QUATERNARY GEOLOGY OF THE UPPER WEBER RIVER DRAINAGE BASIN, SUMMIT COUNTY, UTAH

by

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

Quaternary deposits in part of the western Uinta Mountains of northeastern Utah have been mapped at a scale of 1:50,000. The map area includes parts of eight 7.5-minute quadrangles. Numerous Quaternary landslides, and glacial deposits of one major Pleistocene glaciation (Pinedale) are widespread in the area. Significant potential geologic hazards from mass-wasting processes include large landslides, slumps, rockslides, debris slides, avalanches, and rock fall, as well as high runoff and sediment production caused by heavy land use on steep slopes.

Pinedale (late Wisconsin) glaciers formed an ice field in the highest parts of the map area that was drained by valley glaciers in the main Weber River valley, Smith and Morehouse Canyon, and South Fork. Numerous small cirque glaciers and small valley glaciers occupied steep canyons on north-facing slopes. Equilibrium-line altitudes (ELAs) of Pinedale glaciers were lower in the western part of the map area (about 2520 m), and were progressively higher with distance to the east (about 2800 m at east end of map area). This strong gradient (about 300 m/20 km) was probably caused by a west-to-east moisture gradient during Pinedale glaciation.

INTRODUCTION

This map and report document the Quaternary geology and extent of late Pleistocene glaciation in the upper part of the Weber River drainage basin in the Uinta Mountains, Utah (Figure 1). The map area includes parts of the following 7.5-minute quadrangles: Crandall Canyon, Hidden Lake, Slader Basin, Whitney Reservoir, Kamas, Hoyt Peak, Erickson Basin, and Mirror Lake. Surficial deposits were mapped in the field on aerial photographs at a scale of 1:24,000, and the data were transferred to 1:24,000-scale orthophotoquads, which were then reduced and compiled at a scale of 1:50,000. Field work was completed during the summer of 1991.

The map area has high topographic relief, with altitudes ranging from about 6500 ft (1980 m) along the Weber River east of Oakley, Utah, to over 11,500 ft (3500 m) along the drainage divide in the high mountains. The bedrock in the map area ranges in age from Precambrian to Tertiary and most of it has been mapped previously (Weeks, 1907; Forrester, 1937; Hooper, 1951; Root, 1952; Morris, 1953; Williams, 1953; Larsen, 1954; Meecham and Bitter, 1959; Hansen, 1969, 1986; Lindsay, 1969; Wallace, 1972; Crittenden and Peterman, 1975; Spreng, 1979; Bryant, 1985, 1990).

Although no climate data are available for stations in the map area Barnhardt (1973) has estimated climate parameters for three elevations on Bald Mountain, 3261, 3441, and 3627 m (10,700, 11,290, and 11,900 ft) using weather data from Brighton, Utah, and inferences about lapse rate and radiosonde data. At 3441 m the estimated mean annual air temperature is about -2° C (29° F), mean July temperature is about 9° C (48° F), and mean January temperature is about -11° C (13° F). At Trial Lake, a few kilometers south of the map area (Figure 1) at an elevation of about 3000 m (9800 ft), total mean-annual precipitation is about 96 cm (38 in.), with the greatest monthly average in January and the lowest in July. Precipitation values are lower, and temperature values are higher for localities at lower elevation in the map area.

Vegetation in the map area ranges from thick shrubs or grasslands on south-facing slopes at lower elevations, to coniferous forest dominated by spruce and fir, to alpine tundra on the summits of the highest peaks.

Subalpine meadows and bogs are common high in the glaciated valleys.

Quaternary deposits in the upper Weber River drainage basin can be classified as primarily glacial, mass-wasting, and alluvial or colluvial. All of these deposits are thin over most of the area, and only in a few places do they attain thicknesses greater than a few tens of meters. The most obvious landforms in the area have been produced by fluvial erosion, mountain glaciation, and mass wasting.

