern snowbanks, cave water is aggressive, and adequate meltwater is available to form the karst features observed.

Conversely, cave passage densities are low. Surface sinks are usually shallow and are not directly connected with lower cave systems, indicating that solution has not been occurring over a long period of time. The majority of the karst and caves has been formed since the end of the last glaciation.

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SOLUTIONAL LANDFORMS ON CARBONATES OF THE SOUTHERN TETON RANGE, WYOMING

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ABSTRACT

Well-developed karst landforms occur on the western and southern flanks of the Teton Range at two stratigraphic levels: the Mississippian Mission Canyon and Cambrian Death Canyon limestones. Other karst development occurs within the Mississippian Lodgepole Limestone and the Ordovician Bighorn Dolomite. These formations are part of a 3000-ft thick sequence of carbonate rocks which crop out at elevations of 7000 to 10,600 ft. Karsification is controlled by local structural setting, lithology, and nature of recharge to the carbonate aquifers. Karst development differs significantly in four types of settings:

1) ***Structural surfaces.*** *These contain the largest exposures of limestones. Solution at depth occurs only on the periphery of these areas where aggressive runoff water reaches the limestone. Most solution is by uniformly distributed thin films of meteoric water at depths of less than 30 ft, resulting in classical karren forms.*

2) ***Closed basins (cirque-dolines) with limestone dipping toward major peaks.*** *These areas are usually smaller than structural surfaces. Most water is concentrated runoff from other rock types and is highly aggressive when it reaches the limestone. Subsurface piracy occurs through conduits which are usually partly traversable. Subsurface flow paths 2 to 4 mi long with elevation losses up to 2000 ft have been dye-traced.*

3) ***Shelves parallel to westward-flowing streams.*** *These have topographic and solutional characteristics of both the structural surfaces and the closed basins.*

4) ***Cliff faces.*** *These are the setting for some of the larger karst springs in the area. Several of these springs issue from traversable solution channels.*

INTRODUCTION

SEVERAL TYPES of glacio-karst landform assemblages are found in the southern part of the Teton Mountain Range (defined here as being that part of the range south of North Leigh Creek) of northwest Wyoming. We will describe these, the solutional processes which have occurred, the factors which influence the nature of solution both on the surface and in the subsurface, and the factors which influence the patterns of groundwater movement in the carbonates.

Although a wide variety of solutional phenomena exist in the southern Tetons, they have been described in only a few papers (Stellmack, 1968; Medville and Werner, 1974; Hill, Sutherland, and Tierney, 1976). In this paper, the work which we have carried out during the past ten years as well as that of others which is known to us will be summarized.

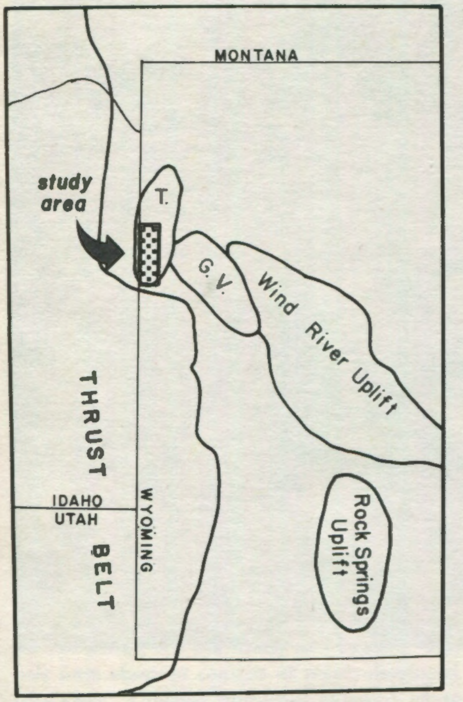

Figure 1. Index map. T—Teton Mountain uplift, G. V.—Gros Ventre uplift.

DESCRIPTION OF THE AREA

The Teton Range is about 15 mi wide and 30 mi long and trends north-south. It is located near the western boundary of Wyoming, at the eastern edge of the Idaho-Wyoming Thrust Belt (Fig. 1). Structurally, the range is a large fault block uplifted and tilted to the west and broken by a number of secondary faults. Descriptions of the geology of the range can be found in Horberg (1938), Edmund (1951), and Love and Reed (1968). While the highest peaks are composed of Precambrian gneisses and schists, the southern and western slopes of the range consist of an uplifted, west-dipping sequence of sedimentary rocks ranging in age from Cambrian to Triassic; this sequence is over 3000-ft thick in this area. About 2500 ft of Cambrian-to Mississippian-age carbonates are included. They crop out over about 120 mi² at elevations of 7000 to 10,000 ft. The rocks generally dip to the west to northwest at 5 to 15°.

The stratigraphic column and a brief lithologic description are given in Fig. 2, and a generalized geologic map is shown in Fig. 3. Karst features and evidence of solutional activity are found at the top of the Mission Canyon limestone, the top and base of the Lodgepole limestone, the top of the Bighorn dolomite, and throughout the Death Canyon limestone. The major landform types on which these features can be found fall into four categories: open structural surfaces, semi-open shelves, closed basins, and cliff faces. The nature of the solutional activity found differs for each of these landform types and depends on lithology, degree to which runoff is concentrated, quantity of runoff, and landform morphology. Groundwater movement in the western Teton carbonates follow a pattern in which water sinks at or near the top of a soluble rock, drops vertically via enlarged joints to an impermeable unit, flows downdip for distances up to 6 mi, and then resurges either from talus, from springs drowned by colluvial deposits, or from enterable openings. Upon reaching the next lower carbonate unit, this process may be repeated until either the base of the lowest limestone or base level drainage is reached, whichever is higher.

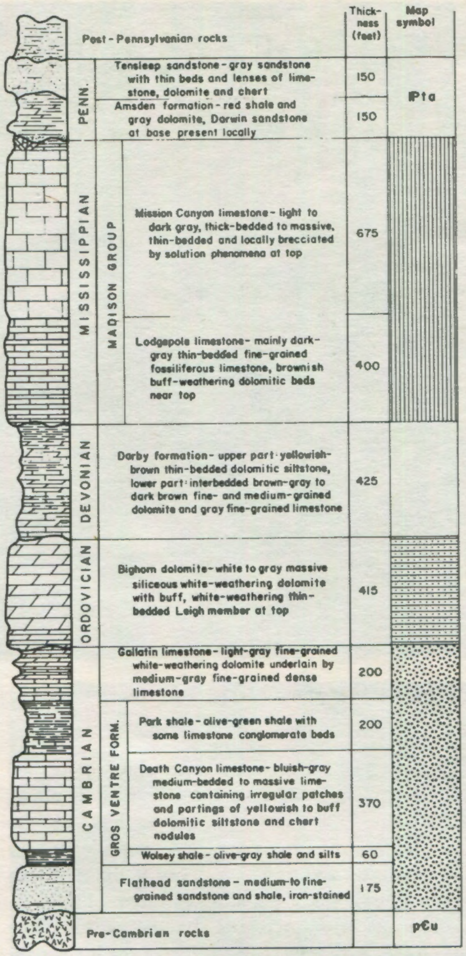

Figure 2. Generalized stratigraphic column of the western Tetons. Data adapted from Edmund (1951) and Horberg (1938). Map patterns used on Figure 3 are shown.

SOLUTIONAL DEVELOPMENT IN VARIOUS SETTINGS

Open Surfaces

Open surfaces, referred to by Edmund (1951) as structural plains, are large, open areas where the slope of the surface coincides with the dip of the rocks. Although they exist at several stratigraphic horizons, open surfaces are most commonly found at the tops of the Lodgepole Limestone, the Bighorn Dolomite, and the Death Canyon Limestone. They are the most areally extensive landforms on which solution takes place and occupy more than 15 mi². The characteristics which distinguish them from the others described in this paper are:

flatness — The surfaces can best be thought of as tilted planes. Small, local variations in relief are caused by stream channels; these are rarely more than 15 ft below the general elevation of the surface.

openness — While individual closed depressions may be found on the surface, the landform as a whole trends downdip and is itself not a closed depression.

lack of recharge — The upper ends of the surfaces are generally at elevations of 9000 to 10,000 ft and receive little recharge from higher elevations. Thus, only small quantities of water enter from above via surface streams. Most of the water found on these surfaces is derived from precipitation and melting snow.

As a result of these characteristics, solution on the open surfaces tends to be diffuse rather than concentrated. A wide variety of karren are found

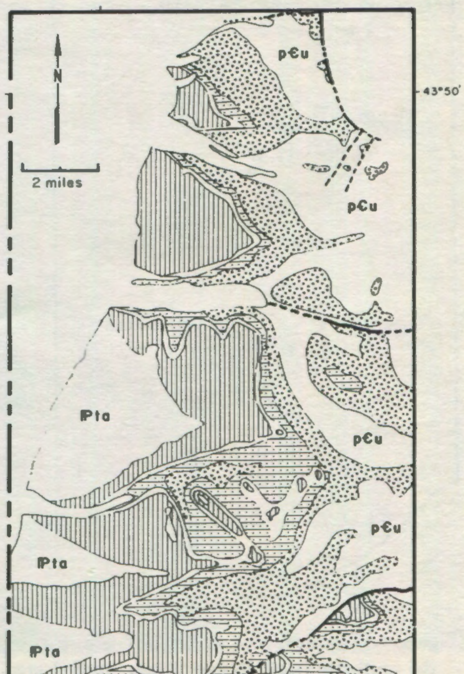

Figure 3. Generalized geologic map of the southern Teton Range. See Figure 2 for key to symbols.

both on the limestone and on the dolomite. Where solution takes place at depths greater than a few feet, it occurs along narrow joints, all of which are blocked by broken rock and surface fill. Although these joints are numerous, no caves of significant horizontal extent have been found on any of the surfaces.

Four representative examples of open surfaces at different stratigraphic horizons are shown in Figure 4:

The Terrace. The Terrace is an open surface at the head of Terrace Creek, a southern branch of

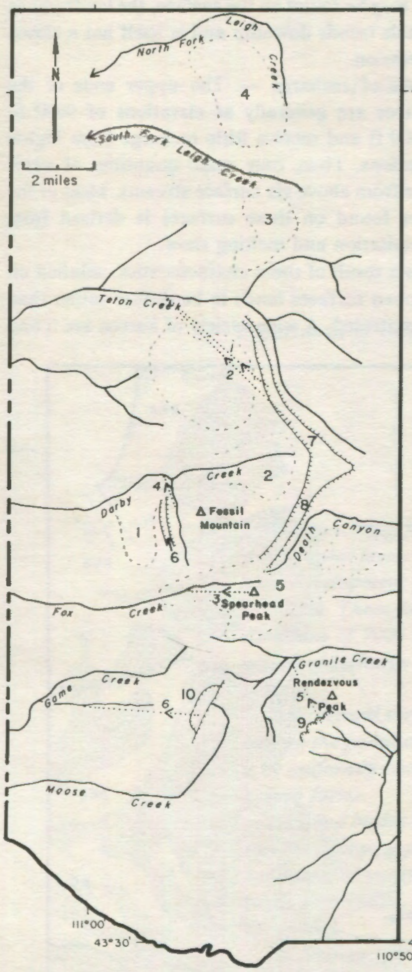

Figure 4. Physiographic and hydrologic map of the southern Teton Range. Areas studied are: 1—The Terrace, 2—Upper Darby Canyon, 3—Beard's Wheat Field (Death Canyon) surface, 4—Leigh Creek (Death Canyon) surface, 5—Spearhead Peak karst, 6—South Darby Creek Shelf, 7—Teton Shelf, 8—Death Canyon Shelf, 9—Rendezvous Peak basins, 10—Moose Lake basin. Hachured lines indicate cliffs; hachures are on the down side. Subsurface dye traces are shown as dotted lines with arrows; numbers correspond to entries in Table 1. Note: Although trace 6 (Moose Lake to Game Creek) is shown rising in the South Fork of Game Creek, this may not be the case. The detector was placed below the junction and thus the rise could be in either fork.

Figure 5. Air photo of The Terrace.

Darby Creek. Developed near the top of the Mississippian Lodgepole Limestone, it is about one mile in length and width (Fig. 5). Its southern, western, and eastern boundaries are ridges rising 200 to 400 ft above it. Its northern (downdip) boundary is at an elevation of 9000 ft, where the slope increases and the canyon floor drops more rapidly than the dip of the limestone. Topographic variation about the mean surface level is ± 20 ft. The Terrace is developed subparallel to the direction of the dip and trends N45°W. The average slope of this surface is the same as the dip component of its major axis — about 12°. The Terrace varies in elevation from 9000 to 9800 ft and, because of the limited area above it, recharge is negligible.

Although solution takes place on all exposed rock surfaces, solution at depth occurs only at the periphery of The Terrace, especially at its southern (updip) margin along the surrounding ridges where snow drifts against and is protected by the ridges above. Here, several enlarged strike joints and vadose shafts are found; the largest of these is over 230 ft deep.

Solution in the center of The Terrace consists of enlarged joints which typically extend downward for 10 to 20 ft. At this level, horizontal solution along bedding-planes results in circular, saucer-shaped chambers up to 30 ft across. These are up to 6 ft high in the center, but decrease to a few inches in height around the periphery.

Karren, notably maänder- and spitzkarren, are in evidence throughout The Terrace. Dolines and other closed depressions follow joints, are up to 50 ft in diameter and 20 ft in depth, and are usually clogged with frost-shattered limestone.

The water which sinks into joints on The Terrace has not been traced; however, probable outlets are either springs in Terrace Creek, 1.5 mi north and 1200 ft lower at the base of the Lodgepole Limestone, or the large resurgence at Wind Cave in South Darby Canyon, 6000 ft northeast and 1000 ft lower.

Upper Darby Canyon. One of the most areally extensive (4 mi^2) open surfaces in the study area is in upper Darby Canyon, at the top of the Ordovician Bighorn Dolomite. Here, the canyon broadens to over 1.5 mi in width. As with the other open surfaces, this one is relatively flat, has no deeply incised streams, and has little recharge from above. It reaches elevations of over 9800 ft at its updip end. The slope of the surface trends downdip to the northwest at about 10° until, as at The Terrace, the slope exceeds the dip. Downcutting into the dolomite begins at about 9000 ft elevation.

Although the Bighorn Dolomite and its upper Leigh Member are pure dolomites and are substantially less soluble than the Madison Group limestones, solution occurs both on the surface as spitzkarren, maänderkarren, and trittkarren and at some depth as solutionally enlarged joints (Fig. 6). These enlarged joints extend across the width of the dolomite pavement, follow the strike, and are up to 1000 ft in length. Where local solution occurs, they reach widths of up to 15 ft and depths of over 50 ft. In no case can they be followed to the base of the dolomite, because they usually become too narrow for human passage. They appear to form in the same manner as the enlarged joints, especially at the outer margins of the area, and vadose movement of snowmelt waters gradually enlarges them.

Water sinking into these joints appears to move downdip within or at the base of the dolomite and rises at several springs about 2 mi downcanyon.

These springs are generally found in talus at the base of the Bighorn Dolomite cliffs and discharge up to 1 cfs each.

The Death Canyon surfaces. Open surfaces are also found at the top of the Cambrian Death Canyon limestone — the lowest carbonate rock in the western Tetons. At least three such surfaces exist: one between Teton Canyon and South Leigh Canyon, a second between South Leigh and North Leigh canyons, and the third near Spearhead Peak at the headwaters of Fox Creek. The first two are each about 1 mi^2 and the third is over 2 mi^2 in area. The Leigh Canyon surfaces were first described by Edmund (1951). These surfaces, or structural plains, are coincident with the dip of the limestone and have the same characteristics as the ones described above; *i.e.*, they have little local relief, no recharge from higher elevations, are well-jointed, and have primarily distributed rather than concentrated solution. Because of the thinner bedding of the Death Canyon Limestone and the existence of siltstone partings within it, its weathering differs from that of the more massively bedded Mississippian limestones, and surface karren are not nearly as extensive. The only form observed consists of poorly-developed trittkarren. Although numerous enlarged joints are found at the updip ends of these surfaces, there is little horizontal cave development. The largest solution feature is found on the surface between North Leigh and South Leigh canyons, at an elevation of 9100 ft. Clearly visible on aerial photographs, this is a large, vertically-walled collapse sink over 80 ft in diameter and about 50 ft deep.

Spearhead Peak karst. As shown in Figure 7, this area is located at the head of Fox Creek and is the highest and largest example of this type of landform developed on the Death Canyon Limestone. This surface of more than 2 mi^2 is between 9400 and 10,200 ft in elevation. It is, basically, a tilted plateau. The full thickness of the Death Canyon Limestone is exposed as cliffs on its northwestern, eastern, and southern margins. As within the other surfaces, groundwater movement is via deep joints and downdip to the northwest. All water sinking on the Spearhead Peak karst rises at two springs of about 5 cfs each (August 1976) found at the base of the limestone cliff at the northwest (downdip) end of the surface.

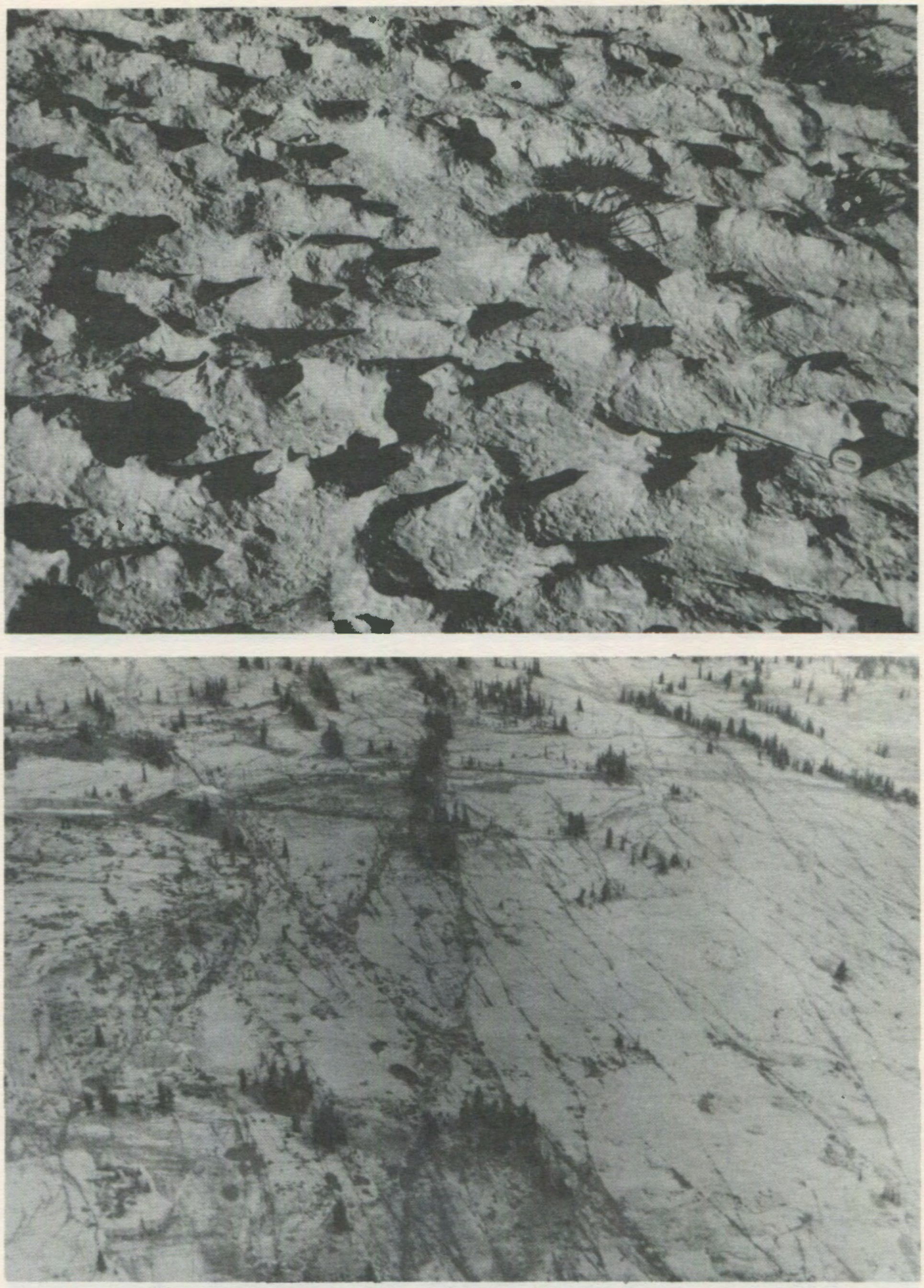

Figure 6. Two typical types of karren forms developed on the Bighorn Dolomite in Darby Canyon.

Shelves

Shelves paralleling the canyons of the western Tetons are found in a variety of structural and stratigraphic settings. These landforms can be described as elongated surfaces bounded on one side by a lower escarpment composed of the same rock on which the shelf is developed and on the other side by an upper escarpment composed of younger rock units. The shelves found on carbonate rocks in the study area vary from 2 to 4 mi long, are up to 1/3 mile wide (although average widths vary from 500 to 1000 ft), and are found on or near the tops of the Mission Canyon Limestone, the Bighorn Dolomite, and the Death Canyon Limestone. Their major distinguishing characteristics are:

flatness — Like the open surfaces, these are basically tilted planes following either one or several closely-spaced stratigraphic horizons. The slope of the shelf surface is always parallel or subparallel to the dip direction.

The shelf surfaces consist, in part, of a stepped bedding plane karst or schichttrippenkarst (Bögli, 1964) on the west-dipping carbonates.

recharge — Streams flow onto the shelves from adjacent higher areas. As a result, solution is both more concentrated and more localized than on the open surfaces.

semi-openness — In all cases, either the downdip direction or a sizeable component of this direction is toward (into) the upper escarpment bordering the shelf. Consequently, streams entering the shelf must cross the dip and downcut stratigraphically in order to reach the outer margin. Although the shelf is not itself a closed depression, the dip direction with respect to the bounding cliffs impedes the flow of surface streams across it.

As a result of these characteristics, solutional activity on the shelves tends to be more

Figure 7. Spearhead Peak karst.

Figure 8. Entrance to Wind Cave, South Darby Canyon shelf above.

concentrated and at greater depth than on the open surfaces. Karst features here include joints and vadose shafts up to 180 ft deep, and substantial sub-surface drainage. Three examples of such shelves are (see Fig. 4 for locations):

The South Darby Canyon shelf. Located at the top of the Mission Canyon Limestone, this shelf parallels and is about 600 ft above the south fork of Darby Canyon (Fig. 8). The shelf is 6000 ft long, 600 to 1000 ft wide, and varies from 9000 to 9800 ft in elevation. Local relief is moderate (±50 ft). Streambeds up to 300 ft long and small canyons up to 20 ft deep are incised in the shelf floor. Intermittent streams in these small canyons are derived from snowmelt and meteoric waters. The direction of flow is downdip to the northwest and toward the upper cliffs which bound the shelf on the west. With the exception of one stream which does cross the shelf, all ultimately sink.