During the late Pleistocene the eastern part of the map area was covered by an ice field from which distributary glaciers flowed down the major valleys (Figures 2 and 3). Glaciers in the western part of the area were smaller and tended to be confined to the upper parts of individual valleys rather than emanating from the ice field (Figure 4). Glacial features were mapped throughout the Uinta Mountains by Atwood (1907; 1909), and smaller areas in the range have been mapped or studied more recently (Bradley, 1936; Barnhardt, 1973; Grogger, 1974, 1975, 1978; Currey, 1980; Carrara and others, 1985; Shroba, 1988). Mapping for this COGEOMAP project shows that many small valleys were glaciated than previously had been recognized, and refines our knowledge of the late Pleistocene glacier limits in this area.

The area has been, and continues to be, developed for private residences and commercial interests, and for public recreational and water resources. This map and report supply information on the distribution of landslides, which constitute a significant geologic hazard. Large landslides are present throughout the area, but the greatest number of them is along the Weber River Valley where development is proceeding rapidly. Because the Weber River drainage basin is an important watershed for thousands of people, including the local residents, land use should be managed carefully to protect water resources.

DESCRIPTION OF MAP UNITS

A system of descriptive and genetic map units, similar to those used on previous COGEOMAP projects, has been utilized in the mapping. The Quaternary deposits were deposited in alluvial (a), glacial (g), and mass-wasting (m) environments, as indicated by the first lower-case letter in the map-unit symbol. The second lower-case letter in the map-unit symbol indicates other distinguishing characteristics of the deposit, such as Qat for alluvium underlying terraces, or Qgt for glacial till. Two small areas of coarse-grained deposits for which the depositional environment is not known are mapped as "non-genetic" diamicton (Qnd).

PRE-QUATERNARY ROCKS

Precambrian, Paleozoic, Mesozoic, and Tertiary Rocks (b): The bedrock in the map area ranges in age from Precambrian to Tertiary. Most of the area mapped as bedrock (b) in Plate 1 is covered by thin colluvium, and no attempt has been made to distinguish colluvium as a map unit at the scale of 1:50,000. The higher parts of the map area are underlain by the Proterozoic Uinta Mountain Group, which consists of quartzite, sandstone, and shale. These rocks are overlain by Paleozoic and Mesozoic sedimentary rocks that within the map area dip to the north and northwest on the north flank of the Uinta arch. Tertiary sandstone, conglomerate, and shale, primarily of the Wasatch Formation, overlie the older rocks unconformably, mostly north of the Weber River.

The most important aspect of the bedrock geology from the point of view of Quaternary geology is its influence on the geomorphology of the area. For instance, the Weber River valley has formed along the contact between the gently dipping Tertiary rocks and the more steeply dipping Paleozoic and Mesozoic rocks (Bryant, 1990). This boundary also coincides with a major structural boundary -- the north flank fault zone and the approximately located Absaroka thrust fault. Major canyons cut by tributaries of the Weber River are eroded through the Precambrian and Phanerozoic rocks by fluvial, mass-wasting, and glacial processes.

The reader should refer to the following studies for further information on the bedrock geology (Weeks, 1907; Forrester, 1937; Hooper, 1951; Root, 1952; Morris, 1953; Williams, 1953; Larsen, 1954; Meecham and Bitter, 1959; Hansen, 1969, 1986; Lindsay, 1969; Wallace, 1972; Crittenden and Peterman, 1975; Spreng, 1979; Bryant, 1985, 1990).

QUATERNARY DEPOSITS

Glacial Deposits: Glacial deposits cover a large area in the Weber River drainage basin, although they are generally thin except where glacial till (Qgt) forms end or lateral moraines. The map unit Qbg is used over a large part of the area and represents thin and discontinuous glacial till overlying bedrock. The heavy line at the boundary of Qbg represents the approximate limit of glacial ice at its late Pleistocene maximum. This line is drawn partly on the basis of the distribution of till and erratics, but clues from erosional glacial landforms, such as striations on bedrock and the inferred upper limit of ice on cirque headwalls, were also used. Within the area mapped as Qbg glacial deposits are generally thin, but hummocky moraines are locally well developed, such as north of Bald Mountain and Reids Peak

(Barnhardt, 1973).

Thick glacial deposits in end moraines and lateral moraines are mapped in the lower parts of valleys. Here they consist of bouldery, hummocky to smooth-sided mounds of debris generally less than 50 ft (15 m) thick. Well-defined moraine ridge crests are indicated in Plate 1 by dashed lines. The best developed moraines are along the main Weber River valley, at the mouth of Smith and Morehouse Creek, and along Nobletts Creek and South Fork. Many smaller canyons are so steep that well-defined lateral and end moraines are not present, but masses of glacial till have been mapped in the lower parts of the valleys.

Although Atwood (1909) mapped moraines of two different ages in the Uinta Mountains, I found no evidence to distinguish moraines of more than one major glaciation. Atwood (1909) used position in the valley as a primary indicator of the relative age of moraines, but my reconnaissance showed no significant difference in morphology or weathering between Atwood's two groups. All the preserved moraines in the Weber River drainage basin are most likely related to the Pinedale glaciation of the Rocky Mountains, which is Wisconsin in age, or correlative with marine oxygen-isotope stages 3 and 2 (Pierce and others, 1976; Madsen and Currey, 1979). No numerical ages are available for the Uinta Mountain glacial deposits. If the correlation with the Pinedale is correct, however, the glaciers were probably at their maximum down-valley extent approximately 20 ka (Porter and others, 1983).

Mass-wasting Deposits: The mass-wasting deposits in the area have been mapped as landslides (Qms) and talus (Qmt). The landslides are primarily large complex landforms that probably included multiple episodes of movement and a number of different processes, including rotational slumping, earth flow, and debris slide. Although some landslides have been derived from the Proterozoic rocks, especially the Red Pine Shale, the sources for most landslides are the Paleozoic, Mesozoic, and Tertiary rocks. On Plate 1 landslide headscarps are marked by hachured lines.

In the higher parts of the area some landslides formed in post-glacial time in quartzite and sandstone of the Uinta Mountain Group. These consist of large rock slumps and rock slides on cirque headwalls that were probably oversteepened by glacial erosion and then left unsupported when the glaciers melted. Two examples of these are west of Island Lake at the head of Smith and Morehouse Creek, and at the head of Dry Fork.

One of the largest landslide complexes is in Broad Canyon, a tributary of Smith and Morehouse Canyon, where the Proterozoic Red Pine Shale and associated units has failed. At least the lower half of Broad Canyon has been glaciated by the main Smith and Morehouse glacier, adding to the complexity. Some of the movement of this landslide is probably post-glacial in age, but the landslide probably existed prior to the most recent episode of glaciation (Pinedale).

A rock slide in South Fork canyon just south of the confluence with White Pine Creek and Pullem Creek was mistaken by Atwood (1909) to be the terminal moraine of the South Fork glacier deposited during the later of the two epochs of glaciation he defined. However, the feature is clearly a rock slide derived from the Madison Limestone cliffs directly above it on the east side of the canyon. The rock slide contains almost no fine material and only a few shrubs and coniferous trees have begun to colonize its surface. Most of the limestone blocks ~~and~~ are disintegrating into angular blocks of cobble to pebble size. The South Fork rock slide is post-glacial in age as shown by its position across the glaciated valley, which it has dammed. South Fork itself may have formed a lake soon after the rock slide dammed the valley, but now drains through a karst swallow hole in the Madison Limestone on the upstream side of the rock slide (Figure 5). Water re-emerges from the karst system to the stream channel on the downstream side of the rock slide at the lower limit of Madison Limestone outcrops. Organic debris and clastic sediment accumulate on the upstream side of the rock slide when the swallow hole is episodically blocked during flood events. Layers of debris and sediment exposed by recent stream incision (Figure 6) indicate multiple episodes of downcutting and channel filling. There is potential for determining the approximate age of the rock slide from these deposits, which contain abundant material for radiocarbon dating.

Landslides constitute a significant geologic hazard along the main Weber canyon downstream from Smith and Morehouse Creek. In this area landslides line both sides of the valley, including many areas where residences and subdivisions have been built. Most landslides on the north side of the valley are derived from the Wasatch Formation, whereas those on the south are derived from Mesozoic rocks, primarily the Ankareh Formation. Although there is no evidence that any of these landslides has moved significantly in recent time, landslide stability is difficult to predict, and any of these landslides should be regarded as potentially hazardous in the future.