Surface karren on the shelf are extensively developed; the most prominent forms are maänder- and spitzkarren. Much of the surface of the shelf consists of a bare, well-jointed limestone pavement with two prominent joint sets trending downdip (about 45°W) and along the strike. These joints are up to 20 ft deep, exhibit vadose fluting along their walls, and at their bottoms are clogged with frost-shattered rock. The largest single solution feature on the shelf is a vertically-walled collapse sink, 120 ft in diameter, 80 ft deep, and floored with angular limestone blocks. Neither bedrock on the floor nor evidence of current solutional activity can be seen in this sink.

The Teton Shelf. This is the most areally extensive shelf in the western Tetons — over 4 mi long and about 1/2 mi wide (Fig. 9). Formed at the top of the Death Canyon Limestone, it parallels the west side of upper Teton Canyon and is 1200 to 1400 ft above Teton Creek. The elevation of this shelf ranges from 9600 ft at its southeast (updip) end to 9400 ft at its northwest end, where it narrows to 500 ft and becomes buried beneath talus from the overlying Bighorn Dolomite.

This shelf and the solution features on it have been described previously (Medville and Werner, 1974). The surface follows the dip of the Death Canyon Limestone (10° at N68°W). Dip joints trend N28°E and N58°E. The shelf, itself, trends N32°W. Thus, water flowing onto it and following the dip will flow toward the inner margin of the shelf, away from Teton Canyon. Most of the solutional features found on the shelf consist of enlarged joints up to 180 ft deep. Several large dolines up to 300 ft wide and 50 ft deep are found at upper end of the shelf.

Two perennial and several intermittent streams flow on the Teton Shelf. All wholly or partially sink into joints in the Death Canyon Limestone. Water on the shelf is derived from snowfields at the bases of Bighorn Dolomite and Madison Group cliffs at the western edge of the shelf. Total flow onto the shelf in early August 1975 was about 1 cfs (less the 0.1 cfs that flowed across the shelf into Teton Creek below).

The northernmost perennial stream on the shelf flows northwest for 3000 ft along the inner (downdip) margin of the shelf and then sinks into a small cave. This stream has been traced to a large alluviated spring (31.25 cfs in August 1976) at the top of the Death Canyon Limestone in Eddington Canyon (see Fig. 10). The spring, at an elevation of 6900 ft, is a tributary of Teton Creek and is about 3.4 mi northwest of and 1650 ft lower than the insurgence on the shelf. The flowthrough time from sink to rise was less than 72 hours. It appears that water sinking on the shelf almost immediately drops 300 ft through the vertically-jointed Death Canyon Limestone and then flows downdip to the northwest, through either joint- or bedding-parting-controlled conduits, until the top of the limestone passes below the local base level for drainage, which, in this case, is Teton Creek. The water then resurges.

The difference between the amount of water sinking on the Teton Shelf and the amount rising at the Eddington Canyon spring indicates that this spring may be a resurgence both for water sinking into the Death Canyon Limestone and for water sinking higher in the 3000-ft carbonate sequence found in the western Tetons, particularly at the top of the Bighorn Dolomite and at the top of the Madison Group. This water has not yet been traced.

A second perennial stream on the Teton Shelf, 2 mi south of the one described above, is also derived from snow melting at the base of higher cliffs. This stream, like the first, flows northwest along the downdip side of the shelf for 2 mi. It then turns to the northeast, crosses the shelf, and passes through a 20 to 30-ft-deep canyon as it flows across the dip of the limestone. Between June and September, 80 to 100% of the water in this stream sinks into joints in the limestone. The flow in August 1973 was 0.5 cfs with the entire stream sinking. This water has also been traced to the Eddington Canyon spring, 4.4 mi northwest of and 2200 ft below the sink. The flowthrough time was under 5 days.

In addition to the two perennial streams described, several smaller, intermittent streams flow for short distances on the shelf and sink into enlarged joints.

The Death Canyon Shelf. This shelf, also developed on the Death Canyon Limestone, is on the north side of Death Canyon and 800 ft above the canyon floor. It is 3 mi long and 300 to 500 ft wide. On its western end, this shelf is adjacent to the Teton Shelf; on its eastern end, it joins the Spearhead Peak surface. Unlike the Teton Shelf, its long axis parallels the strike of the limestone and its elevation is about 9400 ft throughout. The inner margin of the shelf terminates beneath talus from a 300-ft-high Bighorn Dolomite escarpment. There is minimal direct recharge onto this shelf from above, and almost all water found on it is from precipitation and snowmelt. Although several small caves and pits are found on the shelf, none of these are over 50 ft deep.

At the base of the Death Canyon Limestone cliff on which the shelf is formed are several springs discharging over 1 cfs each. These springs are at the easternmost (updip) end of the limestone in contrast to other carbonate springs in the western Tetons, which are found at the downdip margins of the limestone/dolomite exposures which they drain.

The water rising at these springs has not been traced, but probably is derived from water sinking on the Death Canyon shelf, the outer Teton Shelf to the northeast, and possibly the closed basins at the top of the Bighorn Dolomite cliffs, 1000 ft directly above the springs.

Closed Basins

Large, closed basins are found in the southern and western Tetons, where northwest-dipping carbonate rocks occur on the east sides of higher peaks and ridges. The basins appear to have originated as a result of cirque glaciation with subsequent solutional modification to their present configurations. Although, in a few cases, surface water currently entering the basins flows against the dip and breaches the updip side of the basin, the usual pattern is one of streams sinking in the basin floor. The basins, or cirque-dolines, in the study area are up to a mile wide. They are found on the Bighorn Dolomite and on the top of the Death Canyon Limestone. Their primary characteristics are:

non-openness — The basins are generally semi-circular in shape. Their eastern boundaries lie at the updip margins of the carbonate rock, their western boundaries at talus and cliffs of higher rock units, and their northern and southern margins at ridges composed of higher rock units. As noted above, streams entering the basins sink.

recharge — As with shelves, streams from higher elevations flow into the basins. The resulting solution of the carbonates is locally concentrated.

solutional concentration — Although about half of the water found on the shelves and almost

Figure 9. (above) Teton Shelf.

Figure 10. (below) The large spring in Eddington Canyon. Measured flow was 31.25 cfs when the picture was taken.

Figure 11. One of the Rendezvous Peak basins. Note sinking stream near center of picture.

all of the water on the open surfaces is meteoric, almost all of the water in the basins is runoff from higher elevations.

There are two areas in the southern Tetons which contain closed basins (see Fig. 4 for locations):

The Rendezvous Mountain basins. Rendezvous Mountain is on the east flank of the southern Tetons. On the east side of the mountain, five basins or cirque-dolines are formed on northwest-dipping Death Canyon Limestone. These basins are at elevations of 9400 to 9600 ft and are 1000 to 1600 ft wide (Fig. 11). As on the Teton Shelf, streams entering the basins are fed by snowfields at the bases of higher cliffs to the west. The basins have no surface outlets; all water sinks into enlarged joints in the limestone. The limestone dips 15° to the northwest. Two major joint sets are found in the basins: 1) strike joints at N10°E and 2) dip joints. As on the Teton Shelf, the floors of the basins consist, in part, of an inclined pavement with solution taking place along joints. The pavement in the basins is much less prominent than that of the Teton Shelf, however.

Caves have been found in the Rendezvous Mountain basins, at or near points where streams entering these basin sink. Surface runoff is more concentrated here than on the shelves, and, perhaps for this reason, cave development is greater. Two significant caves have been found. At the lowest point in the first basin south of the Jackson Hole Aerial Tram, a small (0.05 cfs) surface stream sinks into a rock-choked, 20-ft-deep enlarged strike joint. A cave is developed along this same joint 100 ft to the north. The entrace is 2 ft wide and leads to a series of shafts totalling 170 ft, vertically. At the base of the shaft series is a horizontal passage developed along a bedding parting. A stream flows northeast along the strike for 800 ft to a rock choke. Water in the stream is, apparently, derived from the surface stream sinking 100 ft south of the entrance.

The second basin to the south also contains a small stream sinking into a cave. The entrace of this cave is at 9600 ft, at the top of the Death Canyon Limestone. The 5-ft-high entrance passage trends downdip for 300 ft and is developed along a bedding parting. This passage leads to a series of shafts, developed along vertical joints, totaling 250 ft and carrying water to the base of the limestone. Below these joints, a lower bedding-parting passage is intersected. This passage averages 20 ft high and wide and contains a stream flowing at an estimated 0.5 cfs. This lower stream is probably derived from streams sinking in basins to the south of the one containing the entrance, although it has not been dye-traced. The lower passage also trends downdip to the northwest and has been followed to a depth of 512 ft below the cave entrance.

Fluorescein dye placed in the lower stream in the second cave described above was traced to a spring (flow about 6 cfs) on the south fork of Granite Creek, 2 mi northwest of and 1300 ft lower than the cave entrance. This spring is on the west side of Rendezvous Mountain and emerges from Quaternary colluvium below cliffs of Bighorn Dolomite. Flowthrough time was under 24 hr. Streams sinking in the other Rendezvous Mountain basins have not been traced, although it is expected that they will also rise at the spring on the south fork of Granite Creek. Because the total amount of water sinking into all of the eastern basins on Rendezvous Mountain is about 1 cfs, it is apparent that this spring is the resurgence for a considerably larger area than is accounted for by these basins. The recharge area for the spring probably includes the headwaters of the south fork of Granite Creek, which sinks into the Madison Group 1.5 mi upstream of the spring. The drainage area of the spring probably also includes basins higher on Rendezvous Mountain which are developed primarily on the Bighorn Dolomite and the Gallatin Limestone.

Directly above the Rendezvous basins described above are smaller closed basins on the Bighorn Dolomite. These upper basins are at elevations of 9800 to 10,200 ft and are about 100 ft across. Although an inclined pavement is found on the top of the Leigh Member and vertical jointing along the strike is pronounced, cave development

Figure 12. Moose Lake basin.

similar to that found in the Death Canyon Limestone is absent.

The Moose Lake basin. On the west side of the Open Canyon Fault and on the east side of the ridge south of Housetop Mountain, the Death Canyon Limestone is exposed for a distance of about 2 mi. Here, at an elevation of about 9400 ft, the largest closed basin in the Teton Range is found (Fig. 12). This area, the Moose Lake basin, is a generally treeless, inclined surface 0.9 mi^2 in area at the top of the limestone. The limestone is well-jointed and dips to the west at 10 to 20°.

The Moose Lake basin contains three ponds. The northernmost of these, Moose Lake, is about 500 ft in diameter. The outlet stream from this lake flows for a few hundred feet and then, during low flow, sinks into rock-filled, enlarged joints in the limestone. Just beyond this point is a small cave in the streambed which takes the discharge from the lake in higher flow conditions. One thousand feet south of Moose Lake and in the same basin are two smaller, unnamed ponds. The outlet streams from both of these join, and the combined discharge enters a 20-ft-deep pit developed along a strike joint. Subsurface flow can then be followed along the strike for about 100 feet. The streams entering this cave have been traced with fluorescein dye to a spring rising in one of the forks of Game Creek, 2.5 mi to the west. Thus, as with the Rendezvous basin streams, initial subsurface flow is downward through vertical joints to or nearly to the base of the limestone and then directly downdip for a substantial distance: 2.1 mi horizontally and over 1400 ft vertically. Although it has not been traced, it is assumed that the Moose Lake water also rises on Game Creek. It should be noted that the Death Canyon Limestone is not exposed in the Game Creek valley; the youngest bedrock exposures there are of Bighorn Dolomite. Consequently, it appears that, at some point in its subsurface flow, the Moose Lake basin water must rise stratigraphically at least to the top of the Death Canyon Limestone. This is also the case with the Rendezvous basin water, which rises just below the Bighorn Dolomite cliffs, and with streams sinking on the Teton Shelf, which rise at the Eddington Canyon spring at the top of the limestone.

Cliff Faces

Carbonate rock escarpments 200 to 400 ft high are commonly found in the western Tetons. These cliffs contain a variety of solutional features, including enlarged joints, natural bridges, frost pockets, and cave entrances. Although the three landform types described above all provide settings where surface streams sink, the escarpments are the major landform type where subsurface water resurges. In a few cases, this water rises at enterable caves. The most notable example of this (and the only example in the Madison Group limestones) occurs at Wind Cave in South Darby Canyon.

Wind Cave is located at the base of the cliff of Mission Canyon Limestone which terminates the South Darby Canyon shelf on its outer (canyon) side. The entrance, near the downdip end of the cliff, is the largest in the study area — 30 ft wide at the base and over 150 ft high (Fig. 8). A stream having an estimated summer flow (July, 1976) of 1/2 cfs rises from broken rock just inside the entrance and flows out. This water has been traced from Fossil Mt. Ice Cave, one mile up canyon. The cave has a surveyed length of 900 ft, is developed along a joint sub-parallel to the strike, and extends beneath the shelf above. A prominent joint is visible in the cave ceiling. Domes occasionally extend upwards toward the shelf. The cave appears to have been developed under phreatic conditions, as indicated by the oval-shaped cross-section of the upper part of the passage. This apparently was followed by entrenching of the cave stream into the base of the tube. A deeply incised vadose canyon, extending to the base of the Mission Canyon Limestone, resulted.

Although several other caves are found in the same cliff as Wind Cave and are are at about the same stratigraphic horizon, none of these contain streams. It would appear that Wind Cave is the outlet for water sinking both on the shelf above and, possibly, on The Terrace as well.

Resurgence caves are also found at the base of the Death Canyon Limestone cliffs below the Spearhead Peak karst and at the base of the cliff beneath another open surface on the Death Canyon Limestone between Battleship Mountain and Hurricane Pass. A majority of the caves found in the cliff faces, however, are dry, abandoned outlets. The most common other solutional features found in the cliff faces are frost pockets, in which voids are created mainly by spalling away of rock through action of freezing water or chemical weathering, and solutionally enlarged joints extending into the cliff for a few feet. Neither of these features are of hydrological significance.

SUMMARY OF STREAM TRACING RESULTS

Several stream traces using fluorescein dye have been referred to in this paper. The major characteristics of these traces are summarized in Table 1 and shown in Figure 4. Stream traces have been conducted on all three of the landform types on which streams sink. Several things should be noted about Table 1. First, almost all of the traces were conducted in the Death Canyon Limestone. As the stratigraphically lowest carbonate rock in the study area, surface flow onto it is most often sufficient for traces to be carried out. Water sinking into Madison Group limestones is more difficult to trace, because of the lack of surface streams at the higher elevations where the Madison is exposed. Second, in all cases, subsurface flow is more or less downdip to the west; substantial flow paths along the strike or against the dip have not been found. Third, two cases of interbasin transfer of water, on a modest scale, are known. One occurs in the Rendezvous Peak basins, where surface water that would otherwise flow directly into the Snake River at the base of the east side of the Teton Range sinks, flows west beneath Rendezvous Mountain, and rises in Granite Canyon (which is tributary to the Snake River). Interbasin transfer also occurs in the Moose Lake basin, where surface water that would otherwise be part of the headwaters of Moose Creek sinks, flows west beneath a southern spur of Table Mountain, and rises in the Game Creek drainage.

The stream traces listed in Table 1 represent only a partial picture of subsurface flow paths in the southern and western Teton carbonates. Additional traces should be carried out to determine whether or not water sinking into units such as the Lodgepole Limestone and Bighorn Dolomite crosses clastic beds in the Darby and Upper Gros Ventre formations and then resurges from the Death Canyon Limestone. There is some evidence that this occurs in the Death Canyon Shelf area and in a large exposure of the Madison Group above the Eddington Canyon spring. In one trace carried out, the stream which flows out of the entrance of Wind Cave and sinks into its bed in upper Bighorn Dolomite a few hundred feet downstream was used as the injection point. This water rises at an alluviated spring in the floor of Darby Canyon, 1200 ft lower. Although bedrock is not seen at the spring, a small exposure of Death Canyon Limestone is found within a few hundred feet, at about the same elevation as the spring. This exposure and the vertical drop from sink to rise (greater than the projected downdip elevation of the Bighorn Dolomite) indicate that subsurface downcutting through the clastics and into lower carbonates can occur.

WATER GEOCHEMISTRY

In this section, we will expand on the discussion in an earlier paper (Medville and Werner, 1974) of 13 samples taken in 1973 from Death Canyon Limestone terranes. An additional 15 samples were collected during field work in 1974 and 1976, during August both years. These newer samples, like the previous ones and like those cited in the hydrogeology studies, have the limitation of being primarily related to Death Canyon Limestone terranes, although a few samples were obtained from other, higher horizons.

Samples were analyzed for pH, temperature, specific conductance (at the time of sampling or shortly thereafter), bicarbonate (in the field shortly after sampling), calcium, magnesium, nitrate, sulfate, and chloride. Because of the conditions of sampling, transportation, and analysis, several of the values obtained were less accurate than normal; calcium and magnesium concentrations are estimated to be ±2%, pH is ±0.15 units, and bicarbonate is within ±5%. The other constituents are usually too dilute to be of concern in this interpretation. The conclusions should be regarded as being accurate only to the extent allowed by the data.

Generally, these waters have a relatively low hardness compared to that of many temperate zone karst waters (*e.g.*, Harmon, *et al.*, 1972; Schuster and White, 1971; Ogden, 1976); however, they are well within the range of figures given for arctic or alpine terrains (Bögli, 1960; Corbel, 1960; Ford, 1971) as well as of those for karst stream and spring waters with low residence times from West Virginia (Werner, 1974). Most of the waters have alkaline earth metal hardnesses below 100 ppm (calcium carbonate equivalent). In general, although the slowest moving water has the highest hardness, pond waters seem to form a special case. Summaries of the water data appear in Table 2.

The new data largely confirm the conclusions of the earlier paper; however, some of the reasoning on which those conclusions were based needs re-examining. The rough classification previously presented still appears to be valid. Based on carbonate dissolution products (*i.e.*, calcium, magnesium, and bicarbonate), this classification places the water samples into pond waters, fast-moving waters, slow-moving surface water, and slow-moving cave water (Fig. 13). Since no further samples of slow-moving cave water have been obtained, the earlier comment that the last two groups may be one still stands. There is no overlap between groups on the basis of alkaline earth hardness (the most reliable analysis).

The waters containing the least carbonate dissolution products are pond waters, except for a single sample. This presents a problem in that, for some ponds at least, the source of the pond water is streams which fall into the fast-moving water classification. The explanation for the particularly low levels of hardness advanced in the earlier paper (Medville and Werner, 1974, p. 101), low carbon dioxide levels in snowmelt, is not entirely valid, as seen from our new data.

In those cases where samples were obtained from both a snowmelt stream and the pond into which the stream flows, the hardness is higher in the stream than in the pond, although the equilibrium partial pressure for carbon dioxide is usually less in the snowmelt stream. Thus, it appears that some of the dissolved material is removed after the water reaches the pond—possibly by repeated warming of the water with the resultant outgassing and reabsorption of carbon dioxide and resulting lowered solubility of the carbonate minerals at intervals, or by simple removal of these substances by plants in the ponds. Certainly the problem requires further investigation, as the data currently available are insufficient for an adequate explanation.

The additional data obtained from the samples of fast moving waters show that relatively little carbonate solution occurs underground. Paired samples, from sinks and rises of the same systems, indicate no differences, within the limits of analytical errors indicated above, of calcium, magnesium, or bicarbonate concentrations. Where sufficient information was collected to compute calcite saturation levels (2 sets of samples), the saturation levels differ only slightly between sinks and springs. The preponderance of solution appears to occur in the first few hundred feet of flow over the carbonate. Solution underground is clearly an extremely small fraction of the total solution occurring in the terranes investigated in the southern and western Tetons.

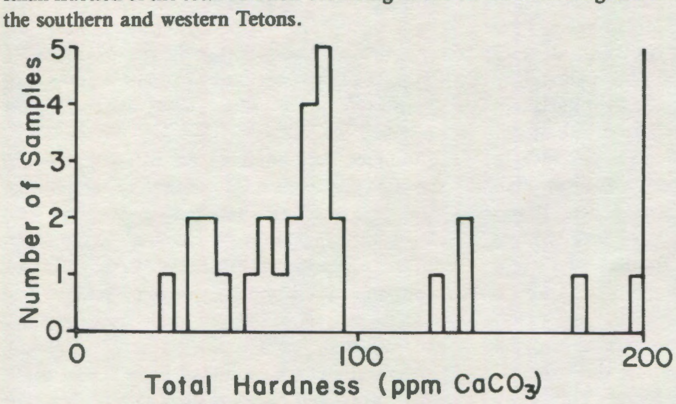

Figure 13. Distribution of measured values of total alkaline earth hardness for water samples collected in the southern Tetons.

TABLE 1. Summary of Stream Trace Results in the Southern and Western Teton Range.

SINK	RISE	DISTANCE (miles)	VERTICAL DROP (ft)	GRADIENT (ft/mi)	FLOW DIRECTION	COMMENTS
1. Teton Shelf above Devils Stairs — top of Death Canyon limestone	Eddington Canyon Spring at top of Death Canyon limestone	3.4	1650	485	N35°W	Drops to base of Death Canyon limestone, flows downdip, and then rises to top of Death Canyon limestone
2. Teton Shelf — outer sinking stream — top of Death Canyon limestone	Eddington Canyon Spring at top of Death Canyon limestone	4.4	2220	505	N45°W	same as in 1. above
3. Spearhead Peak karst — top of Death Canyon limestone	Base of Death Canyon limestone cliff at lower end of Spearhead Peak surface	1.0	400	400	N80°W	Drops to base of Death Canyon limestone, flows downdip, and then rises in talus
4. Wind Cave stream in South Darby Canyon — top of Bighorn dolomite	Spring in Darby Canyon at top of Death Canyon limestone(?)	1.3	1280	960	N14°W	Drops to base of Bighorn dolomite, through Darby and Upper Gros Ventre formations and, possibly, into the Death Canyon limestone; underdrains South Darby Canyon
5. Rendezvous Peak basins — top of Death Canyon limestone	South Fork Granite Canyon in colluvium or talus	1.7	1260	740	N20°W	Drops to base of Death Canyon limestone, flows downdip, and rises to spring, possibly at top of Death Canyon limestone
6. Moose Lake basin — top of Death Canyon limestone	Game Creek — top of Death Canyon limestone	2.1	1440	685	N75°W	Drops to base of Death Canyon limestone, flows downdip, then rises, probably, in upper Death Canyon limestone

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TABLE 2. Selected values from chemical analyses of water samples.

	SAMPLE STATION	Date (mo/yr)	pH	Ca (mg/l)	Mg (mg/l)	TH	HCO_3 (mg/l)
1.	Devilish Cave entrance	viii. 73	7.0	23.2	7.0	87	nd
2.	Snowfire Pit entrance	viii. 73	7.5	36.0	9.2	128	nd
3.	Lake insurgence	viii. 73	7.8	20.4	10.7	95	nd
4.	Diabolical Cave, entrance	viii. 73	7.5	16.0	7.3	70	nd
5.	Eddington Canyon spring	viii. 73	7.7	23.2	7.8	90	nd
6.	Rendezvous Cave entrance	viii. 73	7.2	20.0	9.5	89	nd
7.	RP basin 1, upper pond	ix. 73	9.0	12.4	2.7	42	nd
8.	RP basin 3, pond	ix. 73	8.3	12.0	3.6	45	nd
9.	Rendezvous Cave entrance	ix. 73	8.1	22.0	8.0	88	nd
10.	RP Cave, top of drop	ix. 73	8.1	54.1	15.1	197	nd
11.	Stream into RP cave	ix. 73	7.9	20.0	8.0	83	nd
12.	RP basin 1, small stream	ix. 73	8.3	38.0	10.2	137	nd
13.	South Fork, Granite Creek	ix. 73	7.6	25.0	7.5	93	nd
14.	RP basin 1, spring at pond	viii. 74	8.0	23.1	2.1	66	58
15.	RP basin 1, lower pond	viii. 74	7.8	15.1	2.5	48	70
16.	RP basin 1, lower pond	viii. 74	7.8	15.7	3.3	53	75
17.	RP basin 1, upper end of stream	viii. 74	7.6	11.9	1.4	35	61
18.	RP basin 1, lower end of stream	viii. 74	8.1	28.8	2.5	82	73
19.	RP basin 3, pond	viii. 74	7.8	13.0	3.7	48	71
20.	Spring below Fossil Mountain	viii. 76	7.3	23.4	4.5	77	102
21.	Fox Creek, spring 1	viii. 76	7.4	22.3	5.4	78	88
22.	Fox Creek, spring 2	viii. 76	7.5	19.7	3.8	65	71
23.	Darby Canyon spring	viii. 76	7.5	21.1	5.4	75	87
24.	Eddington Canyon spring	viii. 76	7.4	26.1	5.1	86	102
25.	Teton Shelf, pond	viii. 76	7.6	23.1	6.4	84	99
26.	Devilish Cave entrance	viii. 76	7.7	25.0	4.5	81	97
27.	North Fork Coal Creek, spring	viii. 76	7.6	40.0	9.3	138	163
28.	South Fork Coal Creek, spring	viii. 76	8.1	46.5	14.6	176	206

Notes: RP — Rendezvous Peak
TH — total alkaline earth hardness in mg/l $CaCO_3$
nd — value not determined

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KARST LANDFORMS IN THE WASATCH AND UINTA MOUNTAINS, UTAH

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ABSTRACT

A widespread but subdued karst has developed on the Mississippian carbonate rocks of the Wasatch Front Range and in the Uinta Mountains of northeastern Utah. Surface landforms in the Wasatch Mountains include scattered areas of limestone pavement, a remnant pinnacle karst, and small caves. Underground drainage is more common in the Uinta Range. The caves are somewhat larger and are frequently related to sinking streams. Caves in the Wasatch are mainly controlled by faults and fracture zones and have little relation to the local drainage. Alpine glaciation has obliterated some karst forms and buried others in morainic material. High dolomite content of the Mississippian carbonate rocks is a strong controlling influence and results in subdued landforms. There is some evidence for post-karst tectonic movements.

INTRODUCTION

A WIDESPREAD but rather subdued karst is developed on the Mississippian carbonate rocks that crop out in the Wasatch Mountains and along the flanks of the Uinta Mountains in northeastern Utah. Large expanses of bare carbonate rock occur with some solutional sculpturing. Caves are found in both mountain ranges, and in the eastern Uintas are Big Brush Creek and Little Brush Creek caves, the two largest caves in Utah.

Few formal descriptions of karst landforms from this region appear in the literature, except for remarks made in the course of other geological investigations. The landforms in American Fork Canyon were described briefly by White and Van Gundy (1974). Caves of both the Wasatch and Uinta ranges have been explored by members of the Salt Lake Grotto of the National Speleological Society, and their findings have been published in a series of Technical Notes. Timpanogos Cave (White and Van Gundy, 1974) Big Brush Creek Cave (Palmer, 1975) and Neff Canyon Cave (Green and Halliday, 1958) have been described in some detail. Karst hydrological investigations have been made in conjunction with water resources studies, particularly in the Brush Creek and Ashley Creek basins north of Vernal (Maxwell, *et al.*, 1971).

This paper presents a general survey of the surface and underground landforms. Field investigations were restricted to the Brush and Ashley Creek basins in the eastern Uintas, the Soapstone Basin in the western Uintas, and the portion of the Wasatch between Provo and Little Cottonwood canyons. The first field work in the area was in 1966; the most recent in the summer of 1976.

GEOLOGIC SETTING

The Wasatch Mountains are a north-south trending fault block range generally considered to be the western-most range of the Rockies. The range is about 175 miles long, with the southern end fading into the Colorado Plateau south of Nemo, Utah and the northern end just south of the Idaho border. The western margin of the range is a prominent fault scarp that forms an abrupt boundary between the mountains and the Salt Lake Valley to the west. The Wasatch Mountains have been extensively block-faulted, and tectonic activity has continued through the Pleistocene. Some of the terraces of Lake Bonneville are cut by fault scarps. The Wasatch Mountains are, therefore, very rugged with craggy peaks, many vertical cliffs, and deep canyons. The highest peak is Mount Timpanogos, at 14,250 ft.

The Uinta Mountains are an east-west trending range in northeastern Utah. They extend from the Colorado line to meet the Wasatch at nearly right angles between Salt Lake City and Provo. Between the ranges is the broad expanse of the Heber Valley, at an elevation of 7000 ft. The dominant structural element in the Uinta Mountains is a broad anticline extending the length of the range, with parallel faults along the north and south margins. The mountains have a domal shape, and, although the elevations of individual peaks are greater than 12,000 ft, the landscape is broad and rolling and distinctly less "alpine" than the Wasatch.

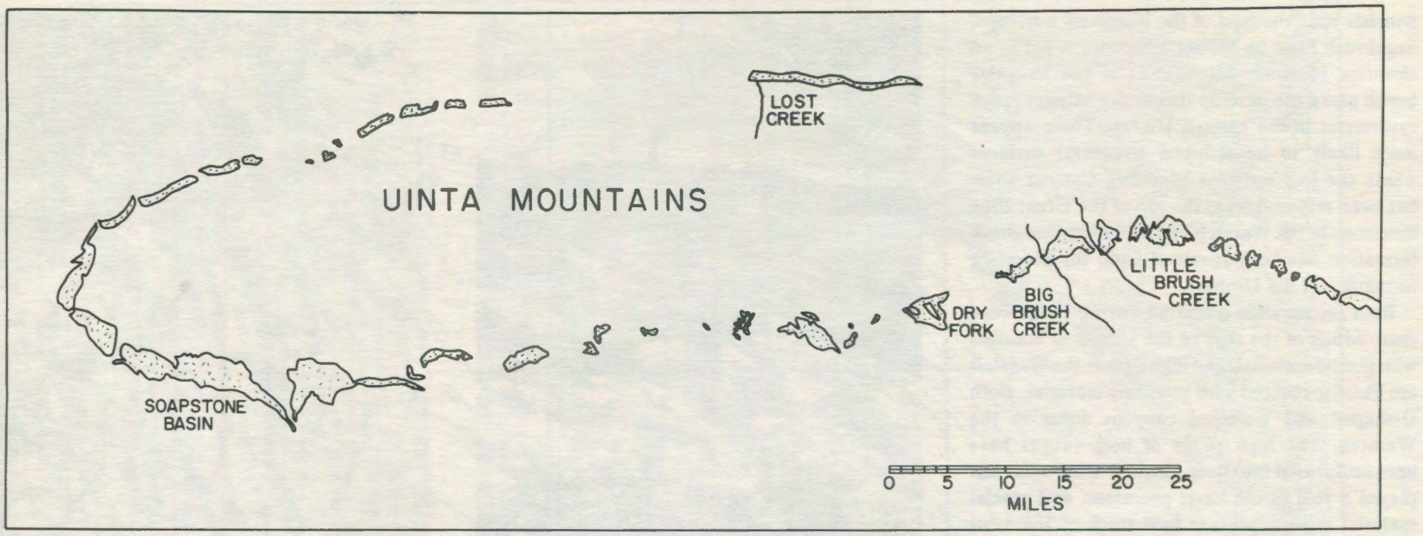


Figure 1. (above) Outcrop pattern of Mississippian carbonate rocks in the Uinta Mountains.

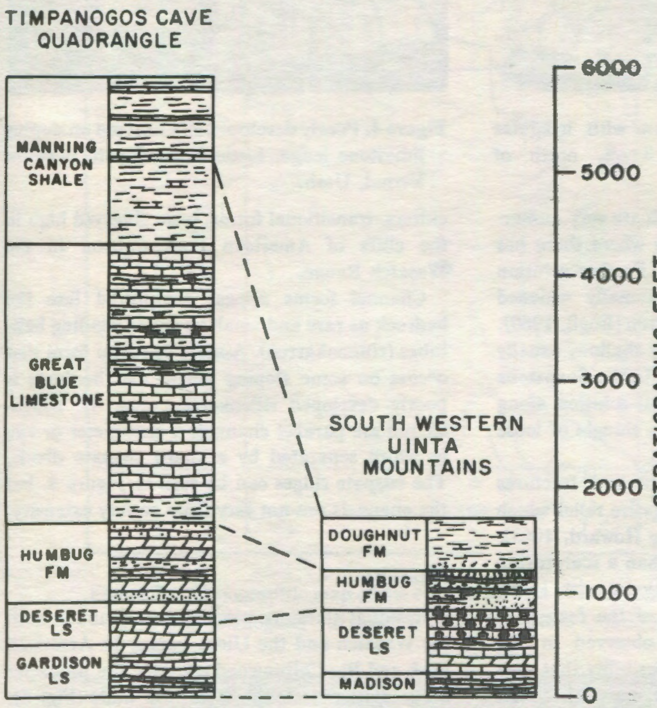


Figure 2. (left) Somewhat generalized stratigraphic sections for the thick sequence and the thin sequence of Mississippian rocks. Adapted from Crittenden (1959) and Baker and Crittenden (1961).

The geology of the Wasatch Mountains is complex because of the faulting. Much of the range is underlain by a mixed carbonate rock and clastic sequence, although intrusives and metamorphic rocks occur as well. It is not possible to generalize the Wasatch geology, and detailed studies of the karst landforms must depend on such individual quadrangle geology maps as may be available.

The central core of the Uinta Mountains is underlain by pre-Cambrian metaquartzites of the Uinta Mountain Group. Paleozoic sediments are stacked along the flanks of the range, where they have been uplifted by Uinta Mountain folding (generally agreed to be of Paleocene to Eocene age [Ritzma, 1969]). The outcrop area of the Mississippian carbonate rocks forms a well-defined band around the perimeter of the range (Fig. 1).

The stratigraphy of the sedimentary rocks is shown in Fig. 2. The Mississippian rocks in the area west of the Wasatch fault and in some parts of the Wasatch Mountains are composed of a very thick section of limestones and dolomites with interbedded shales and sandstones. The eastern boundary of the thick section is demarcated by a series of overthrust faults (see Crittenden, 1959, for discussion). North and east of the Charleston overthrust, the thickness of the Mississippian section is about one third that of the thick section. The northern part of the Wasatch and all of the Uinta mountains are in the region of thin section. However, the rocks in which most of the caves and karst features are developed, the Gardison (Madison) and Deseret limestones, have roughly the same thickness and lithologic character in both areas.

None of these carbonate rocks are very pure either lithologically or stratigraphically. Both Madison group rocks and the Deseret are mainly limestones with interbedded dolomite and some dolomite concentration throughout the section. Portions of the section contain bedded chert.

The geomorphology of the Wasatch and Uinta mountains has been discussed by Bradley (1935), Threet (1959), and Marsell (1969). Early descriptions, couched in Davissian terms, identified four erosion surfaces, of which the Gilbert Peak surface was highest, reaching 13,000 ft along the crest line of the Uintas. The Bear Mountain surface was said to be 400 to 500 ft lower, and there were two still lower surfaces represented by terraces or straths. Later writers have assigned less significance to these surfaces (see Threet, 1959). There are pronounced flat areas or benches near the tops of the limestone outcrops. Sagebrush Flats on Mount Timpanogos are at an elevation of about 8500 ft, as is the irregular bench along the outcrop line of the Mississippian carbonates in the eastern Uintas. These appear more likely to be stripped structural surfaces where the less resistant Manning Canyon shale has been removed from the top of the Great Blue limestone in the Wasatch and where the Doughnut formation has been removed from the Humbug formation in the Uintas.

Both ranges were glaciated during Pleistocene time. Much of the core of the Uintas is mantled with glacial material, and high flats in the Wasatch are likewise covered with morainic material. Both U-shaped and V-shaped canyons occur in the Wasatch. The high peaks of both ranges have been sculptured into deep cirques. Glaciation has played a role in the karst processes, and glacial material is an important feed stock of sediment transported through the underground drainage system.

Atwood (1909) distinguished two periods of glaciation and offered some evidence for a third in what remains to this day the most extensive investigation of the glacial geology. Bradley (1935) also recognized an older glacial period on the north slope of the Uintas. Table 1 shows the presently accepted correlations between the glacial periods.

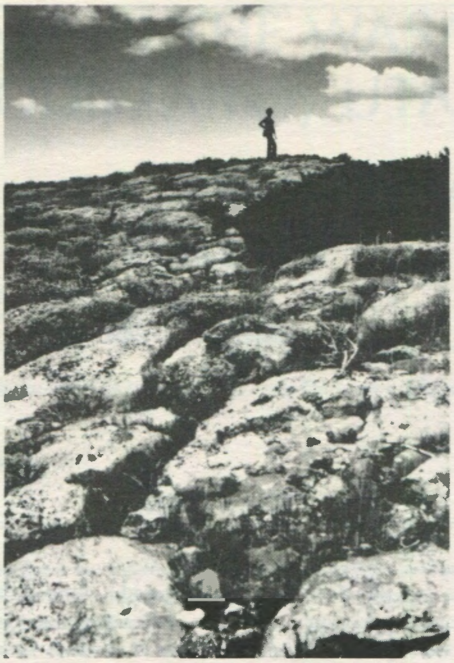


Figure 3. Limestone pavement with irregular kluftkarren. Little Brush Knob, north of Vernal, Utah.

Figure 4. Poorly developed rillenkarren on sloping limestone ledge, Little Brush Knob, north of Vernal, Utah.

LANDFORMS

Pavements and Karren

Most of the Mississippian carbonate outcrop is covered with thin soil and sparse vegetation. Here and there in both the Wasatch and Uinta ranges are patches of bare limestone pavement cut by joints which have been widened by solution to form kluftkarren and cutters. The pavement of Figure 3 is typical. Bare rock areas, many with solution grooving, are more common in the Wasatch Range with its steeper slopes and more rugged terrain.

Solution along lines of structural weakness produces a criss-cross pattern of solution grooves one to a few millimeters deep. Solution grooves follow small joints and fractures, which are very numerous in the Wasatch carbonates where there has been active tectonic movement. Further solution along joints produces a solutionally widened fracture form known as kluftkarren (Bögli, 1960). Most of those examined are very shallow, usually spanning only one or two beds. In the Soapstone Basin, there has been additional solution along the bedding planes to produce a shingle of loose slabs, mostly still in place.

Solution along vertical joints and fractures produces a linear feature of negative relief which will be called a cutter (following Howard, 1963). Cutters are a landform rather than a sculpturing of the bedrock but are transitional with kluftkarren as the size and depth of the feature is decreased. Cutters were not observed in the Uintas, although there is the possibility that they are masked by soil and glacial material. Small cutters, transitional forms, were observed high in the cliffs of American Fork Canyon in the Wasatch Range.

Channel forms appear sculptured into the bedrock as rare and small irregular winding half-tubes (rinnenkarren). A more common form that occurs on some sloping ledges and boulders is poorly developed rillenkarren (Fig. 4). Rillenkarren are parallel channels a centimeter or two in width separated by a sharp cuspate divide. The cuspate ridges can be seen in Figure 4, but the channels are not very deep or very extensive.

Pinnacles

Residual limestone pinnacles are found in both the Wasatch and the Uinta karsts. In American Fork and Big Cottonwood canyons, the pinnacles occur scattered along the canyon walls (Fig. 5).

TABLE 1. Correlation of Glacial Deposits.

Uinta-Wasatch Mountains	Wind River Mountains	Central United States
Smith Fork (Atwood's "Younger" Deposits)	Pinedale	Late Wisconsinan
Blacks Fork (Atwood's "Older" Deposits)	Bull Lake	Early Wisconsinan
Little Dry Creek	Buffalo	Illinoian

Figure 5. (right) Pinnacles on north wall of Big Cottonwood Canyon, Wasatch Mountains.

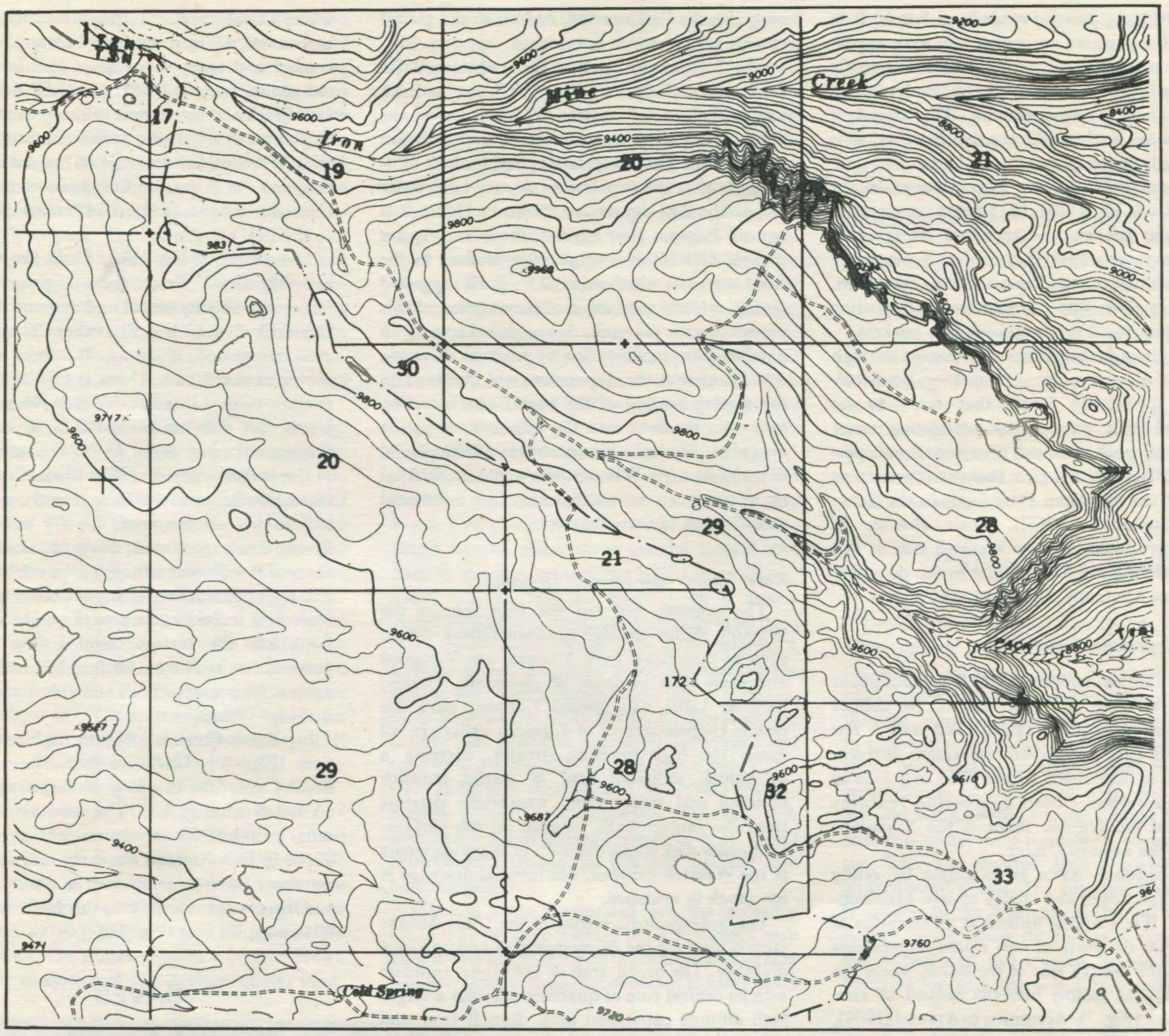


Figure 7. Topographic map of the eastern portion of the Soapstone Basin, showing location and distribution of closed depression features. From U.S. Geological Survey Iron Mine Mountain Quadrangle. Section lines (1-mile grid) give the scale.

Figure 6. Pinnacles on east wall of Dry Fork Canyon, eastern Uinta Mountains.

Figure 8. Closed depression features, Soapstone Basin, western Uinta Mountains.

The individual pinnacles range from 5 to 25 ft in height. The surfaces are rough and have apparently been fractured by frost action. Smoothly sculptured surfaces are uncommon. Both of these Canyons have V-cross-sections. The abundant pinnacles are further evidence that these canyons escaped glaciation.

Pinnacles occur in Dry Fork Canyon in the eastern Uinta karst (Fig. 6). These have more-or-less regular spacing and appear to be remnant rock masses separated by more deeply weathered zones along fractures. Some surfaces are rough, but others show smooth, solutional sculpturing and contain small tubes and solution pockets.

Small pinnacles occur here and there in the high Wasatch. However, the rounded or toadstool-shaped spitzkarren forms that seem to be associated with solution by percolating water under snow pack were not observed. Small pinnacles of the toadstool form occur in other alpine karsts of the American West (*e.g.*, in the Teton Karst [Stellmack, 1965]). Their absence in similar environments in the Wasatch may be due to the differing rates of solution in the dolomitic limestones. Certainly, the snowpacks occur.

Closed Depression Features

Dolines are an uncommon landform throughout both the Wasatch and Uinta karsts. No dolines were observed in the Wasatch, and only a few small and irregular closed depression features were seen in the eastern Uintas. The only area in which numerous closed depression features appear on topographic maps is the Soapstone Basin in the western Uintas.

The Soapstone Basin is underlain by gently dipping limestones and shales of late Mississippian age. It is a rolling upland at an elevation of about 9500 ft. Most of the closed depressions occur on the eastern side of the basin, where the upland breaks steeply into the canyon of Iron Mine Creek (Fig. 7). According to Atwood (1909), the Soapstone Basin was glaciated only during the early (Blacks Fork) period. Moraines and glacial gravels are not much in evidence.

The closed depressions of the Soapstone Basin are broad and shallow. Their walls are bedrock with a shallow soil mantle. The bottoms of the dolines are flattish and contain an accumulation of soil, although the vegetative cover (or lack of it) suggests that the depressions do not hold water effectively. The depression shown in Figure 8 is typical. It is not clear whether these features are entirely of solutional origin (*i.e.*, dolines in the strict sense) or whether (as D.C. Ford suggested during discussion of the oral presentation of this paper) they are primarily due to glacial scour with possible later modification by solution. It seems significant that the depressions are clustered in the eastern portion of the basin, where a steep hydraulic gradient into the adjacent canyon is available. Atwood's interpretation of the glacial deposits as early Wisconsinan would imply that there has been substantial time for solutional modification of these features.

Karst Valleys and Internal Drainage

The canyons that cut the west face of the Wasatch Range are high-gradient valleys. Some, such as American Fork Canyon, are V-shaped and show little evidence of glaciation. Others, such as Little Cottonwood Canyon, have the classic U-cross-section of a glacial valley. In all cases, the high gradient streams, carrying a substantial sediment load of glacial outwash material and more recent weathering detritus from the canyon walls, remain on the surface. Carbonate rock forms the walls of many segments of the Wasatch canyons, but internal drainage is not much in evidence.