A landslide in sec. 6, T. 1 S., R. 7 E., on the south side of the Weber River valley across from the mouth of Perdue Creek at and near Hidden Lake, is composed of rock debris derived from the Ankareh Formation and Nugget Sandstone. It is perched 120-440 ft (35-130 m) above the valley bottom, is being undercut by the Weber River, and appears to be in a vulnerable position.

A number of houses in a subdivision are built directly on the landslide, and other houses and a highway are built in the narrow river valley below the landslide. It is difficult to know when or if this landslide will move again in the foreseeable future, but if people are aware of the potential hazard they can be prepared with emergency procedures. Other landslides in developed areas pose similar potential hazards.

Another mass wasting problem that is less of a hazard, but still a concern for water and soil quality, is a large area of steep slopes on the Wasatch Formation on the north side of the Weber River east of Smith and Morehouse Creek. In this area numerous small debris slides have formed in the headward parts of the drainage networks and are feeding abundant sediment into the steep stream channels that are tributary to the Weber River. Steep channel bottoms in this area are choked with the sediment. The locations of some of the larger debris slides are indicated on Plate 1 by short curved hachured lines, which mark the headscarps. It is clear that the ratio of surface runoff to infiltration is higher than it should be in this area, probably as a result of overgrazing by sheep. This is a major cause for concern because it indicates that part of the Weber River watershed is in an unhealthy state -- less water recharges ground water systems in this area and surface water flowing from this area is probably of lower quality than in surrounding areas.

Talus deposits (Qmt) are most abundant and thickest in the high glaciated part of the area. Talus deposits are abundant along cirque headwalls and below cliffs around peaks that stood as nunataks during periods of maximum glaciation. The talus consists of loose angular boulders of locally derived rock, in most cases quartzite or sandstone, on steep slopes with little vegetation. On Plate 1 dashed lines within areas mapped as talus indicate the ridge crests of protalus ramparts, which formed at some time when late-season snow or ice on the slopes allowed blocks to slide or roll farther from the cliff and accumulate in a ridge or rampart (Figure 6). Lichen cover on many of the boulders in protalus ramparts is extensive (≥ 50 %), suggesting that the protalus ramparts may have formed prior to the late Holocene Neoglaciation, probably in late Pleistocene time.

Talus cones continue to be active today, but no protalus ramparts are presently forming in the area. Boulders accumulate in talus cones through processes of snow avalanching and rock fall directly from cliffs. Avalanche boulder tongues, which are accumulations of rock debris deposited dominantly by avalanching (Barnhardt, 1973), are mapped with the talus (Qmt) on Plate 1.

Alluvial Deposits: Alluvium in the map area is divided into four map units, Qal, Qac, Qat, and Qab. The distinction between Qal and Qac is arbitrary in some places, but in general the alluvium mapped as Qal is found along major perennial streams and underlies flood plains and the lowest young terraces directly above flood plains. Qac consists of the deposits of smaller-order streams, many of which are ephemeral and tributary to the larger perennial streams. The designation Qac refers to the presence of a large component of slope-wash colluvium in these deposits. Alluvium mapped as Qal and Qac is post-glacial (Holocene) in age.

Along the Weber River valley, terraces underlain by alluvial gravel and graded downstream from the late Pleistocene end moraines of the Weber River glacier and the Smith and Morehouse glacier are mapped as Qat. The terrace surfaces are 30 to 50 ft (10 to 15 m) above the modern Weber River, and represent outwash deposition during glaciation.

In higher parts of glaciated valleys near drainage divides, the glacially scoured rock surfaces are uneven and are sites where lakes and bogs have formed (Figure 7). The bog deposits are mapped as Qab because many of them contain an alluvial detrital component; however, much of the total volume of the bog deposits consists of sedge and moss peat. Barnhardt (1973) and Currey (1980) reported radiocarbon ages of 8235 ± 345 yr B.P., and 2970 ± 150 yr B.P., for peat and wood from bogs near Bald Mountain. The 8200 yr age was for peat near the base of a bog, and indicates that deglaciation was complete on the north slope of Bald Mountain by at least this time in the early Holocene. By comparison with other mountain ranges in the Rocky Mountain region, the Uinta Mountains were probably deglaciated long before 8000 yr B.P. (Porter and others, 1983).

Non-genetic diamicton (Qnd): Two small areas of coarse gravel of uncertain origin were mapped in the central part of the area. Both deposits consist of boulders of Uinta Mountain quartzite and fine-grained material in low vegetated mounds on remnants of a high-level erosion surface -- Mud Lake Flat and Slader Ridge. The erosion surface is probably part of the Gilbert Peak surface of probable Oligocene age (Bradley, 1936; Hansen, 1986). The boulder deposits themselves, however, may be much younger than the erosion surface, and are mapped as Quaternary deposits (Qnd), even though no definite evidence of their ages has been found.

The designation "nd" is used for the boulder deposits to indicate their undetermined origin. They could have been deposited by glaciers or by streams, but in the absence of exposures of fresh material and surficial geomorphic expression, the depositional environment cannot be determined.

QUATERNARY GLACIATION

During the Pleistocene, glaciers repeatedly accumulated in the Uinta Mountains and carved the majestic alpine landscape. Although there is a preserved stratigraphic record of only a few major advances of ice in the Uinta Mountains (Atwood, 1909; Bradley, 1936), the global record of glaciation shows many cycles of ice accumulation and melting (Imbrie and others, 1984), each with a magnitude similar to the last major glaciation. Therefore, the large-scale landforms, such as cirques, glacial troughs, and horns, are probably the result of many episodes of glacial erosion in the mountains. The preserved record of the last major glaciation in the Weber River drainage basin can be thought of as a general template for the character of previous, poorly documented glaciations.

The maximum extent of glaciers in the Weber River drainage basin during the last major Pleistocene glaciation (Pinedale) is shown in Plate 1. This map can be used to describe the glaciers and to compare the climate of Pinedale time to that of today. At maximum extent, large valley glaciers flowed down the major valleys (Weber, Smith and Morehouse, South Fork) from an extensive ice field, only part of which was in the Weber River basin. Other valley glaciers outside the map area drained the ice field down the Bear River, Provo River, and smaller valleys on the south flank of the range (Atwood, 1909).

That the ice mass was an ice field and not an ice cap is shown by the distribution of erosional features that indicate ice-flow directions in the vicinity of modern drainage divides along the crest of the range. In the saddles or cols between horns or nunataks, striations and other erosional features are common. Crescentic marks, lunate gouges, and possible cavitation pits (Figures 8 and 9) on quartzite bedrock indicate ice flow in both directions away from the cols, and therefore demonstrate that the ice divides were essentially the same as the modern drainage divides. Flow in an ice field is strongly influenced by the underlying topography, in contrast to flow in an ice cap, which is determined by the morphology of the ice dome. If the directional features in the Uinta Mountains showed there had been significant flow through the divides the ice mass would be more likely classified as an ice cap (Sugden and John, 1976).

In addition to the large ice-field--valley-glacier system, a number of smaller valley glaciers or enlarged cirque glaciers formed in the steep north-facing valleys tributary to the Weber River (Plate 1). These glaciers were fed in most cases by single cirques and ended at intermediate elevations in the valleys or terminated at the lateral moraines of larger valley glaciers (e.g., the small glaciers on the north slope of Slader Ridge). Atwood (1909) recognized that Swifts Canyon and Bear Basin had been glaciated, but did not recognize the glacial features in the numerous other small canyons in the area.

The map of the extent of glacial ice (Plate 1) can be used to estimate the equilibrium-line altitude (ELA) of the Pinedale glaciers. The equilibrium line on a glacier is the boundary between the glacier's accumulation zone, where more snow falls during the winter than melts the following summer, and the ablation zone, where during a melting season the previous winter's snowfall, in addition to part of the glacier that had formed many years earlier and had flowed downvalley to the ablation zone, melts. In the accumulation zone there is a net increase in mass each year, and there is a net decrease in mass in the ablation zone. The glacier is in mass balance when the inputs equal the outputs. When inputs exceed outputs for a sufficiently long period, the glacier advances downvalley, and the altitude of the equilibrium line is lowered.