The hydrogeologic setting of the Uinta Mountains is ideal for the development of underground drainage. The broad arch of the Uinta anticline with its central core of quartzite provides a large, high altitude catchment area. Runoff from the central core flows down the flanks of the Uintas, where it meets the outcrop line of the Mississippian carbonates. Many streams go underground at the contact and reappear as large springs at lower elevations. Marsell (1969) mentions the sink of Lost Creek and its relation to Sheep Creek Spring on the north slope. Big and Little Brush creeks and the associated drainage of the Ashley Creek Basin on the eastern end of the south slope have been the subject of much attention (Maxwell, *et al.*, 1971).

The valleys of Dry Fork, Little Brush Creek, and Big Brush Creek form a progression of increasing effectiveness of underground drainage. Dry Fork Canyon is a dry valley. The tributary streams are lost along an ill-defined reach of cobble-armored bed. There are no well-defined swallow holes. Little Brush Creek flows from the Uinta core and has incised a blind valley into the original valley floor. At the downstream end of the incised canyon, Little Brush Creek flows into Little Brush Creek Cave (Fig. 9). Above the cliff at the cave entrance, the dry bed of Little Brush Creek continues down the valley. The channel is well-maintained, with a cobble armoring. It is apparent that during periods of high flow, the entire incised canyon back-ponds and overflows into the surface channel. The flooding generates an additional 50 ft of hydrostatic head against the obstructions in the underground drainage system.

Big Brush Creek is lost into Big Brush Creek Cave (Fig. 10). Upstream from the cave, Big Brush Creek flows in a deep, U-shaped valley with a V-notch cut in its floor. The discontinuity in the valley profile at the cave entrance is on the order of 100 ft. Downstream, above the cliff at the cave entrance, there is no trace of a stream channel and there is no evidence that this blind valley ever fills and spills over (Fig. 11).

CAVES

Caves are widely but sparsely distributed in both the Wasatch and the Uinta mountains.

Figure 9. View down the incised blind valley of Little Brush Creek, toward the entrance to Little Brush Creek Cave. The stream flows directly into the entrance. The spillover onto the continuation of the surface channel is at the top of the cliff, in the notch to the left of the cave entrance.

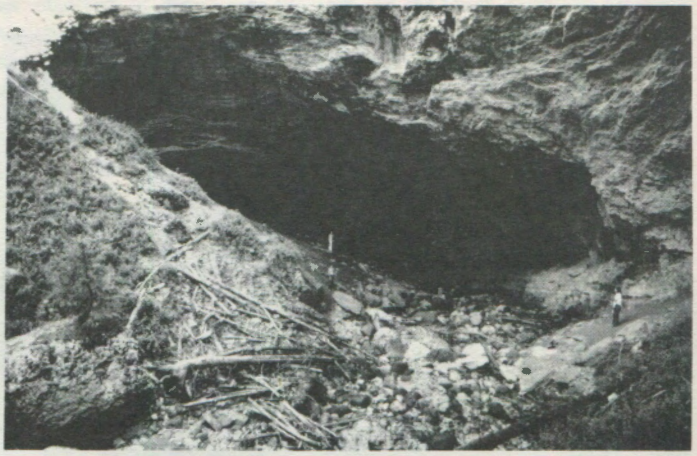


Figure 10. Entrance to Big Brush Creek Cave. Big Brush Creek sinks in Capture Cave, a half mile upstream, during dry weather. The cliff rises about 100 feet above the entrance to the floor of the old surface valley.

Entrances are obscure, and the collection of cave descriptions that appear in the Salt Lake Grotto Technical Notes represent many man-hours of searching. The length distribution (Fig. 12) shows that the caves, with the exception of the Brush Creek caves, are rather small. There is no significant difference between the lengths of Wasatch and Uinta caves, but the sample is too small to be very conclusive. The length distribution pattern is poorly developed. The population statistics of cave length distributions can only be determined if the cave areas have been searched systematically, so that the correct proportions of long and short caves have been described. Systematic searches of only a few areas in the western Uintas, including the Soapstone Basin, have been reported (Haman, 1963, 1967, 1968).

Timpanogos Cave (White and Van Gundy, 1974; Green, 1975-b) is the largest cave in the Central Wasatch. It is really three different caves, with three entrances which have been connected by tunnels. Two of the segments, Hansen Cave and Middle Cave, are parts of the same cave and have developed along two parallel faults. The remaining and larger segment, Timpanogos Cave, is also developed along a fault, but a fault that is rotated about 15° with respect to the Middle Cave fault. The Hansen and Middle Cave faults are nearly vertical, while the Timpanogos Cave fault dips at about 45°. The northwest wall of the main passage of Hansen Cave is directly on the fault. It is shattered and unmodified by solution. Flowstone is cracked along the Middle Cave fault.

The main passage of Timpanogos Cave has a pronounced dip with a vertical relief of about 50 ft, whereas the main trend of Hansen-Middle Cave is more or less horizontal. A small side passage in Timpanogos Cave has been trenched to provide a walking-height bypass to part of the tourist trail. The trench exposes a bedded brown sediment consisting of fine-grained quartz sand. The beds slope about 15° and parallel the slope of the passage. Since the block faulting of the Wasatch Mountains is known to have taken place throughout the Pleistocene, the evidence in the cave is that some movement of these small faults post-dates the formation of the cave and even, in the case of the Middle Cave fault, post-dates the flowstone deposits. The fault may be in motion at the present time. The dipping sediments suggest a tipping of the fault block that contains Timpanogos Cave relative to the block that contains Hansen-Middle Cave since the development of the cave passage and the deposition of the sediments.

Most of the larger caves are associated with streams draining from the eastern portion of the Uinta Mountains. These include the Pole Creek Caves (Green, 1975-a), the Dry Fork Cave System (Green, 1957), and the Brush Creek Caves (Green 1971-a, b; Lyon, *et al.*, 1972).

The Brush Creek caves are anomalous. At 26,000 ft and an explored depth of 800 ft (Big Brush Creek) and 15,000 ft and 500 ft (Little Brush Creek), these caves are the longest and deepest in Utah (except for Neff Canyon Cave, which is deep but not long). Each is associated with a large surface catchment and associated blind valley. Each has the same basic pattern: A large-diameter, elliptical trunk channel at the stream sink which disintegrates into a complex, three-dimensional maze further in. The walls are mainly bare bedrock, intricately sculptured and scalloped. Palmer (1975) used Big Brush Creek Cave as his type example of a floodwater maze—a maze pattern produced when extremely high hydrostatic heads during periods of high runoff drive water through all possible fracture and bedding planes.

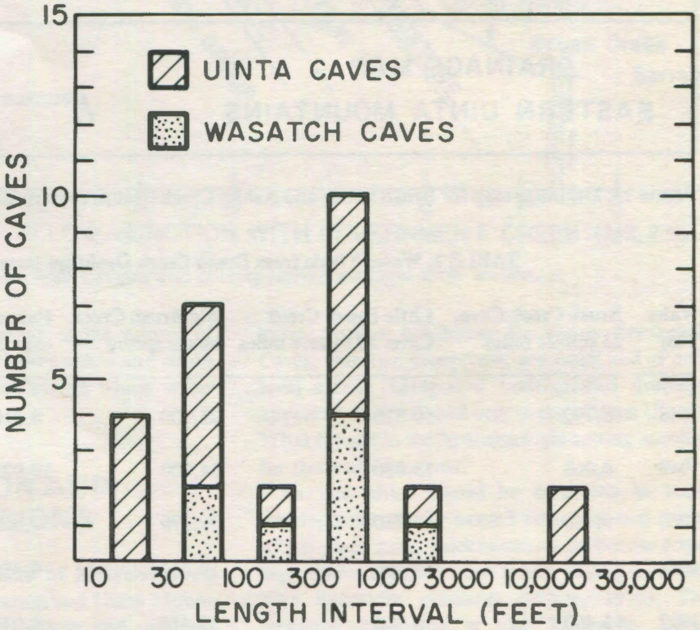
What is more mysterious are the entrance tubes. The entrance passage in Little Brush Creek Cave is almost choked with breakdown and stream boulders, and, indeed, the openings to inner parts of the cave open and close as annual floods re-arrange the sediments. The entrance trunk in Big Brush Creek Cave can be followed for about 1000 ft, to a point where the ceiling becomes low and the passage is interrupted by breakdown. What is unknown is whether the trunks and the interior mazes developed as continuous parts of the same passage, or whether the trunks are completely filled with boulders and sediment and the mazes have been developed above them. Certainly, the profiles of the caves appear to be continuous, but the possibility exists that large, conduit caves developed earlier than the mazes and were filled, perhaps by outwash during the glacial retreat from the Uintas. Alternatively, one could imagine the entrance trunks having formed in the zone of intense weathering. Seasonal freezing and thawing, combined with the erosive power of the annual flood which would strip away the fractured rock, would convert a maze of smaller passages into a conduit.

HYDROLOGY

Underground drainage through conduits in the Mississippian limestones is most extensive in the eastern end of the Uintas. The karstic drainage has provided no end of problems for residents of the area, who for years have tried various devices to maintain flows in the creeks draining from the Uinta core. The water supply problems are most severe during the summer months, when flows are low and much needed for stock water and irrigation. It is under these conditions, of course, when the flows are diverted almost completely into underground routes. The discussion that follows is limited to the drainage basins of Brush and Ashley creeks, where the results of extensive flow monitoring and dye tracing by the Bureau of Reclamation and the Soil Conservation Service

Figure 11. (above) View down-valley from a point directly above the entrance to Big Brush Creek Cave, showing the broad U-shaped cross-section and the lack of a stream channel.

Figure 12. (left) Distribution of cave lengths for caves in the Wasatch and Uinta Mountains. The source of data was the Salt Lake Grotto Technical Notes.



(Maxwell, *et al.*, 1971) provide some clues to the underground drainage routes.

The drainage pattern is summarized in Figure 13. Both Big and Little Brush Creeks terminate at the entrances to the respective caves at elevations a little above 8000 ft. The water from both caves is known to resurge at Brush Creek Spring, rising along a fracture zone through the Weber sandstone at an elevation of 6040 ft. All three forks of Dry Fork sink into their beds. Water that sinks in a swallow hole on the main (western-most) fork reappears at Deep Creek Spring, at the head of Deep Creek, from whence it flows into the Uinta River. Water that sinks in the beds of all three forks flows to the south and east. Structural complexity due to the Deep Creek fault zone complicates things. It appears that there is a direct connection with Ashley Spring, but that some of the water also appears at Dry Fork Springs. Dry Fork Springs discharges groundwater over an area of several acres. Ashley Spring is a major spring in the Weber sandstone.

The U.S. Geological Survey has maintained gaging stations on Big and Little Brush creeks just above the cave entrances and on the main channel of Brush Creek at the Route 44 highway bridge, about 3.5 mi below Brush Creek Spring. These data permit some estimate of the overall water balance and flow conditions through the unobserved portions of the drainage system. Water yields for five consecutive years are tabulated in Table 2. By assuming that all water that passes the gages in the upper basins flows through the underground drainage system, and by comparing the sum of these two with the flow in Big Brush Creek at the downstream gage, it is possible to estimate at least minimum percentages of the overall runoff from the high Uintas that must pass through the cave systems. These numbers range from 67 to 86%, depending on precipitation conditions. The remaining water can be ascribed to surface runoff from the basin downstream from the sinks or to groundwater contributions to the spring from other sources. No discharge measurements are available at the spring itself, since it is located in the bottom of Brush Creek Canyon and is not easily accessible from any direction.

Most of the recharge to the Brush Creek and Ashley Creek basins is in the form of winter snow pack in the high Uintas. Discharge of the creeks during the winter months is very low (Fig. 14). When the snow melt begins in late April or early May, discharge rises rapidly and there is maximum discharge during May and June. With complete melting of the snow pack and the advance of summer, flows again decrease until by July they have reached their base flow values again. Flows of both Big and Little Brush creeks are interrupted by reservoirs and diversions into canals at Oaks Park and East Park, so that no significance can be attached to the shapes of the hydrographs.

Data for two water years, plotted with somewhat greater resolution in Figure 15, show that the discharge from Brush Creek Spring remains in phase with the inputs into the caves. When the input begins to rise at the beginning of the snow melt period, the discharge of the spring rises abruptly within a 24-hour period, which is the resolving power of the data. This does not, of course, necessarily mean that the water moves through the system in less than 24 hours. Rising water levels in the lower portions of the caves generate a pressure head which, in turn, begins to force water from storage in the aquifer out through the spring. The fast response does, however, imply a very well integrated system which allows rapid transmission of the pressure pulse.

The dye tracing experiments on Big Brush Sink described by Maxwell, *et al.* (1971) gave 77 hours as the travel time through the system. This included flow time in the mile of surface channel between the highway and the cave and

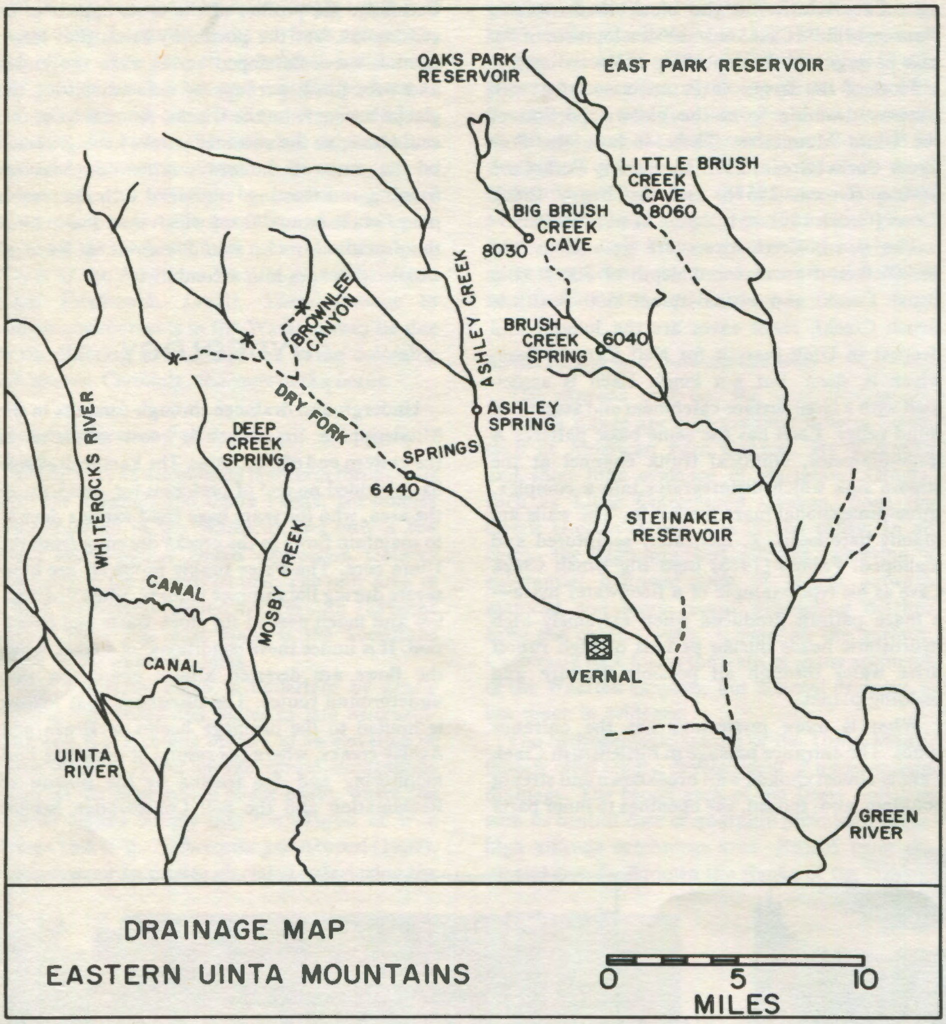


Figure 13. Drainage map for Brush Creek and Ashley Creek Basins, eastern Uinta Mountains.

TABLE 2. Water Yields from Brush Creek Drainage (acre feet).

Water Year	Brush Creek Cave, 23 square miles	Little Brush Creek Cave, 28 square miles	Big Brush Creek below spring	Excess flow over cave input	Input through caves
1948	5,790	10,820	25,100	8,310	67%
1949	8,960	15,020	34,000	10,020	71%
1950	9,440	19,760	36,270	7,360	81%
1951	3,910	9,300	19,660	9,450	67%
1952	11,460	18,920	35,410	5,030	86%

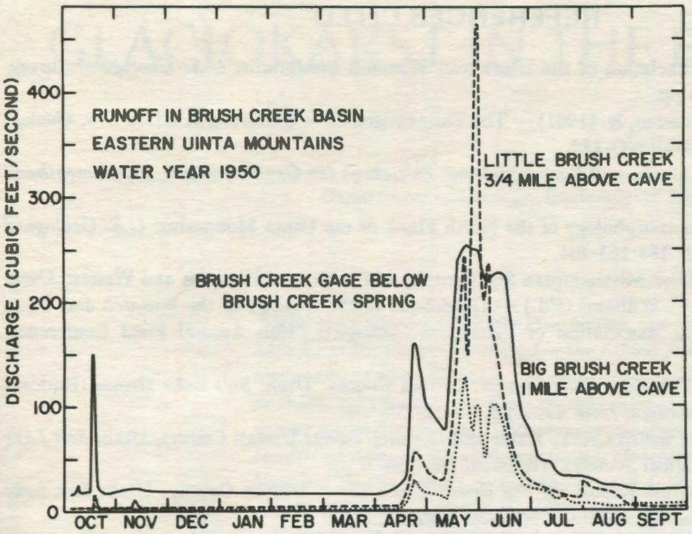


Figure 14. Annual hydrograph for runoff from Little Brush, Big Brush, and downstream Brush Creek basins. Data from U.S. Geological Survey surface water records.

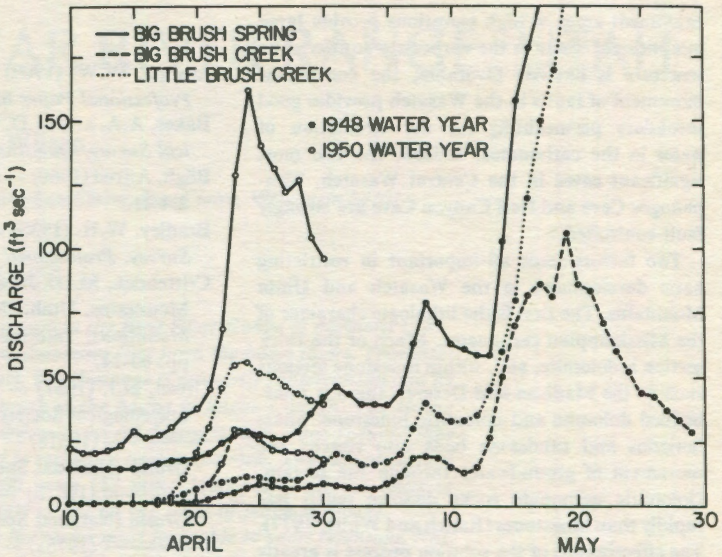


Figure 15. Detail of discharge behavior of Brush Creek Spring in comparison with runoff into the Brush Creek Caves.

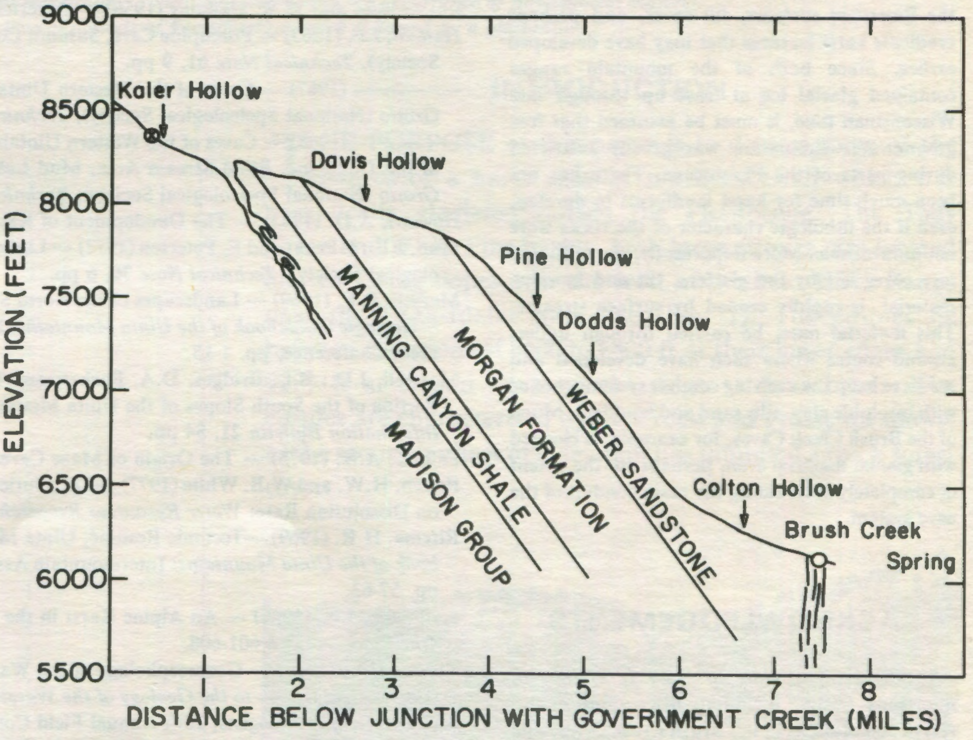


Figure 16. Profile of Brush Creek Canyon and corresponding geologic cross-section.

the 3.5 miles of surface channel between the spring and the collection point at highway 44.

The rapid flow-through times and response times pose some interesting and unanswered questions about the underground drainage system. The dip of the Madison limestone group is steeper than the slope of the surface channel, so that the conduit system in the limestone is carried deeper and deeper into the sub-surface at lower elevations (Fig. 16). Big Brush Creek Cave has been explored to a depth of 800 feet (D.J. Green, personal communication) but continuation appears to be difficult. In Little Brush Creek Cave, explorers have hit standing water levels at 500 foot depths. This is only a fraction of the more than 2000 ft difference in elevation between the caves and the spring. Further, if the dip is extrapolated linearly, the Madison group passes under the location of the spring at depths on the order of 2000 ft below the spring orifice.

The question, then, is how does the water get from the conduits in the Madison limestone across the stratigraphic section to the spring in the Weber sandstone? The simple picture of Figure 16 would require water to move up nearly 2000 ft along a fracture zone through the Manning Canyon shale, the Morgan formation and much of the Weber sandstone to reach the spring mouth. The driving force would be the hydrostatic pressure of water dammed in the limestone conduit system by the impermeable Manning Canyon shale. However, this long flow path, much of it through fractures in non-soluble rocks, does not seem consistent with the observed flow behavior, and one must postulate additional structural changes at depth that would allow a more direct connection. Although this interpretation does imply an open conduit system in the limestone at least to depths equal to the elevation of the spring, it does not follow that these conduits will be traversable by humans. The cobble and boulder fills that occupy much of the volume of the cave passage are highly permeable and could effectively block a passage through which water could pass very quickly.

DISCUSSION AND CONCLUSIONS

Extensive outcrop areas of Mississippian carbonate rocks in the Wasatch and Uinta Mountains have produced only a sparse and subdued karst. Surface landforms are poorly developed. Caves, with two exceptions, are small and of only local extent. Integrated underground drainage appears well developed only in the eastern Uintas. What factors in the hydrogeologic setting account for these observations?

Factors which would be favorable to karst development are the overall structure and relief, which place great thicknesses of carbonate rocks high above regional base levels, thus generating high hydraulic gradients (White, 1977). The topographic setting is also favorable. Large catchment areas at high elevations provide large quantities of water to the carbonate aquifer. The structure is likewise favorable; the continuous movement of faults in the Wasatch provides good secondary permeability for the circulation of water in the carbonates. Indeed, the two most significant caves in the Central Wasatch, Timpanogos Cave and Neff Canyon Cave are strongly fault-controlled.

Two factors seem all-important in restricting karst development in the Wasatch and Uinta Mountains. The first is the lithologic character of the Mississippian carbonates. Much of the thick section is dolomite, and, within limestone sections such as the Madison and Deseret, there is interbedded dolomite and dolomitic limestone. Shale horizons and sandstone beds also restrict the movement of groundwater through the section. Dolomitic carbonate rocks dissolve much less rapidly than limestones (Rauch and White, 1977). The effectiveness of the solution process is greatly diminished by the character of the bedrock.

The second factor is the influence of the Pleistocene glaciations. Glaciation has acted to scour the limestone surfaces, fill caves, and perhaps eradicate karst features that may have developed earlier. Since both of the mountain ranges contained glacial ice at least up through late Wisconsinan time, it must be assumed that free groundwater circulation was greatly restricted during parts of the Pleistocene. There has not been much time for karst landforms to develop, even if the lithologic character of the rocks were not unfavorable. More importantly, the sedimentary cover left by the glaciers, till and moraine material, is rapidly eroded by surface streams. This material must be carried through underground routes where such have developed and serves to keep the evolving conduit systems choked with insoluble clay, silt, sand and boulders. Much of the Brush Creek Caves, for example, is clogged with glacial material even, perhaps, to the extent of completely obliterating the main conduit of the cave system.

ACKNOWLEDGEMENTS

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GLACIOKARST IN THE BEAR RIVER RANGE, UTAH

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ABSTRACT

The Tony Grove and White Pine basins, two large cirques in the Bear River Range of northern Utah, have karst landforms developed in the folded Paleozoic dolomites that form the basin floors. Precipitation averages 50 to 60 inches per year (water equivalent). The drainage is primarily subterranean and its probable resurgences are three springs several miles south of the basins in Logan Canyon. The spring waters are undersaturated with respect to calcite and dolomite except during winter, when reduced flow conditions exist. Degree of saturation is closely related to residence time in the rock. The rate of surface lowering in the basins is estimated at 2.1 in./ 1000 years (53 mm/ 1000 years) on the basis of geochemical data and 1.8 in./1000 years (46 mm/1000 years) on the basis of perched blocks.

Surface solution features tend to be concentrated in the sector most favorable to snow retention, N5E to N90E with respect to the nearest ridge. Scouring by glaciers has produced dolomite pavement broken by scarps formed along chert-rich beds. Caves within the basins are small, joint-controlled, and are generally restricted to areas peripheral to the path of glacial advance. The anomalous position of the caves relative to the present topography, solution breccias, and the occurrence of fluvioglacial material in karst features suggest several periods of karstification, each followed by a glaciation.

INTRODUCTION

THE PURPOSE of this paper is to discuss some of the features of a karst area in northern Utah which was periodically glaciated during the Pleistocene. The Tony Grove and White Pine basins, both large cirques, are located on east-facing slopes near the crest of the Bear River Range near the Utah-Idaho state line (Fig. 1). The cirques lie between elevations of 8600 to 8800 ft at their lower edges and 9979 ft at Naomi Peak on the ridge at their heads. The cirques are characterized by bare rock and grassy meadows, as the scouring effect of glaciation has inhibited a thick forest cover.

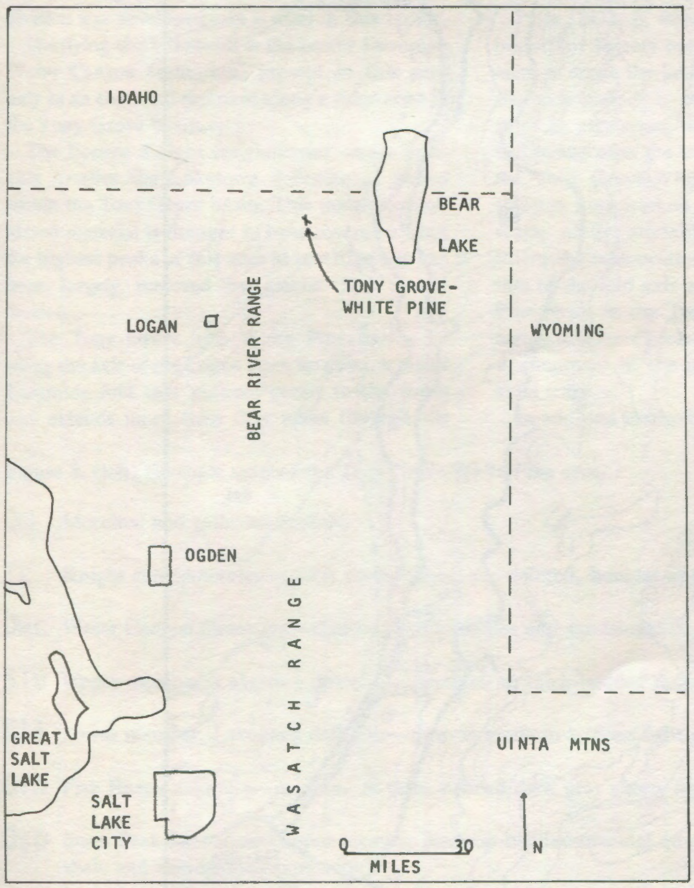
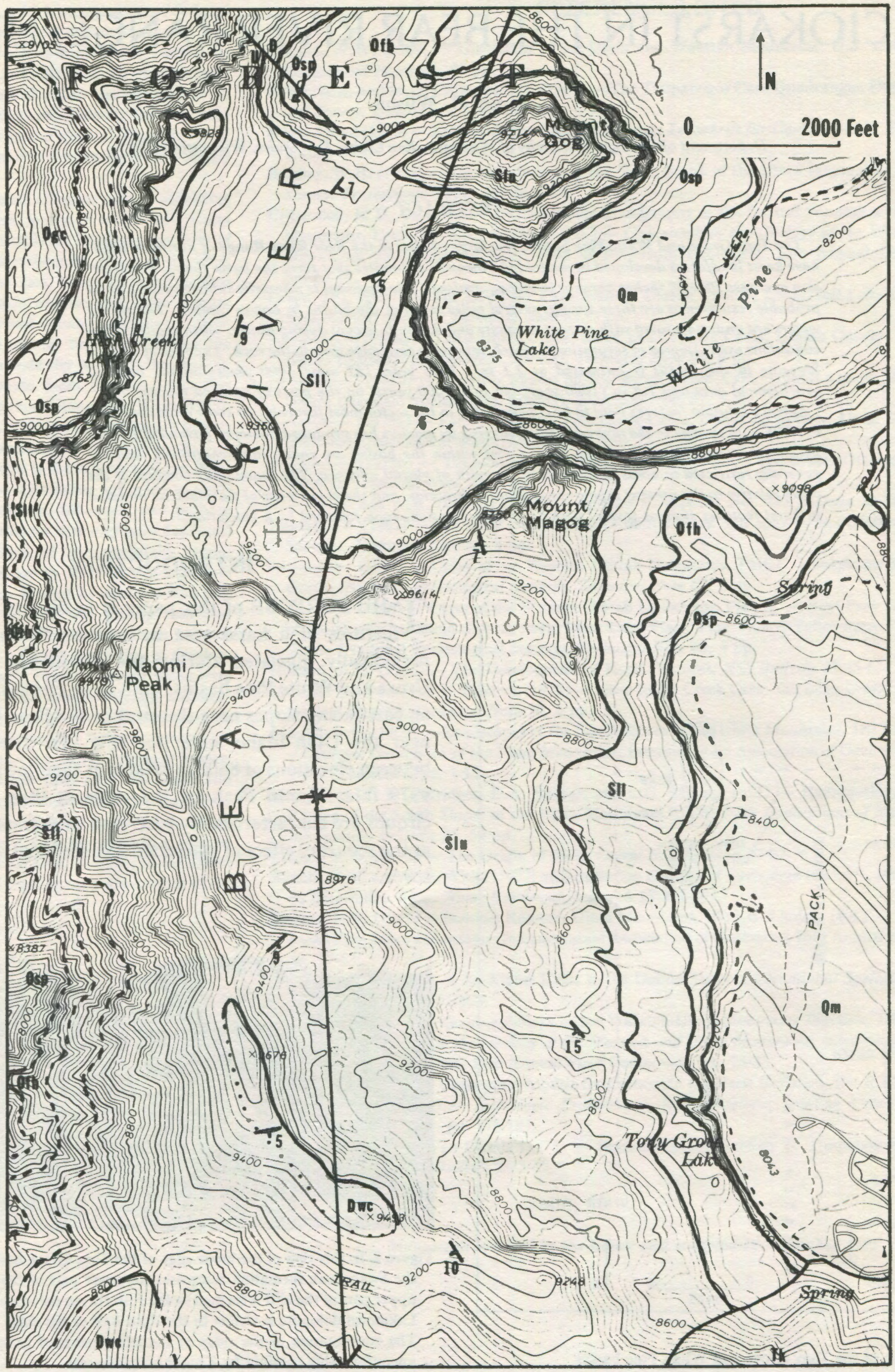


Figure 1. Location of the Tony Grove and White Pine basins.

Figure 2. In this view of the lower edge of the Tony Grove basin, quartzite of the Swan Peak formation comprises the lower part of the cliff face. The upper part of the cliff is formed by the Fish Haven dolomite. The overlying Laketown dolomite is visible in the right half of the ridge on the skyline. The continuation of the ridge to the left consists of Laketown dolomite unconformably overlain by Knight conglomerate.


N
0
2000 Feet
F O R E S T
B E A R R I V E R
Mount Gog
Mount Magog
Naomi Peak
White Pine Lake
White Pine
High Creek Lake
Tony Grove Lake
Spring
Spring
PACK
TRAIL
JEEP
Qm
Osp
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Slu
Sll
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8600
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9400
9800
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9350
15
10
5

GEOLOGY

Rocks exposed in this area comprise the Lower Ordovician to Lower Devonian section plus one Eocene formation (Fig. 2). The Lower Ordovician Garden City formation, predominantly limestone, does not crop out in either of the basins but underlies both and is exposed a few hundred yards downstream from Tony Grove Lake.

The upper quartzite member of the Middle Ordovician Swan Peak formation is a distinct cliff former and useful stratigraphic marker. The lower shale member, in contrast, is seldom seen, for it is generally covered by colluvium from the cliffs above. The shale forms an impermeable horizon and the lakes at Tony Grove, White Pine, and High Creek occupy bowls that were carved into it by the glaciers.

Above the Swan Peak formation is the Upper Ordovician Fish Haven dolomite, a dark gray, cherty, cliff-forming unit. Most of this formation is exposed in the cliff faces around the lower edges of the basins.

The Silurian Laketown dolomite is a thick unit which forms the floors and surrounding ridges of the basins. The Laketown section at Tony Grove was measured by Budge (1966, pp. 72-76), who divided the Laketown into four members. This writer was unable to identify these members with any certainty while in the field so a two-fold division was developed and is used in this report.

Overlying the Laketown is the Lower Devonian Water Canyon formation, present in this area only as an erosional remnant along a ridge crest in the Tony Grove basin.

The Eocene Knight conglomerate unconformably overlies the Laketown dolomite at places within the Tony Grove basin. This poorly consolidated material is thought to have covered all but the highest peaks of this area at one time but has been largely removed by glacial and fluvial erosion.

The Tony Grove and White Pine basins lie along the axis of the Logan Peak syncline, a major Laramide fold that plunges gently to the south and extends more than fifty miles through the Bear River Range. The axis of the syncline passes southwestward through Mt. Gog, then makes a relatively sharp bend to a more southerly direction immediately south of the ridge between the two basins (Fig. 3).

The Laketown dolomite has been intensely jointed, and many of the joints have been enlarged by solution. The major trend of the joints is subparallel to the fold axis and may represent expansion fractures created when the pressure causing the folding was released.

A small fault occurs in the northern extremity of the area studied. Five or six small springs occur along the fault trace. At other places in the Tony Grove-White Pine area, fracture zones up to three feet wide have been noticed, but there is no noticeable displacement of the rocks.

EROSION AND MASS WASTING

Snow accumulation on the eastern side of the Bear River Range during the Pleistocene led to the development of glaciers and to erosion by glacial ice; in contrast, the western slopes have been shaped by fluvial erosion. In a study of the glacial history of the Tony Grove basin, Williams (1964) distinguished two Bull Lake advances and at least two Pinedale advances, on the basis of morainal deposits in the area.

Price (1973, p. 49) stated that one of the most important factors controlling the ability of glaciers to erode the land "...is the preparation of the rock surface by various forms of weathering prior to glaciation, and structural failure, both before and after the initiation of the ice cover." In the Tony Grove-White Pine area, jointing and solution have worked together to provide a landscape easily erodable by glaciers. Extensive fracturing is associated with the changing orientation of the fold axis as it passes from the White Pine basin to the Tony Grove basin. This fracturing may have been a factor in localizing cirque development in the immediate proximity of the axial trace.

In addition to the cirques, glacial erosion in the basins is evidenced by ice-scoured bedrock and by a series of scarps quarried by the ice. In addition, glacial striae and polish are visible on bedding planes of the quartzite of the Swan Peak formation at the lower edge of the Tony Grove basin. (Fig. 4).

Fluvial activity is quite limited in both basins. Surface streams exist only in early summer, when runoff from the melting snowpack exceeds the infiltration capacity of the alluvium-filled dolines.

Mass wasting in this area at the present time consists of earth slides and flows in the Eocene Knight conglomerate and rockfall and snow avalanches along the steep headwalls of the cirques. Slides in the conglomerate are facilitated by the lubricating action of water flowing along the interface between this formation and the underlying Laketown dolomite.

HYDROLOGY

Winter storms in northern Utah deposit large amounts of snow on the mountains that lie along their path. Precipitation maps for Utah prepared by ESSA-USWB (no date) indicate that the basins along the crest of the Bear River Range

Figure 4. Glacial striae and polish on the Swan Peak formation, near Tony Grove Lake.

Figure 3. (left) Geologic map of the Tony Grove-White Pine area.

Qm Morainal and colluvial deposits

Tk Knight conglomerate—weakly consolidated, red-colored, boulder and cobble conglomerate

Dwc Water Canyon formation—thin-bedded dolomites and sandstones

Slu Upper member, Laketown dolomite—medium- to thick-bedded dolomites and cherty dolomites

Sll Lower member, Laketown dolomite—thin- to medium-bedded light gray dolomite

Ofh Fish Haven dolomite—medium- to thick-bedded dark gray cherty dolomite

Osp Swan Peak formation—upper member medium-bedded fucoidal quartzite, lower member black shale and thin-bedded sandstone

Ogc Garden City formation—thin- to medium-bedded limestone, dolomitic limestone, and shaly limestone.

Contact, dashed where approximate

D U Fault, showing relative movement

Axis of plunging syncline

7 Strike and dip of beds

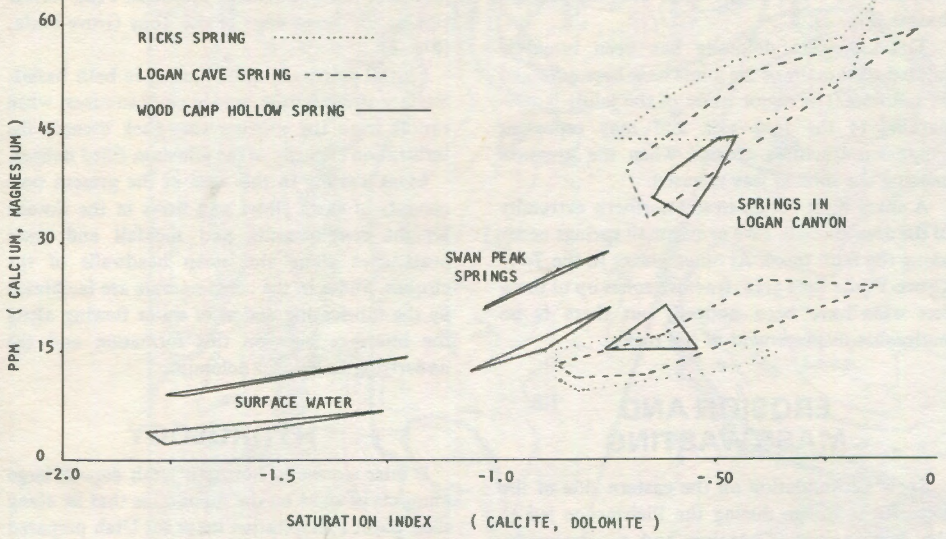

Figure 5. Range of saturation of water samples with respect to calcite (upper figures, calcium plotted against the calcite saturation index) and dolomite (lower figures, magnesium plotted against the dolomite saturation index) for three samples of surface water, three samples of Swan Peak spring water, and 17 to 28 samples from the three springs in Logan Canyon.

receive 50 to 60 inches (water equivalent) annually, of which 80% falls between October and April.

Most of the precipitation that falls in the basins is drained internally. Probable resurgences are three large springs in Logan Canyon, three to six miles south of the basins. Ricks Spring lies on a north-south trending fault in the basal part of the Garden City formation. Its discharge varies from nothing during March of some years to 150 cfs in June, with a strong correlation between discharge and the rate of snowmelt occurring in the basins.

Logan Cave Spring is located in dolomitic limestone of the uppermost part of the Garden City formation. It flows from a talus slope below the entrance to Logan Cave with a discharge that varies from 1 to 25 cfs. The water is visible within the cave for approximately 1000 ft.

Wood Camp Hollow Spring flows from a series of risings in a talus slope along the bank of the Logan River. Stratigraphically, it is located in the Laketown dolomite near the contact with the underlying Fish Haven dolomite. Discharge varies from 10 to 65 cfs.

Below the Tony Grove and White Pine basins, a number of small- to intermediate-sized springs rise on the colluvium covered slopes. These springs are apparently located at the contact between the quartzite and shale members of the Swan Peak formation.

Geochemical sampling of the water in these springs was conducted at frequent intervals from April, 1975, to August, 1976. Some early results are presented in Wilson (1976). More complete data and analysis are in preparation. Data collected at the springs during the study indicates that all the springs are undersaturated with respect to both calcite and dolomite, except at periods of negligible flow. Residence time of the water in the rock seems to be a major factor in determining degree of saturation (Fig. 5). For both calcite and dolomite, surface streams are least nearly saturated, springs in the Swan Peak formation are more nearly saturated, and the three large springs 3 to 6 miles distant in Logan Canyon are the most nearly saturated.

The mean, standard deviation, and coëfficient of variation of the hardness for the three springs in Logan Canyon have been calculated (Table 1) and can be used to classify the springs under the system proposed by Shuster and White (1971). According to their criteria, Logan Cave Spring represents flow in an open conduit and, since it is visible for some distance in Logan Cave, open conduit flow is demonstrated for this spring. In contrast, Ricks Spring and Wood Camp Hollow Spring have coëfficients of variation indicating flow through an open fracture system, not a conduit system (White, personal communication, 1976).

An estimate of the denudation rate within the basins can be made by assuming an annual precipitation of 50 inches and, also, assuming that the amount of dolomite being dissolved by a given volume of water is that necessary to give the water a magnesium content of 16 ppm (a figure representing near-surface solution, as indicated by analysis of water from springs in the Swan Peak formation). These assumptions result in a denudation rate of 2.1 in./1000 years (53 mm/1000 years).

TABLE 1. Statistical Parameters of the Total Hardness (ppm $CaCO_3$)

	Mean Hardness	SD	CV
Logan Cave Spring	190	30.7	16.2%
Ricks Spring	182	13.0	7.2%
Wood Camp Hollow Spring	171	10.2	6.0%

Another estimate of the denudation rate can be obtained from dolomite blocks perched on pedestals in the White Pine basin. These are not glacial erratics, but are residual pieces of an overlying, less soluble member of the Laketown dolomite. The distance from the top of the pedestal to the general rock surface is the minimum thickness of rock removed by weathering since deglaciation. Radiocarbon dating (sample # GX-3481) of a glacial sequence near Salt Lake City (D. R. Currey, personal communication, 1975) suggests that the Bear River Range was probably free of glaciers about 12,000 years BP. Calculations based on this time period and the greatest distance measured, 21.6 in., yields a denudation rate of 1.8 in./1000 years (46mm/1000 years); roughly equivalent to the value computed on the basis of geochemistry.

DOLINES, MEADOWS AND CORRIDORS

Moderate- to large-scale karst landforms in this area are (in generally decreasing size): cirque-dolines, solution dolines, meadows, collapse dolines, and corridors.

Cirque-dolines are glacial cirques which have been deepened by solution. Two of these features are found in the area; one each in the White Pine and Tony Grove basins. Their positions below the headwalls of the basins and the steepness of their slopes distinguish them from dolines of purely solutional origin.

Apart from the cirque-dolines, the other dolines found in the basins fall into two distinct groups. The first group (assumed to be solution dolines) consists of dolines that have little depth (generally less than 20 ft), small (50 to 200 ft) diameters, and flat soil-covered floors. They frequently contain standing water in early summer, and smaller secondary dolines may develop in their alluviated floors as the water drains away.

Dolines in the second group (assumed to be collapse dolines) have nearly vertical bedrock walls, are narrow relative to their length, are deep (20 to 60 ft), and commonly contain permanent snow and ice. Their form is strongly controlled by joints and fracture zones. In a few dolines, the snow melts enough by October to reveal blocks of dolomite rubble on the floor.

Meadows are morphologically similar to solution dolines in that they are elongate, shallow, alluvium-filled features. They differ only in that they are not internally drained, although local depressions may exist in a given meadow. The meadows probably originated as dolines, but water overflowing them during the spring runoff cut through their thresholds and established external drainage.

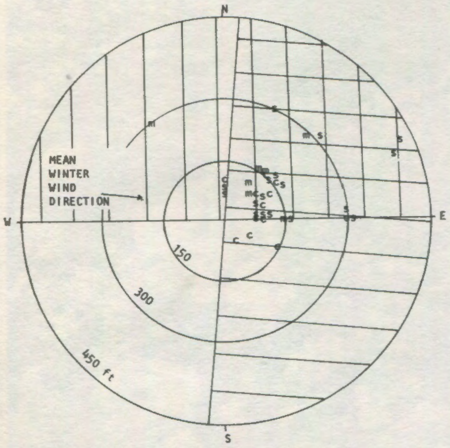

Figure 6. Relationship between surface solution features and lingering snow patches (see text). Cross-hatched area represents the zone where greatest snow drift and least insolation combine to favor snow retention. m—meadow, s—solution doline, c—collapse doline.