On a modern glacier the equilibrium line can be identified readily by field observations -- it often appears at the end of the ablation season as the distinct lower boundary of the fresh snow and firn that fell the previous winter in contrast to the debris-laden ice exposed at the glacier surface in the ablation zone. For paleoglaciers, however, other means must be employed to determine the ELA. A number of methods, some of which are more subjective than others, have been employed by researchers in mountain ranges throughout the world. Meierding (1982) compared the results from five different techniques applied to a large number of glaciated valleys in the Colorado Front Range, and found that two methods, accumulation-area ratio (AAR) and toe-to-headwall altitude ratio (THAR), gave the most consistently accurate results. These two methods also yielded approximately the same estimates of ELA when applied to the same reconstructed glaciers.

For this COGEOMAP project the toe-to-headwall altitude ratio (THAR) was used as the method for estimating paleo-ELAs of the Pinedale glaciers in the Weber River drainage basin. The THAR method was used instead of the AAR method because it was easier and faster to employ. In the THAR method the lower and upper altitudinal limit of ice in each valley are estimated from the map and the ELA is calculated using the following formula:

$$ELA = Z_t + [0.4(Z_h - Z_t)]$$

where Z_t is the lower limit of ice at the end moraine, and Z_h is the upper limit of ice at the cirque headwall. All values are in meters above sea level. A THAR of 0.4 is used for the glaciers in the Weber River basin because Meierding (1982) found that this ratio gave results that were most consistent with other methods, especially the AAR method.

The results from the THAR analysis are listed in Table 1 and shown in Figure 10. A spatial pattern that appears in the data is that the ELA for glaciers in the Weber River drainage basin increased with distance to the east from the western end of the mountain range (Figure 10). The regional ELA gradient amounts to almost 300 m in a horizontal distance of about 20 km, and is probably the result of the primary moisture source for the glaciers being from the west.

Mulvey (1985) used a single ELA value of 2819 m for the western Uinta Mountains in his regional study of Pinedale ELAs in the eastern Great Basin. He used the THAR method with a THAR of 0.5. Mulvey did not report which glaciers he used in determining an ELA of 2819 m, but by comparison, the data on Z_t and Z_h for Smith and Morehouse glacier in Table 1 would yield an ELA of 2825 m if a THAR of 0.5 were used. Porter and others (1983, Fig. 4-2) show the 3000 m contour crossing the central part of the Uinta Mountains on a map of cirque floor altitudes for late Wisconsin glaciers in the western United States. A cirque-floor altitude of 3000 m is reasonable for the eastern part of the Weber River basin. Zielinski and McCoy (1987) compared the data of Porter and others (1983, Fig. 4-2) to modern snowpack data in altitudinal gradients, and suggested that the changes in temperature or snow accumulation patterns from the late Pleistocene to the present were nonuniform across the Great Basin.

No attempt has been made to estimate the amount of ELA depression in the Weber River basin during the Pinedale as compared to modern ELAs in the area. Currey and James (1982) suggested that late Pleistocene ELAs in north-central

Utah may have been on the order of 800 m lower than modern ELAs. Madsen and Currey (1979) estimate that late Pleistocene mean annual temperature in the Little Cottonwood Canyon area of the Wasatch Mountains was at least 5° C colder than today.

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REFERENCES

Atwood, W. W., 1907, Glaciation of the Uinta Mountains: Journal of Geology, v. 15, p. 790-804.

Atwood, W. W., 1909, Glaciation of the Uinta and Wasatch Mountains: U.S. Geological Survey Professional Paper 61, 96 p.

Barnhardt, M. L., 1973, Late Quaternary geomorphology of the Bald Mountain area, Uinta Mountains, Utah, [M.S. thesis]: University of Utah, Salt Lake City.

Bradley, W. H., 1936, Geomorphology of the north flank of the Uinta Mountains: U.S. Geological Survey Professional Paper 185-I.