This view is strengthened by a study of the relationship between lingering snow patches and solution features. Rock beneath snow patches is more likely to exhibit solution features than other rock in the area, since meltwater from patches slowly percolates into the ground instead of rapidly running off. The permanence of snow patches is controlled by accumulation and insolation, both of which are functions of the topography within a given area. In this region, the vector mean wind at an altitude of approximately 10,000 ft during the winter months is N85W (Crutcher, 1958). The greatest accumulation of snow occurs on lee slopes, while the least insolation occurs on north-facing slopes. Thus the zone where the combined influence of these two factors is most favorable to snow retention is on slopes facing between N5E and N90E.

Figure 6 is a polar coördinate plot of 35 solution dolines, collapse dolines, and meadows that lie within 500 ft of a significant ridge or steep slope that would affect insolation or snow drift. Their orientation ($\pm 5°$) and distance (± 75 ft) with respect to the ridge or slope has been plotted. The correlation between the solution features and the sector favorable to snow retention is striking. Note the close association of meadows and solution dolines.

Several trough-like depressions, referred to as corridors, occur in the basins. Within the White Pine basin, a large corridor (approximately 800 ft long, 20 to 50 ft wide, and up to 20 ft deep) is oriented N30E coinciding with the major joint trend. Individual dolines have formed along the length of the corridor. Smaller, but similar, corridors are found in the Tony Grove basin.

PAVEMENT AND KARREN

Terminology relating to pavement and karren is discussed by Bögli (1960) and by Williams (1966). A "karst pavement" is an extensive exposure of carbonate bedrock, with its surface more or less parallel to bedding, that exhibits solutional modification. The usual implication is that the pavement was created by glacial scouring. A variety of pavement types are found in the Tony Grove-White Pine area. Exposures of horizontal pavement occur along the axis of the syncline, where the bedding is nearly horizontal. Inclined pavements are developed along the eastern limb of the fold in the Tony Grove basin, and a stepped pavement is found on the western limb of the fold in the White Pine basin (Fig. 7).

Why particular beds are able to support a pavement and others aren't has yet to be fully explained, but it is probable that both structural and lithologic properties are involved. Williams (1966) suggests that thickness of bedding, degree of jointing, purity, and compaction are important factors. Observations in the Tony Grove-White Pine basins tend to confirm this. The best developed pavement is found in the White Pine basin, where glaciers were able to move in the direction of dip and, thus, scoured more effectively. In the White Pine basin, the pavement is all within the lower Laketown dolomite, which has very little chert. Poor pavement development occurs in areas of jointing, such as near the change in orientation of the synclinal axis.

"Karren" refers to minor solutional sculpturing of the rock surface. Few studies have been made of dolomite karren, but notable is the work of Pluhar and Ford (1970); their comment (p. 397) that "...the overwhelming number (of karren) at all scales are developed in joints, bedding planes, or smaller lithologic features..." is also applicable to the Tony Grove-White Pine area. A number of linear, joint-controlled forms occur, ranging in size from fine etched lines to features 3 ft wide and 20 ft deep.

Much of the rock in this area is speckled by numerous small pits. These pits are commonly less than ¼ inch in diameter and depth. An examination of beds exposed in cliff faces shows that the pitting is stratigraphically controlled, apparently reflecting minor differences in lithology.

A larger scale manifestation of pitting and cavernous weathering has been labeled "spongy karren" by this Writer. Spongy karren has a number of forms, such as pinnacles, scallops, and tunnels, all on the order of one inch in size. Spongy karren is also stratigraphically controlled, but other factors, such as the turbulent flow of water, are suspected to be important in its creation.

"Kamenitzas" (solutionally formed rock basins) occur where relatively horizontal bedding planes are exposed. They range in diameter from a few inches to several feet, but are not abundant in this area.

Numerous small escarpments in the Tony Grove-White Pine area have developed smooth curved slopes from solution. The slopes have been observed with both concave and convex profiles and have been named "beveled karren" by this Writer. Water has been observed flowing over them in thin sheets, but whether this is the process that formed them or simply a result of the profile is uncertain.

The most spectacular karren of these basins are sharp pinnacles and ridges known as "spitzkarren." Isolated pinnacles 4 to 8 inches high form as pedestals beneath perched blocks. When the block is eventually dissolved or falls off, the pedestal is soon sharpened by solution. Other spitzkarren seem to form from spongy karren and kamenitzas; as these features enlarge and coalesce, they leave residual pinnacles and ridges. Narrow, blade-like spitzkarren develops where jointing is closely spaced. In a few locations, extremely thin, sharp spitzkarren were noticed. Investigation revealed these to be in areas of solution breccia, and the spitzkarren were forming on dolomite blocks within the breccia oriented so that their bedding was vertical. Differential weathering along fine laminae had produced sharp, wafer-thin ridges.

CAVES

There are approximately 20 solution features in the basins that would qualify as caves and 6 deep collapse dolines. All the caves show signs of

Figure 7. Stepped pavement in the White Pine basin. View looking north across lower two-thirds of the basin shows trees growing along solutionally-enlarged joints containing soil and small, alluvium-filled dolines (dark areas).

vadose solution, are generally quite small, and exhibit strong vertical development. Most of them are vertical fissures slightly enlarged by solution and collapse.

The stratigraphic positions of the caves are strongly influenced by the presence of chert-rich horizons which act as impermeable beds, perching ground water and leading to the creation of horizontal passages. Pits are often found immediately below the chert-rich beds, where the water was able to pass through along vertical fractures.

Cave entrances in the Tony Grove-White Pine basins are anomalous with respect to topography. Twelve of the twenty caves are on ridge summits or steep slopes, where they receive little or no runoff. Most of the other caves are elevated to some degree above the meadows and dolines and also have very little water flowing into them. The position of the entrances is evidence of the derangement of drainage and destruction of caves that resulted from glaciation.

The effects of glaciation are also visible in the areal distribution of the caves. The caves in the Tony Grove basin occur principally along the southern rim, because the central part of the basin was the main path of ice advance. As mentioned earlier, glacial erosion was particularly effective in the White Pine basin and, as a result, only two minor caves occur there.

Detrital material of apparent fluvioglacial origin is present in some of the caves. At Thundershower Cave, a quartzite boulder 3 ft in diameter and 5 ft long has been wedged into a passage in the lower part of the cave (Fig. 8). Smaller boulders and cobbles of quartzite are found in this and other caves.

Figure 8. A large quartzite boulder jammed in a cave passage, probably by glacial meltwater.

Fluvioglacial material is also associated with bodies of solution breccia in the basins. These breccia bodies probably represent solution features that were open to the surface, so that the detrital cobbles could be washed into them by glacial melt water. Subsequent weathering has released many of these quartzite boulders and cobbles onto the land surface (Fig. 9). They were no doubt originally derived from the Knight conglomerate.

The solution breccias are also associated with secondary calcite (speleothems) on the land surface. This is particularly common along the sides of what appear to be stream valleys, but which are more likely unroofed cave passages.

Figure 9. The bush at right center is at the contact between a solution breccia (to left) and normal bedding (to right). A quartzite boulder (arrow) has weathered out of the breccia.

CONCLUSIONS

The Tony Grove and White Pine basins are located where conditions of structure and climate are favorable to both glacial and solutional processes. The basins were subjected to repeated glaciation during the Pleistocene, and it seems quite likely that periods of karstification occurred during the warmer interglacials. The actual number of glaciations and the timing of the initial karstification remain subjects for further research. It may be that examination of the sedimentary material in the caves and solution breccias may help in this regard.

Existing surface landforms, such as the solution dolines and karren, are products of solutional activity in the past 12,000 years. Holocene glaciations were probably of minor importance at this altitude and latitude. Karst processes are active to some degree at the present time but are creating new conduits rather than enlarging the open remnants of pre-existing hydrologic systems.

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KARST DEVELOPMENT ON THE WHITE RIVER PLATEAU, COLORADO

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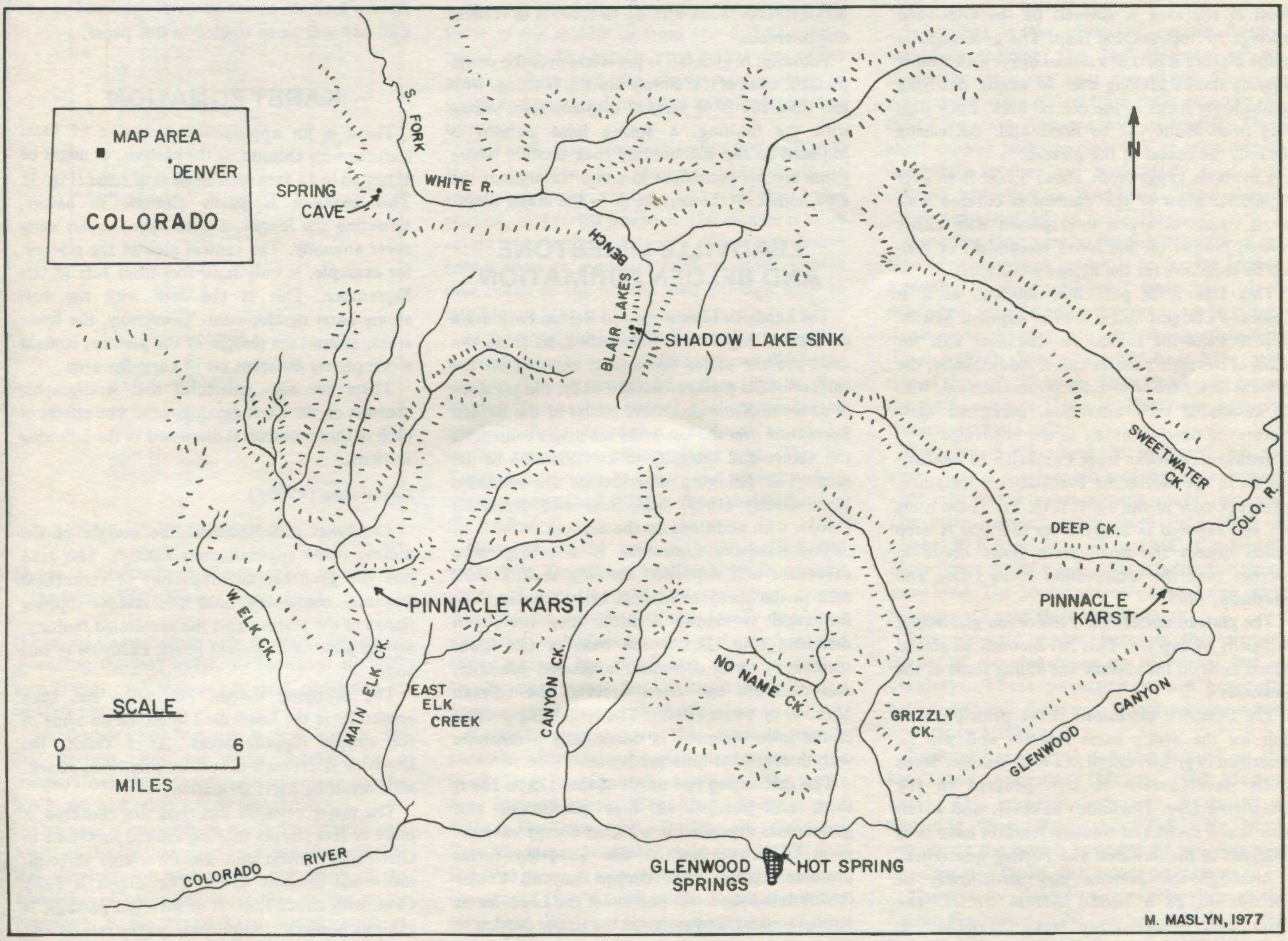

Figure 1. White River Plateau, showing the major drainage lines and some of the karst features.

ABSTRACT

The White River Plateau in northwestern Colorado is a broad, 30 by 40 mile wide, domal uplift which underwent Pleistocene glaciation, primarily by cap ice. Mississippian-age Leadville Limestone is exposed over large areas of the plateau and possesses a wide variety of karst features, including the state's four largest caves. There is an approximate zonation of the karst features by altitude: areas below 7000 ft exhibit minor karren; from 7000 to 10,000 ft, the major caves are developed; areas above 10,000 ft contain well developed alpine karst. The latter two divisions reflect the presence or absence, respectively, of a protective cap of Belden Shale over the Leadville. Where the shale is present, major caves are developed. Where the shale has been removed, in the central area of the plateau, alpine karst exists.

In the latter area, sinkholes are abundant and reach one-half mile in length. Dry valleys up to one mile in length are also present, with solution shafts up to 90 ft deep along their sides. Throughout the central area, surface drainage is quickly diverted underground to joint-controlled caves containing stairstep drops and passages. These caves span much of the 200-ft thickness of the Leadville. The water resurges as canyon-side springs in either the lower Leadville or the underlying Dyer Dolomite. Smaller-scale solution features are also present in the area. Locally, modern solution has exhumed red-siltstone-filled Mississippian-age solution features as well. Faulting has locally influenced the development of the karst features.

INTRODUCTION

THE WHITE RIVER PLATEAU is a 30 by 40 mi wide uplift located approximately 120 mi west of Denver in northwestern Colorado (Fig. 1). Most of the area is covered by the Glenwood Springs 30' topographic map. The general form of the plateau is that of a domal uplift with steeply dipping flanks arching over to nearly flat-lying sedimentary rocks in the central area. Rock dips vary from about 40° to horizontal, decreasing towards the center of the plateau.

Elevations range from about 5,500 ft to over 11,000 ft. Most of the plateau is covered with small stands of aspen interspersed with either twisted juniper (at the lower elevations) or lush alpine meadows (at the higher elevations).

This area is of particular interest, as it is Colorado's largest single area of exposed Mississippian Leadville Limestone, the host unit for many of the state's major caves. Additionally, the general lack here of metallic mineralization, with its associated rock alteration, permitted karst features to develop widely in the Leadville.

Sedimentary rocks from Cambrian to Pennsylvanian in age overlie the Precambrian metamorphic rock core of the uplift (Fig. 2). These units are well exposed in canyons up to 2,000 ft deep which dissect the area. Lava flows occur in patches over the sedimentary rocks (Bass and Northrup, 1963).

The plateau underwent Pleistocene glaciation, primarily by cap ice. This left mounds of glacial debris both on the surface and filling some of the sinkholes.

The Leadville Limestone is the principal host unit for the area's karst features and will be described in greater detail in a later section. Some karst development is also present in the underlying Dyer Dolomite, however, and a few anomalous shafts and solution features have been observed in the Sawatch and Parting quartzites.

Although the plateau may structurally be considered as a broad domal uplift, the description breaks down when examined in detail. Dips along the flanks are steep, and the transition from arching to nearly flat-lying beds occurs in distances as short as a few miles. Abrupt changes in rock dip are accomplished through a series of east-west trending step faults and a few major reverse faults with up to 1,400 ft of vertical displacement.

Faulting, in general, is prevalent over the entire plateau; most of it is simple dip-slip faulting, with less than 100 ft of vertical displacement. Along with the faulting, a strong joint pattern is imposed on the sedimentary rock section. These joints are not consistent in trend throughout the area, either on the surface or in the major caves.

LEADVILLE LIMESTONE AND BELDEN FORMATION

The Leadville Limestone and Belden Formation are treated together in this section, as these two units are the major formations exposed on the surface of the plateau. Additionally, the presence or absence of the protective shales of the Belden Formation over the Leadville is a major control on the nature and intensity of karstification in the area. The following descriptions are modified from Conley (1972) and Bass and Northrup (1963), with additions by the authors.

The Leadville Limestone is a transgressive carbonate unit deposited across a shallow shelf area in the Early to Middle Mississippian. The formation is readily divisible into two main members (Fig. 2). At the base of the lower member is sandy dolomite, laminated dolomite, and dolomite and chert breccia, the Gilman Member of Tweto (1949). The remaining portion of the lower member is dominantly a dolomite with interbedded lime mudstone.

The succeeding two-thirds of this 175 to 225 ft thick unit consists of lime packstones and grainstones deposited in a higher energy environment. This portion of the Leadville forms prominent cliffs in the deeper canyons. Conley (1972) dated the lower portion of the Leadville as Kinderhookian and assigned the upper portion to the Osageian, or early-to-middle Mississippian.

The Leadville and its equivalents were uplifted over much of the Rocky Mountain area in mid- to late-Mississippian time. Karst features developed in the unit at this time and are locally preserved. In the Aspen area, 40 mi south of the plateau, paleokarst sinkholes up to 1,500 ft in length and 250 ft in depth are found (Maslyn, 1976). Fossil karst towers are exhumed along a short section of Deep Creek in the southeastern portion of the plateau. Generally however, fossil karst is not as well developed on the plateau and is exposed only in some of the cavern roofs and as exhumed fillings in modern sinkholes.

A 1- to 25-ft thick residual terra rosa soil developed on the emergent Leadville at this time and is now preserved as the dark-red Molas Formation (shale, siltstone, and conglomerate). The Molas fills some of the paleokarst sinkholes, such as at Karst Cave. Directly overlying the Molas is the 675-ft thick Pennsylvanian Belden Formation (shale, siltstone, and sandstone). The Belden was deposited by the transgressing Early Pennsylvanian sea, which covered the area and fossilized the Late-Mississippian karst features.

For purposes of description, the Molas and Belden formations are generally grouped as one unit and will be so treated in this paper.

KARST ZONATION

There is an approximate zonation of karst features with altitude on the plateau, as might be expected in an area with 6,000 ft of relief (Fig. 3). This zonation is partly climatic in nature, reflecting the length of time spent under snow cover annually. The central area of the plateau, for example, is only snow-free from July to late September. This is the area with the best alpine karst development. Conversely, the lower areas, around the margin of the plateau, contain a completely different set of karst features.

There are also structural and stratigraphic controls on the karst development. The effects of each of these controls is discussed in the following sections.

Lower zone (<7000')

The lower zone includes the margin of the plateau below approximately 7,000 ft. This zone has the greatest development of structural features, containing both the steeply dipping flanks of the plateau and the associated faulting. Strong jointing is present in the Leadville in this zone.

The overlying Belden Formation has been eroded from the Leadville Limestone on some of the steeply dipping areas. As a result, the Leadville is exposed to surface solutional attack, with resulting karst formation.

The major caves in this zone are clustered in more or less steeply dipping faulted limestone in Glenwood Canyon and are invariably three-dimensional phreatic caves. The largest is Fairy Cave, with about 7,500 ft of surveyed passage. It exhibits phreatic iron-mineral encrustrations and the best-developed dripstone in Colorado. The

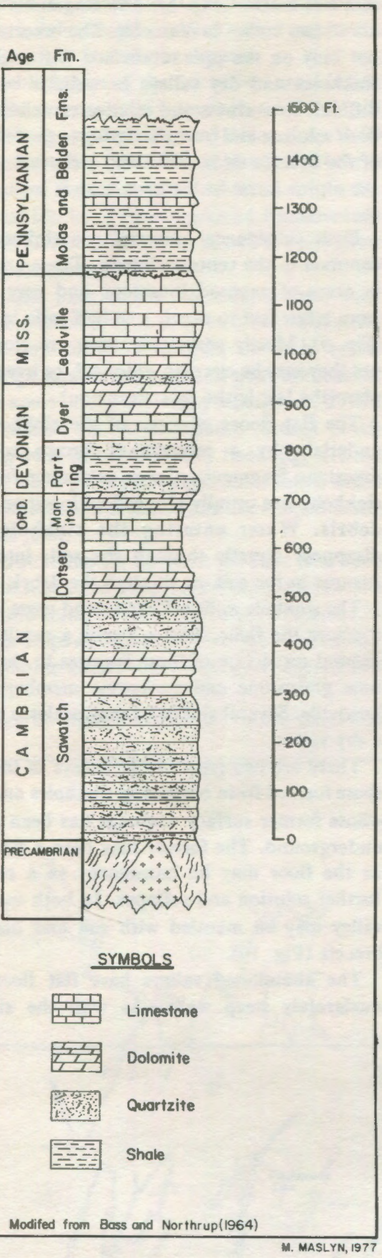

Figure 2. Generalized stratigraphic section for the White River Plateau area.

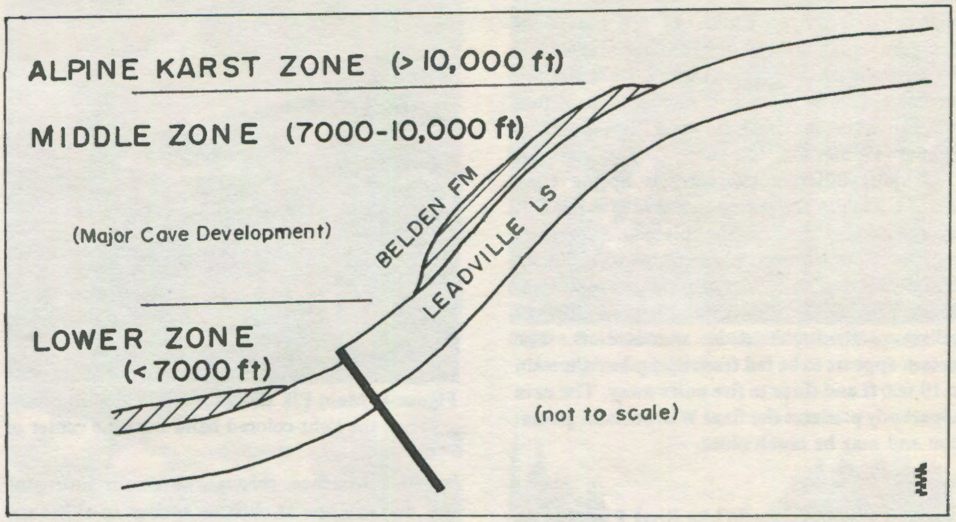

Figure 3. Karst zonation diagram (actual rock dip in the middle zone varies from approximately 2° to 40°).

Glenwood Canyon caves were probably formed by ascending thermal water, as indicated by their morphology, mineralization, and proximity to existing hot springs.

Small-scale karst features, such as karren, are also present, but are often masked by a comparatively thick regolith densely overgrown by scrub-oak and other woody vegetation.

Drainage is primarily on the surface, partly as a result of the limited outcrop areas for potential sink development, and partly as a result of the restriction of perennial water flow to the major drainages. A few major karst springs such as Twenty-Pound-Tick Cave are present, however.

7,000—10,000 ft

There is no abrupt transition either from the lower to the middle or from the middle to the upper zones. The middle zone is that area of the plateau where rock dip changes from steep to gentle and faulting diminishes in both degree and intensity.

This area, from approximately 7,000 to 10,000 ft in elevation, is characterized by alpine meadows, aspen stands, and, in the lower portion, by gnarled junipers. The climate is more severe than in the lower zone, which experiences only a short to moderate winter. Snow remains over much of the middle zone through mid- to late-May.

In this zone, the Leadville Limestone is exposed more often in the walls of deep canyons dissecting the area than on the upland surface (Fig. 4). The Belden Formation covers and protects the Leadville from aggressive solutional attack by surface waters over much of the area (Fig. 3). Faulting is present, however, and allows surface water to enter the Leadville through fractures in the Belden.

Drainage in the area is both surface and subsurface. Occasional swallow holes in exposed areas of the Leadville pirate some water underground. Fissure pits have locally developed where the Leadville is exposed at the surface, and minor karren has developed. The regolith is usually one to several feet thick and less rocky and brush covered than in the lower zone.

Karst development in this zone is mostly phreatic, with subdued surface expression, but is locally large in magnitude, as three of the state's four largest caves, Groaning, Spring, and Fixin-to-Die Caves, occur in this zone. Groaning Cave, the largest, contains more than 29,000 ft of surveyed passage, with perhaps twice that amount unsurveyed (Fig. 5). All of the above-mentioned caves have cliffside entraces near the tops of deep canyons.

The authors believe that the extensive cave development is the result of several factors, including faulting and the presence of the protective shale cap over the caves. This cap prevented surface waters from easily entering and filling the cave passages with sediment. These shale-capped phreatic caves are of considerable antiquity, as they are not clearly related to existing hydrology or landforms. Their speleothems are varied but of moderate size.

Groaning Cave may be the best example of the type of cave present, and it is by far the largest. Basically a two-dimensional maze cave, it is developed nearly horizontally in Leadville Limestone beds dipping 2½ degrees northeasterly (Davis, 1971). Several marker beds have been traced through the cave. They clearly cut across the main cave level (Jerry H. Hassemer, personal communication, 1975).

Complex development is indicated by the alternation of north-south trending high canyons with intervening anastamosing maze passages. The main canyons reach more than 60 ft in height, 30 ft in width, and over 600 ft in length. Their floors are as much as 20 ft below the main, maze level. The canyons rarely connect with other passageways near their terminations.

The phreatic mazes between the canyons consist of rounded tubes up to several feet in diameter. These passageways are generally horizontal, and their rock floors are commonly overlain by sediments.

Faulting has exerted a major control on the area's cave development (Munthe, 1969; Maslyn, 1975). This is particularly well illustrated at Groaning Cave. Here, the 800-ft wide block containing the cave is bounded by two east-west trending step faults with minor displacement.

These faults allowed water to pass through the Belden Formation to the Leadville Limestone and to some extent influenced the water flow. Many of the largest rooms in the cave are located in the vicinity of these two faults, as are the major dripstone areas. The paleohydrology responsible for the maze pattern of Groaning Cave is not well understood. The cave fits neither the diffuse infiltration or the back-flooding hypotheses of Palmer (1975).

Of quite different character is Spring Cave (Fig. 6), a major active resurgence cave at 8,000 ft on the west flank of the plateau. Complete exploration has been blocked by a series of sumps, but over two miles of passage are known. This is primarily a large, strike-oriented, phreatic gallery partly undercut by stream slots. The stream appears to be fed from updip karst terrain at 10,000 ft and three to five miles away. The cave apparently predates the final Wisconsinan glaciation and may be much older.

>10,000 ft.

Near the crest of the White River Plateau, the Leadville arches up to over 11,000 ft in elevation. Rock dips here are usually gentle, and the faulting is diminished in intensity but not in magnitude. The landscape is one of the gently rolling hills of weathered limestone, mantled by limestone fragments. This area of alpine meadows and conifers is the upper zone.

Here, the protective Belden Formation cap has been removed, leaving large areas of Leadville Limestone exposed to the harshest climate on the plateau. Winters are long, and the area is snow-free only two to three months of the year. This area also shows the greatest effects of Pleistocene glaciation. Caves here, as in glaciated alpine karsts elsewhere, are mostly youthful, actively growing, visibly related to existing landforms, and contain few and small speleothems.

Most of the major streams draining the plateau have their headwaters in this area. Drainage is largely subsurface, through numerous insurgences, and emerges at cliffside springs in either the lower Leadville or upper Dyer Dolomite.

Figure 4. Main Elk Canyon. Leadville Limestone forms the light-colored band near the center of the view.

ALPINE KARST

For the purposes of this paper, the term "alpine karst" will be applied to those solution features developed in glaciated carbonate terraines. Within this definition are karren, caves, sinkholes, dry valleys, solution shafts, and surficial accumulations of weathered fragments. Surface karst features present in the area are karren and a surface mantle of karst breccia. The karren include groove karren, cleft karren and solution runnels up to ½-inch deep. Individual cleft karren may extend several hundred feet. These features are developed where the Leadville forms either a small cliff or a limestone pavement, and large-scale karst features such as sinkholes are absent (Fig. 7).

The Leadville is everywhere mantled by a breccia composed of angular limestone fragments in a soil matrix (Fig. 8). The fragments average about two inches in diameter. The breccia occurs not only on the upland surface, but within the sinkholes and dry valleys as well. It would be difficult to evaluate the relative contributions of both solution and frost-shattering to the formation of the breccia. It is, however, a characteristic of the alpine karst in the area and is included for that reason.

Both subsidence and collapse sinkholes are common in the central plateau. These are found in areas of exposed limestone and vary in size from a few feet to nearly one-half mile in length (Fig. 9). Mostly commonly, they are elongated, but they may be circular, elliptical, or irregular in plan (the last is the most common).

The flat floors of some of the sinkholes are underlain by a mixture of brown soil and limestone fragments. Caves extending from the sinkholes are usually plugged with soil or glacial debris. Water entering the sinkholes may disappear directly through the soil, into small fissures in the soil, or through the debris.

The sinkhole walls seldom extend more than 80 ft above the floor. This is partly a result of the general restriction of these features to the upper lime grainstone and packstone member of the Leadville. Several sinkholes may coalesce to form a dry valley.

There are two types of dry valleys in the area: those formed from coalescing sinkholes and those whose former surface drainage has been pirated underground. The former may have a flat floor, or the floor may be hummocky as a result of further solution and collapse. In both cases, the valley may be mantled with soil and limestone breccia (Fig. 10).

The abandoned valleys have flat floors and moderately steep walls. As with the sinkhole

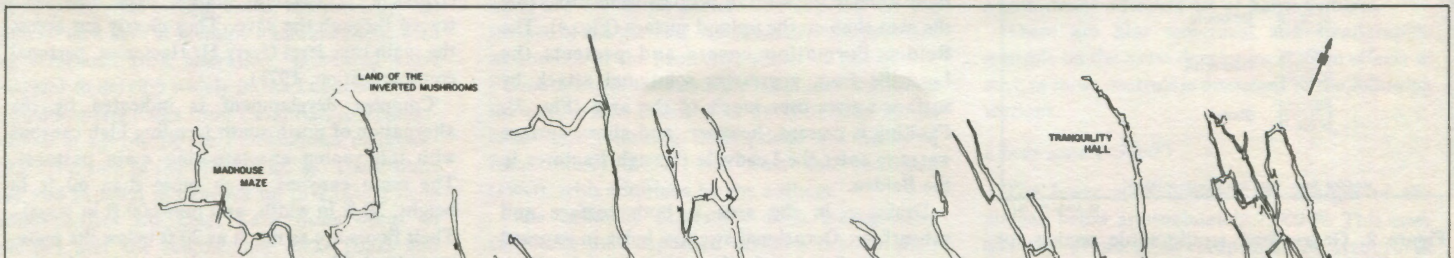

Figure 5. Groaning Cave.

valleys, they are generally restricted to the upper member of the Leadville. The valleys may reach nearly a mile in length, ending either against a bedrock wall or at a major canyon.

A notable feature of the dry valleys are solution shafts. These are entered from sinkholes developed along the edge of the valley and descend up to 90 ft below the valley floor. The entrance sinkhole is commonly a few tens of feet across and contains individual blocks of frost-shattered limestone up to six feet across. Both the sinks and the pits show strong joint-control.

From the base of the sinkhole, a shaft may continue vertically, or it may be offset to a series of stair-step drops. Examples of both types, Organ Pipe and Red Dutchman pits, are located 1,500 ft apart in one dry valley (Hassemer, 1975).

The former is a rounded, vertical shaft plunging 64 ft below the valley floor (fig. 11). At the base of the pit, a short slope leads into a dome slightly offset from the shaft, but developed along the same joint.

Red Dutchman Pit opens in the base of a 40-ft -long sinkhole containing large blocks of limestone which have slumped into it. The pit is a series of stair-step drops along narrow fissures, terminating in a moderate-sized room 90 ft below the surface (Fig. 12).

Neither of these two pits has much enterable

Figure 7. Karren on Leadville Limestone near the center of the plateau.

Figure 9. Sinkhole about 15 ft in diameter in the alpine karst zone.

Figure 8. Surficial limestone breccia.

Figure 10 Dry valley near Johnson Creek.

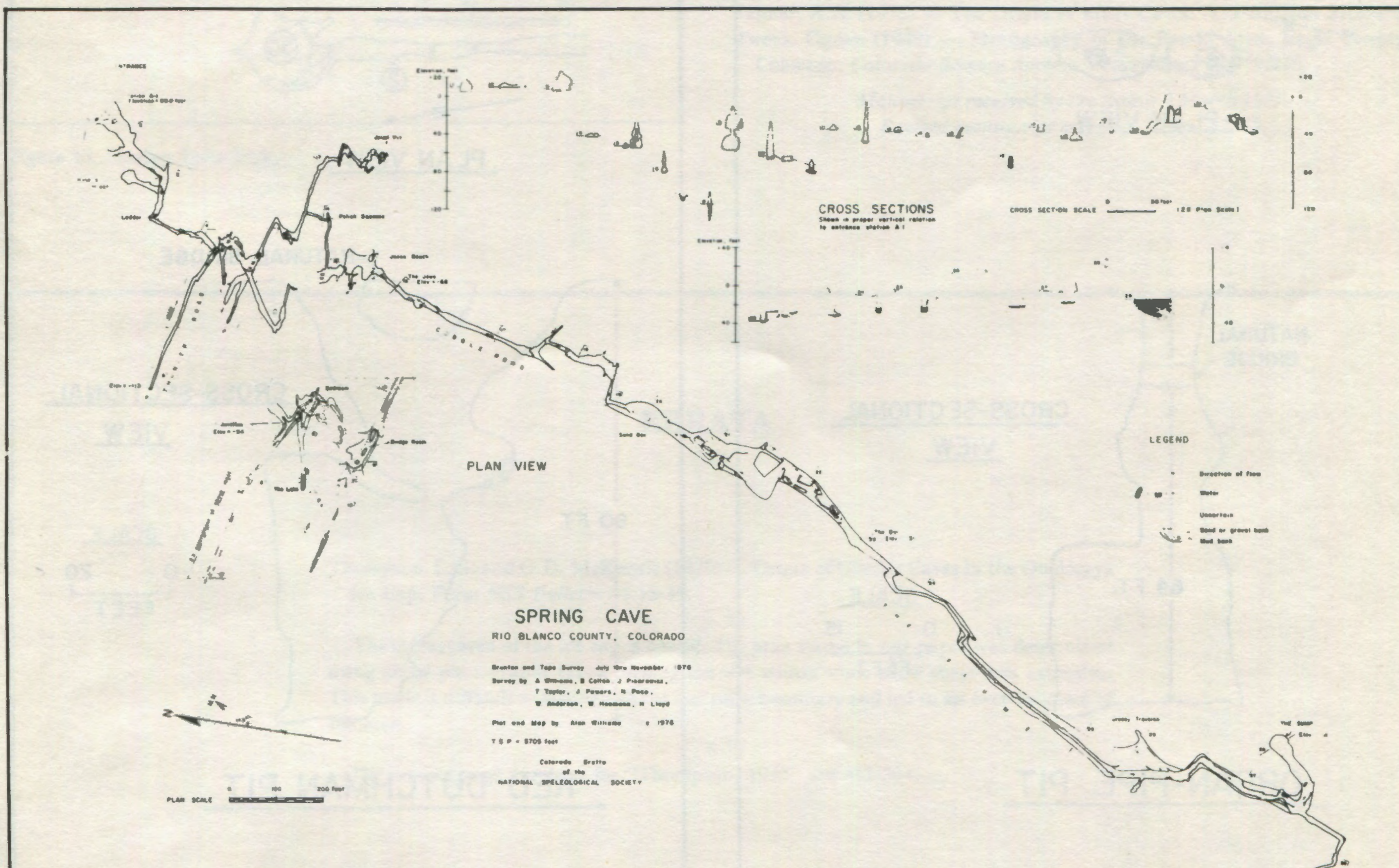

Figure 6. Spring Cave (Republication permission granted to The National Speleogical Society, Inc.).

passage extending out of the bottom. Water, however, escapes either through the surface fill washed into the pits or through small bedrock outlets at the base of the pit.

Campbell (1973) has suggested that similar features in Montana developed beneath long-lasting snow drifts. The authors feel, however, that in this area other factors have caused the pits to form at or near the break in slope between valley walls and floor.

STRUCTURAL INFLUENCE ON ALPINE KARST DEVELOPMENT

All of the alpine karst features thus far described have been in gently dipping beds. There is one highly deformed area in the central portion of the plateau which merits treatment, however. This is the Blair Lakes Bench (Fig. 1), a 6-mi-long by ⅓-mi-wide area of Leadville Limestone down-dropped against Precambrian metamorphic and Lower Paleozoic sedimentary rocks along the Blair Lakes reverse fault.

This area is notable for structural control of both the location and the magnitude of karst features. Tight drag folding along the Blair Lakes fault has brought the Leadville vertically against insoluble rocks at the base of Blair Mountain.

Water flowing across these rocks enters sinkholes in the upturned edges of the Leadville, traverses the unit, and resurges as cliffside springs in gently dipping beds of the same formation a short distance to the east. Shadow Lake Sink (Fig. 13) is an example of this type of cave development.

A surface stream flowing east out of Shadow Lake sinks at the base of a 30-ft -high wall of Leadville Limestone a short distance beyond the Blair Lakes fault. A 2-ft-high, cobble-strewn, vadose crawlway leads from the entrance to the top of the first of a series of stair-step drops in highly joint-controlled passageways. The stream traverses approximately 160 vertical ft of Leadville Limestone before it disappears into a fissure at the base of the last drop.

The depth to which the cave developed is in large part the result of lithologic differences. The stream continues to dissolve deeper into the unit, until it reaches a less permeable bed and is directed laterally to eventually emerge as a cliffside spring in the lower portion of the Leadville. Other, similar, resurgences are located in both the lower Leadville and the upper part of the underlying Dyer Dolomite.

SUMMARY

Karst development on the White River Plateau occurs in three zones determined by the combined influences of altitude and climate, structure, and the presence or absence of the protective Belden Formation above the Leadville Limestone.

In the lower zone, below 7,000 ft, the Belden is locally absent and karren has developed. In Glenwood Canyon, this zone contains clusters of phreatic caves formed by ascending thermal water.

The protective Belden Formation is often present in the middle zone, from approximately 7,000 to 10,000 ft. This zone contains most of the major caves. These caves show a close relationship to nearby faults and are, mostly, drained phreatic systems. Some resurgence caves are present.

The upper zone, above 10,000 ft, contains large areas of exposed Leadville Limestone. This area shows the greatest effects of Pleistocene glaciation on the plateau and contains well-developed alpine karst features. These include youthful caves, sinkholes, dry valleys, solution shafts, karren, and a mantling limestone breccia. Drag folding of the Leadville along the Blair Lakes fault in this area has locally enhanced vertical cave development.

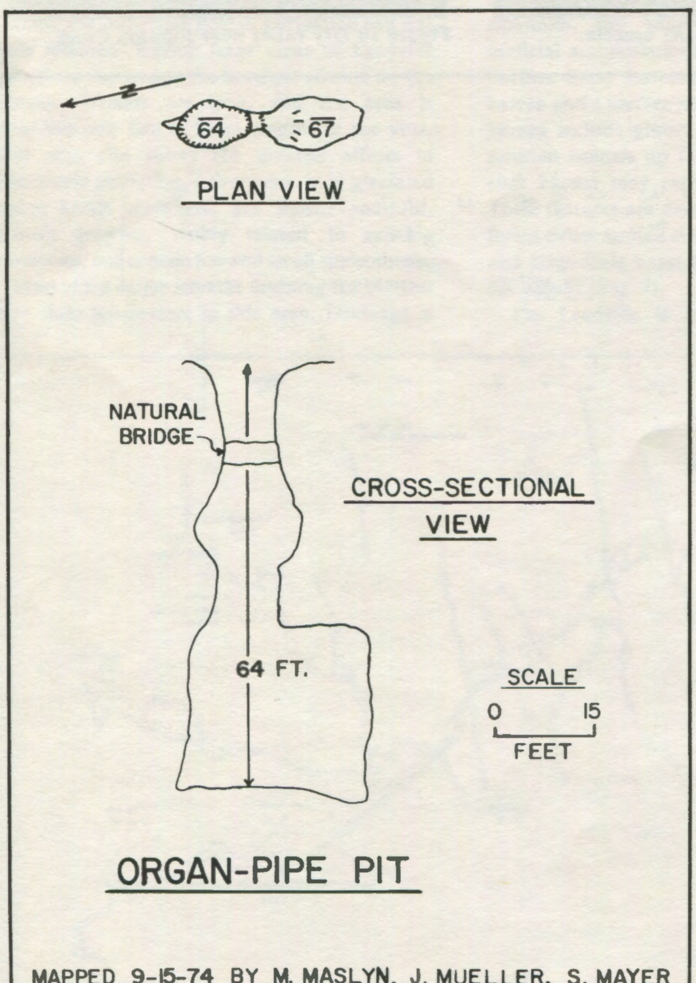

Figure 11. Organ Pipe Pit.

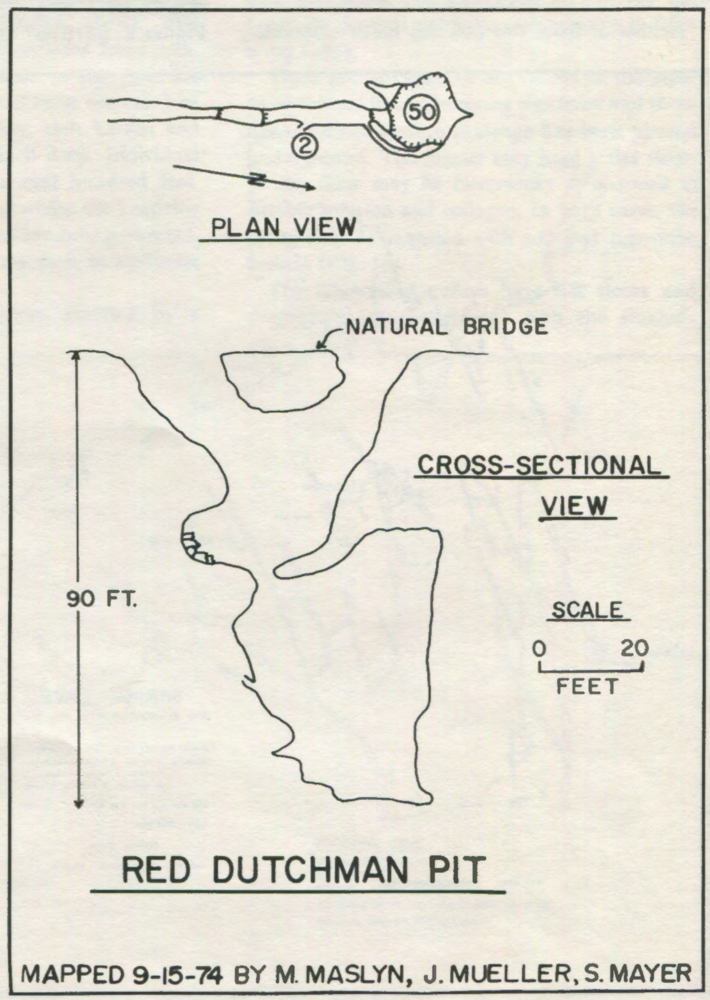

Figure 12. Red Dutchman Pit.

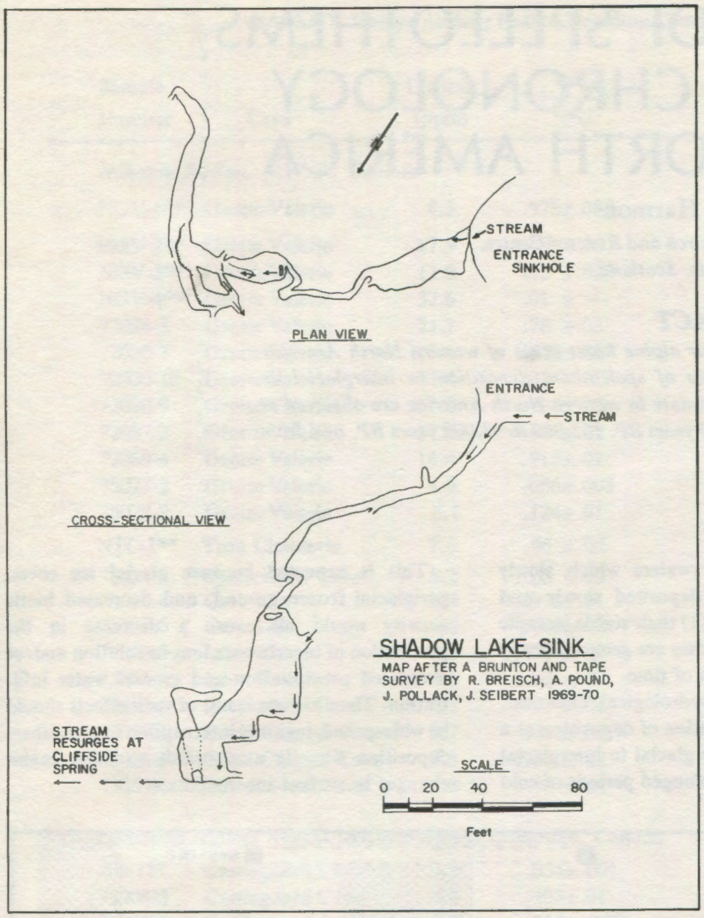

Figure 13. Shadow Lake Sink.

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ERRATA

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p. 16

—The correct area of the ice cap is 55km². The area stated in our paper was determined using aerial photographs taken turning the wet season when snow cover was extensive. This made it difficult to distinguish the ice/snow boundary and led to an over-estimate of the area.

p.19

—The correct page numbers for "Thompson, 1975" are 351-364.

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U-SERIES DATING OF SPELEOTHEMS AND A-GLACIAL CHRONOLOGY FOR WESTERN NORTH AMERICA

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ABSTRACT

$^{230}Th/^{234}U$ ages for calcite speleothems from four alpine karst areas of western North America cluster into four distinct groups. Attributing periods of speleothem deposition to interglacial or interstadial episodes and vice-versa, times of warm climate in western North America are observed at about 320,000 to 280,000 years BP, 220,000 to 180,000 years BP, 155,000 to 99,000 years BP, and from 10,000 years BP to the present.

INTRODUCTION

CALCIUM CARBONATE SPELEOTHEMS are precipitated from ground waters which slowly percolate through cavernous carbonate rock. These deposits tend to be deposited slowly and continuously over their period of growth and are of paleo-climatic interest because (1) their stable isotopic composition is a reflection of the climatic conditions under which they formed, (2) they are geographically widespread, and (3) they tend to be preserved without alteration for long periods of time.

The deposition of speleothems in caves is dependent on a variety of geological, hydrological, chemical, and climatic factors. A change in any one of these factors could cause random cessation of deposition at a single drip site unrelated to climate change. However, in alpine areas of extreme glacial to interglacial climatic contrast, deposition of speleothems would be expected to cease during prolonged periods of cold climate. This is expected because glacial ice cover, periglacial frozen ground, and decreased biotic activity would all cause a decrease in the production of bicarbonate ions in solution and/or decreased precipitation and ground water infiltration. The ultimate result of such effects should be widespread, regional interruption in speleothem deposition directly attributable to these major changes in surface macroclimate.

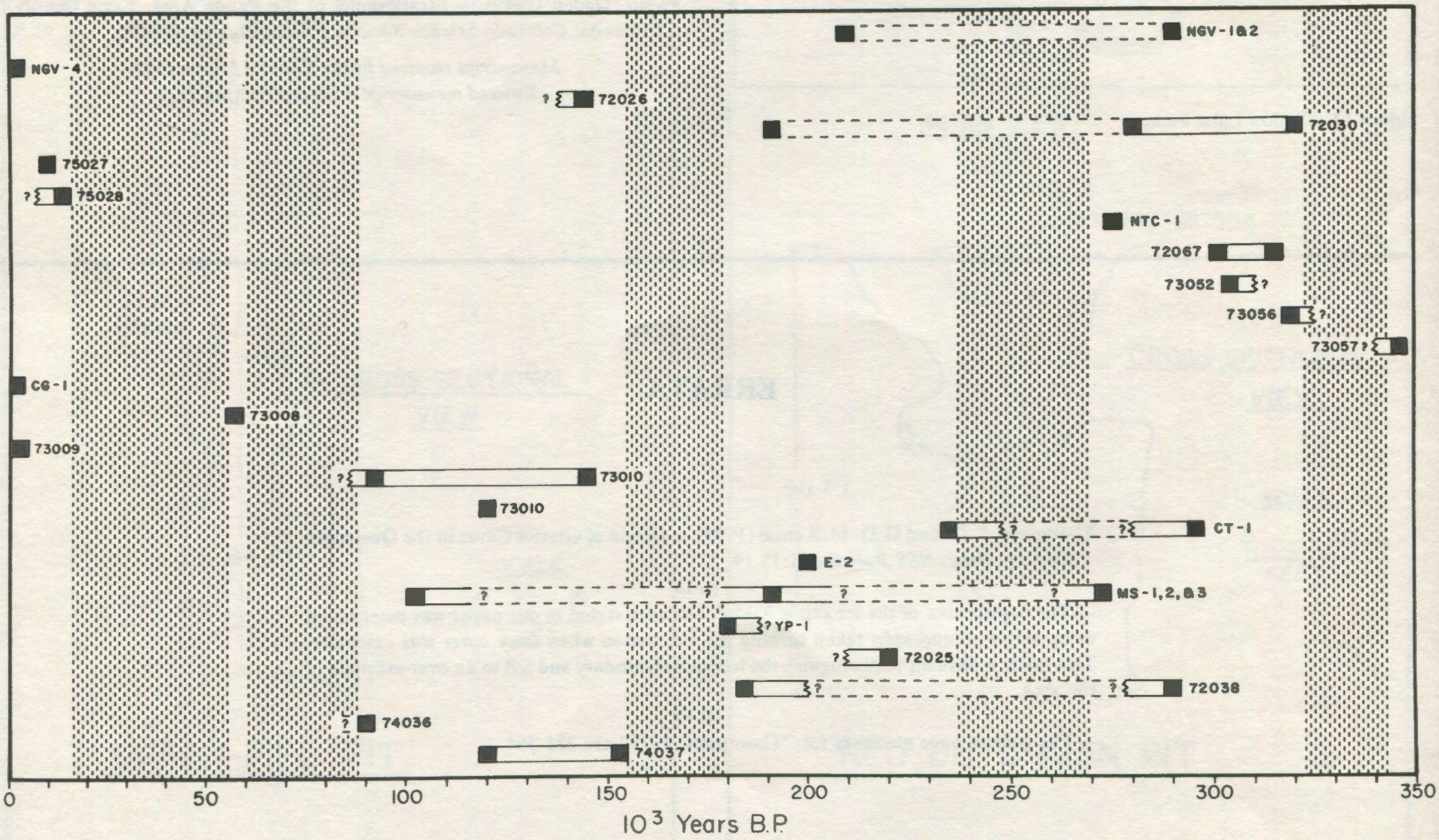

Figure 1. Depositional history of the alpine speleothems dated in this study. Inferred periods of glacial cold climate are indicated by the stippling. See text for discussion.

TABLE 1. Isotope Activity Ratios, Uranium Concentrations, and Calculated Ages for Alpine Speleothems

Sample Number	Cave	U conc. (ppm)	$\{\frac{^{230}Th}{^{234}U}\}$	$\{\frac{^{234}U}{^{238}U}\}$	$\{\frac{^{230}Th}{^{232}Th}\}$	Age (x 10^3 years B.P.)	Comments
Nahanni Region, N.W.T., Canada							
NGV-1**	Grotte Valerie	4.3	.975±.085	.98 ±.01	>1000	290^{+00}_{-50}	stalagmite—top of inner layer
NGV-2**	Grotte Valerie	17.9	.84 ±.03	1.01 ±.005	450	200±20	stalagmite—bottom of outer layer
NGV-3**	Grotte Valerie	12.9	.85 ±.025	1.045±.006	300	217±17	stalagmite—top of outer layer
NGV-4**	Grotte Valerie	32.6	.01 ± —	.97 ±.01	26	<2	stalactite—"modern"
72026-3	Grotte Valerie	21.1	.78 ±.03	1.40 ±.02	137	145±16	stalagmite—basal layer
72030-7	Grotte Valerie	19.6	.825±.02	.98 ±.01	>1000	191±21	flowstone—top layer
72030-10	Grotte Valerie	33.4	.92 ±.02	.92 ±.01	145	280±27	flowstone—top of basal layer
72030-9	Grotte Valerie	60.2	.965±.02	.94 ±.01	86	320±35	flowstone—bottom of basal layer
72057-2	Grotte Valerie	17.8	1.05 ±.02	1.03 ±.01	71	>350	flowstone—middle layer
72066-4	Grotte Valerie	18.4	.915±.02	.97 ±.01	90	278±27	flowstone—top layer
75027-2	Grotte Valerie	4.9	.086±.005	.94 ±.02	>1000	10±0.5	stalagmite—bulk sample
75028-2	Grotte Valerie	5.1	.124±.01	.97 ±.02	10	15±2	stalagmite—basal layer
NTC-1**	Trou Claudette	7.6	.96 ±.02	.99 ±.01	250	275^{+75}_{-35}	stalactite—bulk sample
72034-1	Speleothem Cave	5.2	1.03 ±.02	1.04 ±.01	96	>350	flowstone—top layer
72067-4	Cave 12A	4.9	.975±.01	.84 ±.02	78	301±24	stalagmite—top layer
72067-5	Cave 12A	4.6	.965±.02	.88 ±.01	97	315±41	stalagmite—basal layer
73051-5	Igloo Cave	3.3	1.03 ±.02	1.02 ±.01	120	>350	stalagmite—top layer
73051-4	Igloo Cave	4.1	1.02 ±.02	1.00 ±.01	90	>350	stalagmite—basal layer
73052-6	Ice Curtain Cave	6.7	.93 ±.01	.97 ±.01	178	304±22	flowstone—top layer
73056-1	Coral Canyon Cave	0.7	.955±.02	1.03 ±.02	>1000	319±32	flowstone—top layer
73056-3	Coral Canyon Cave	0.8	1.05 ±.02	1.07 ±.02	20	>350	flowstone—basal layer
73057-1	Tower View Cave	51.8	.98 ±.01	1.08 ±.01	231	346^{+00}_{-26}	flowstone—basal layer
Columbia Icefield Region, Alberta-British Columbia, Canada							
CG-1**	Castleguard Cave	3.2	.034±.001	1.21 ±.014	64	4±0.2	stalactite—"recent"
73008-3	Castleguard Cave	2.6	.405±.01	1.08 ±.02	42	57± 2	flowstone—bulk sample
73009-3	Castleguard Cave	5.9	.012±.005	1.33 ±.02	>1000	1±0.5	stalagmite—top layer
73009-4	Castleguard Cave	19.1	.024±.001	1.34 ±.02	35	3±0.2	stalagmite—basal layer
73010-7	Castleguard Cave	2.5	.58 ±.01	1.09 ±.01	26	92± 3	stalagmite—top layer
73010-6	Castleguard Cave	2.5	.75 ±.03	1.08 ±.02	159	147±12	stalagmite—basal layer
73011-2	Castleguard Cave	8.7	.69 ±.02	1.34 ±.02	68	120± 6	stalagmite—bulk sample
Crowsnest Pass Area, Alberta-British Columbia, Canada							
CT-1**	Coulthard Cave	0.2	.862±.026	1.00 ±.02	99	235± 9	stalactite—outer layer
CT-2**	Coulthard Cave	0.25	.925±.029	1.07 ±.02	162	296±32	stalactite—inner layer
E-2**	Eagle Cave	0.3	.869±.026	1.40 ±.02	46	198±13	flowstone—bulk sample
MS-1**	Middle Sentry	0.12	.595±.024***	1.10 ±.03	10	102± 6	flowstone—top layer
MS-2**	Middle Sentry	0.07	.808±.032***	1.07 ±.04	24	191±16	flowstone—middle layers
MS-3**	Middle Sentry	—	.873±.038	.97 ±.03	18	273±13	flowstone—basal layers
YP-1**	Yorkshire Pot	3.1	.756±.028	1.21 ±.015	402	178±10	stalagmite—top layer
GAR-1**	Gargantua Cave	3.7	.993±.016	1.07 ±.015	141	>350	stalactite—inner layer
72025-4	Gargantua Cave	18.1	.85 ±.02	.94 ±.05	56	219±16	stalactite—inner layer
72038-1	Gargantua Cave	8.9	.88 ±.05	1.55 ±.02	59	184±22	stalagmite—top layer
72038-2	Gargantua Cave	38.1	.91 ±.04	.92 ±.01	277	290^{+60}_{-45}	stalagmite—basal layer
69001-6	Middle Caves	2.5	1.04 ±.03	.98 ±.03	45	>350	stalagmite—top layer
Bear River Range—Uinta Region, Utah, U.S.A.							
74033-1	178' Pit	0.15	1.15 ±.10	1.43 ±.07	>1000	>350	flowstone—bulk sample
74034-2	Surface	1.16	1.43 ±.24	1.025±.06	84	>350	flowstone—top layer
74034-1	Surface	0.18	1.07 ±.07	1.07 ±.05	20	>350	flowstone—basal layer
74036-1	Logan Cave	0.15	.59 ±.03	2.57 ±.03	>1000	90± 5	flowstone—basal layers
74037-1	Porcupine Cave	0.17	.69 ±.04	1.38 ±.09	>1000	120±12	flowstone—top layers
74037-2	Porcupine Cave	0.12	.78 ±.08	1.12 ±.09	>1000	153±34	flowstone—basal layers
74038-1	White Rocks Cave	1.11	1.06 ±.03	1.54 ±.03	66	>350	stalactite—bulk sample
74040-1	White Rocks Cave	0.62	1.07 ±.03	1.13 ±.02	>1000	>350	flowstone—middle layer

* { } denotes isotope activity ratios ** data from P. Thompson (1973) *** corrected for detrital Th assuming $\{^{230}Th/^{232}Th\}_o = 1.7$

Four alpine areas in Canada and the United States in which significant glacial/interglacial climate contrast could be expected were sampled. From north to south, the areas are: (1) the Nahanni Plateau Region of the Mackenzie Mountains, N.W.T. (61°31'N, 124°W), (2) the Columbia Icefield Region of the Rocky Mountains, Alta.-B.C. (52°N, 117'W), and (3) The Crow's Nest Pass Area of the Rocky Mountains, Alta.-B.C. (49°30'N, 114°30'W), and (4) the Bear River Range Uinta Region of the Southern Rocky Mountains (41°N, 111°W). All of these regions have mean annual surface temperatures within a few C° of freezing, with only a minor, limited amount of speleothem deposition occurring at present. It would thus seem likely that any worldwide change from interglacial to glacial climate, such as has occurred at least ten times in the past two million years (Emiliani and Shackleton, 1974), should result in concurrent interruption or cessation of speleothem deposition within the study areas.

ANALYTICAL METHODS

Twenty-five calcite speleothems showing no physical signs of recrystallization or diagenetic alteration were dated by the $^{230}Th/^{234}U$ method. Between 10 and 100 grams of speleothem calcite were dissolved in about 1l of $2MHNO_3$, spiked with a $^{232}U/^{228}Th$ tracer, an Fe^{3+} carrier added, and filtered if any detrital material was present. After boiling, U and Th were co-precipitated with $Fe(OH)_3$. The precipitate was washed, aged, and then dissolved in 9 *M* HCL. U was separated from Th by ion exchange of Dowex 1X-8 resin in a chloride form and Th subsequently separated from other contaminates by ion exchange on the same resin in a nitrate form. After elution from the resins with 0.1 *M* HCl, the U and Th solutions were evaporated to dryness. Further isolation of Th and U was accomplished by double extraction into a 0.2 *M* TTA/benzene solution at pH's of 1.2 and 3.5, respectively. Finally, the purified U and Th samples were plated from the organic onto stainless steel discs for alpha counting.

U concentrations, isotope activity ratios, and calculated ages are given in Table 1. A total of 35 age determinations were made on 25 samples. Also included in the data set are 14 analyses for 9 samples previously reported in Ford, *et al.* (1972) and in Ford (1973). All samples except MS-1 and MS-2 contained less than 1% insoluble residue. U contrations ranged from 0.07 to 60.2 ppm. $^{230}Th/^{232}Th$ ratios were generally high (>20), indicating an initial absence of detrital Th. However, 12 samples had $^{230}Th/^{234}U$ ratios of unity or greater, having been deposited before 350,000 years B.P. and subsequently having attained a state of isotopic equilibrium. The final data set for the study thus consisted of 37 analyses for 26 different speleothems from the four study areas.

DISCUSSION

The data from Table 1 have been plotted as a function of time in Figure 1. From the figure, it is seen that the 37 speleothem ages cluster into four distinct groups. That there is little overlap between age uncertainties among the groups suggests that there is a real intervening time break between them.

Interpreting the four depositional episodes in Figure 1 to correspond to major periods of continental interglacial climate, such as that at present, and vice-versa, a general picture of past climatic change within western North America during the past 350,000 years can be obtained. The individual age determinations have been plotted in Figure 1 as filled squares. Solid lines connect filled squares where textural evidence indicates continuous speleothem growth between two or more dated horizons within a single specimen. Dashed lines indicate interrupted growth characterized by surface alteration and/or erosion of the deposit between dated horizons. Jagged terminations mean that the complete record of growth of a speleothem is not known. The periods of speleothem deposition within the study areas occur from (1) the present to about 15,000 years BP, (2) 90,000 to 155,000 years BP, (3) 180,000 to 240,000 years BP, and (4), 275,000 to 320,000 years BP. Single dates at 60,000 and 345,000 years represent interstadial events within these major cold periods.

It is important to note that the observed grouping of the age data is consistent with the physical characteristics and morphology of the specimens studied. The 11 analyses within the 275 to 320 thousand year group all belong to what is identifiable as an ancient, massive phase of deposition throughout the four areas sampled. Likewise, four of the analyses in the 180 to 220 thousand year group are overgrowths on specimens assigned to the former group, renewed deposition having occurred after a depositional hiatus during which the original, exposed surfaces of the older deposits were deeply weathered.

The climatic cycles within the past 350,000 years observed in this study agree well with other generally accepted records also dated by U-series techniques. Interglacial periods of high sea stand characterized by intense coral reef growth are observed in Barbados and New Guinea at 350,000 to 300,000 years BP, 230,000 to 190,000 years BP, 145,000 to 80,000 years BP, at 60,000 years BP, and from 10,000 years BP to the present. (Mesolella, *et al.*, 1969; Veeh and Chappell, 1970; Bender, *et al.*, 1973; Bloom, *et al.*, 1974). Likewise, periods of maximum ^{18}O depletion in the marine foraminiferal isotope record taken to be indicative of interglacial climatic conditions are observed at about 340,000, 270,000, 222,000, 190,000, 125,000, 105,000, 82,000 and 10,000 years BP (Broecker and VanDonk, 1970; Shackleton and Opdyke, 1973). Within the analytical uncertainty in age, these marine interglacial events are synchronous with the interglacial periods of speleothem deposition recognized in this study.

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It is occasionally necessary to replace members of the Board of Editors. In order to ensure that no interested party may be overlooked at these times, and to facilitate our obtaining referees for papers submitted, everyone interested in being considered for a place on the Board of Editors is invited to forward a paragraph or two [NO resumes] describing their background and interests to the Managing Editor. This will help us avoid making the Bulletin the exclusive domain of an "invisible college" of buddies; it will also help induct beginning scholars into the system.

Those who have submitted their names previously are still on record and need not re-apply.

Even if you are unable to participate in the operation of the Bulletin, we would appreciate having your suggestions and criticisms. Many of the changes made by the present staff were suggested to us by readers.

INFORMATION FOR CONTRIBUTORS TO *THE NSS BULLETIN*

Papers discussing any aspect of speleology are considered for publication in *The NSS Bulletin*. We particularly welcome articles describing important caves and cave areas, on the history of caves and of speleology, on problems and techniques of cave conservation, and critical reviews of current literature, in addition to papers on the more traditional subjects of cave geology, geography, anthropology, fauna, and ecology. The material presented must be original and of lasting interest. Authors should demonstrate the significance of their work to speleological theory and should elucidate the historical antecedents of their interpretations by reference to appropriate literature. Presentations consisting of raw data, only, will not be accepted.

A narrative style of writing is preferred. Fine prose is terse yet free from lacunae, sparkles without dazzling, and achieves splendor without ostentation. Data and interpretations blend effortlessly along a logical continuum so that the reader, having read, neither knows nor cares how many pages he may have turned while following the author's exposition.

As written language must communicate through time as well as across space, neologisms should be introduced only if needed to express new concepts or to record new percepts. Standard usage, therefore, is required of all authors. For general style, refer to papers in this *Bulletin* and to the following handbooks: "Suggestions to Authors" (U. S. Geological Survey), "Style Manual for Biological Journals" (American Institute of Biological Sciences, Washington, D. C.), and "A Manual of Style" (The University of Chicago Press).

Articles on earth sciences (including pseudokarst), life sciences, conservation, social science (including history), and exploration should be sent directly to the appropriate specialist on the Board of Editors (see masthead); articles not clearly falling into any of those categories may be sent to the Managing Editor. Potential contributors, especially those not professional scientists or writers, are invited to consult with the editors for guidance or aid in the presentation of their material.

Two double-spaced, typewritten copies of each manuscript, including all illustrations, are required. Manuscripts should not exceed about 10,000 words in length (approximately 40 pages of typescript), although this limit may be waived when a paper has unusual merit. Photographs must be sharp, high in contrast, and printed on glossy paper. All line drawings should be neatly rendered in "india" ink or its equivalent; the smallest lettering must be at least 2 mm high after reduction. Typed lettering is not satisfactory. Captions will be set in type and added in proof. The dimensions of original drawings and of cropped photographs should be made some multiple of the length and width of a column or of a page, when possible, in order to avoid problems with the layout. In case of doubt regarding length or illustrations, consult with the editors.

Abstracts are required of all papers; these must be brief and must summarize the author's discoveries and conclusions, not merely tell what he did. Captions are required for all illustrations. All unusual symbols must be defined. Authors should give their institutional affiliation (if any) and address exactly as they are to appear in print. Direct quotations from non-English language sources should be given in the original languages, with English translations (if desired) in footnotes. References to the literature must be by author and date, with specific pages where desirable. Literature cited must be listed in an end bibliography, with entries arranged alphabetically by the author's surname, typed in the format employed in this *Bulletin*. References must contain all information necessary for locating them, with titles and journal names completely spelled out in their original language and including all diacritical marks. Inclusive page numbers of articles and the total number of pages of books must be given. All persons to whom "personal communications" are attributed should be named in the bibliography and a current address provided for each.

Contributed papers will be refereed by one or more authorities in the appropriate specialty and will be edited for style before publication. After being refereed and again after being edited, papers will be returned to the authors for inspection and for any revisions which may be necessary. Please enclose a self-addressed, stamped envelope for the return of your manuscript.

By act of the Board of Governors of the NSS (#81-277, dated 8-12-74), a charge of not less than $25 per printed page will be levied against the author's institution or other funding agency after a paper has been refereed, edited, and accepted for publication. Payment will not be expected of scholars whose research was not sponsored or whose budgets do not include money earmarked to subsidize publication. In no event, will the ability to pay page charges be discussed until after final acceptance of a manuscript.

Reprints may be ordered when galleys are returned by the authors to the Managing Editor; these will be supplied at cost.

Summary: (1) data and/or interpretations must be original; (2) use a narrative style of writing; (3) follow standard English usage; (4) do not exceed 10,000 words (40 double-spaced pages of typescript) without receiving permission from the Managing Editor in advance) (5) submit two complete copies, including abstracts and all illustrations; (6) enclose a self-addressed, stamped envelope for the return of your manuscript.

TECHNICAL NOTE

Many authors' manuscripts are delayed in publication because the photographs submitted as illustrations are not of acceptable quality. Most problems could be avoided by careful darkroom work or by patronizing laboratories experienced in technical photography. These guidelines should be kept in mind:

All photographs submitted for publication *must* be in black-and-white; color prints and color negatives or slides will be converted to black-and-white at the author's expense, or returned. Black-and-white prints made from black-and-white negatives *must* be on black-and-white paper—many commercial laboratories now print black-and-white negatives on color paper because there is so little demand for black-and-white work by the general public, but this results in a gray picture lacking in definition. Color negatives should be printed on Kodak "Panalure II RC" paper or a similar material. Color slides (positive transparencies) *must* have an internegative (a color-film negative) made; the internegative is then printed on "Panalure II RC" paper.

Color conversion costs about $6 per picture. We strongly recommend that, when taking photographs for possible future publication, you carry two cameras. Load one with a slow-speed, fine-grained black-and-white film, such as Kodak "Panatomic-X," and the other with the color film of your choice. If only one camera is available, use a medium-speed, fine-grained color negative film, such as "Kodacolor" film. This will yield good black-and-white prints as well as good color prints and projection slides.

Photographs of line drawings submitted in lieu of the original artwork *must* be on a high-contrast black-and-white paper or on Kodak "Kodalith ortho type 3" film or a similar material. Photographs on color paper or even on general-purpose black-and-white paper contain shades of gray which may prevent satisfactory copying by the printer.

The best prints are finished by ferrotyping; however, ordinary "glossy" prints are adequate if made on the proper paper from good negatives. Under no circumstances will photographs on matte-surfaced, sepia, portrait, or other special-purpose papers be accepted. *JH, R.E. Werner* (Eastman Kodak Company)

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