Bryant, B., 1985, Structural ancestry of the Uinta Mountains, *in* Picard, M. D., ed. Geology and energy resources, Uinta Basin of Utah: Utah Geological Association Publication 12, p. 115-120.

Bryant, B., 1990, Geologic map of the Salt Lake City 30' X 60' quadrangle, north-central Utah, and Uinta County, Wyoming: U.S. Geological Survey Miscellaneous Investigations Series Map I-1944, 1:100,000.

Carrara, P. E., Short, S. K., and Shroba, R. R., 1985, A pollen study of

Holocene peat and lake sediments, Leidy Peak area, Uinta Mountains, Utah: Brigham Young University Geology Studies, v. 32, pt. 1, p. 1-7.

Crittenden, M. D., Jr., and Peterman, Zell, 1975, Provisional Rb-Sr age of the Precambrian Uinta Mountain Group, northeastern Utah: Utah Geology, v. 2, no. 1, p. 75-77.

Currey, D. R., 1980, Radiocarbon dates and their stratigraphic implications from selected localities in Utah and Wyoming: Encyclia, v. 57, p. 110-115.

Currey, D. R., and James, S. R., 1982, Paleoenvironments of the northeastern Great Basin and northeastern basin rim region: A review of geological and biological evidence, in Madsen, D. B., and O'Connell, J. F., eds, Man and environment in the Great Basin: Society of American Archaeology Papers No. 2, p. 27-52.

Drewry, D., 1986, Glacial geologic processes: London, Edward Arnold, 276 p.

Forrester, J. D., 1937, Structure of the Uinta Mountains: Geological Society of America Bulletin, v. 48, p. 631-666.

Grogger, P. K., 1974, Glaciation of the High Uintas Primitive Area, with emphasis on the northern slope, [Ph.D. dissertation]: University of Utah, Salt Lake City.

Grogger, P. K., 1975, Neoglaciation of the northern slope of the High Uintas Primitive area, Utah: Great Plains-Rocky Mountain Geographical Journal, v. 4, p. 37-43.

Grogger, P. K., 1978, A late Wisconsin fourth stade in the Uinta Mountains, Utah: American Quaternary Association Abstracts, v. 5, p. 206.

Hansen, W. R., 1969, The geologic story of the Uinta Mountains: U.S. Geological Survey Bulletin 1291, 144 p.

Hansen, W. R., 1986, Neogene tectonics and geomorphology of the eastern Uinta Mountains in Utah, Colorado, and Wyoming: U.S. Geological Survey Professional Paper 1356, 78 p.

Hooper, W. G., 1951, Geology of the Smith and Morehouse-South Fork area (Summit County) Utah [M.S. thesis]: University of Utah Salt Lake City, 55 p.

Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix., A. C., Morley, J. J., Pisias, N. G., Prell, W. L., and Shackleton, N. J., 1984, The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $\delta^{18}O$ record, in Berger, A. L. and others, eds., Milankovitch and climate, Part 1, D. Reidel Publishing Company, p. 269-305.

Larsen, W. N., 1954, Precambrian geology of the western Uinta Mountains (Summit, Wasatch, and Duchesne Counties) Utah [M.S. thesis]: University of Utah, Salt Lake City.

Lindsay, J. B., ed. 1969, Geologic guidebook of the Uinta Mountains, Utah's maverick range: Intermountain Association of Petroleum Geologists Guidebook, 16th annual field conference.

Madsen, D. B., and Currey, D. R., 1979, Late Quaternary glacial and vegetation changes, Little Cottonwood Canyon area, Wasatch Mountains, Utah: Quaternary Research, v. 12, p. 254-270.

Meecham, D. F., and Bitter, R. K., compilers, 1959, Geology of the Wasatch and Uinta Mountains transition area, northeastern Utah and southwestern Wyoming, *In* Guidebook to the geology of the Wasatch and Uinta Mountains transition area: Intermountain Association of Petroleum Geologists Guidebook, 10th annual field conference, map scale 1:190,000.

Meierding, T. C., 1982, Late Pleistocene glacial equilibrium-line altitudes in the Colorado Front Range: A comparison of methods: Quaternary Research, v. 18, p. 289-310.

Morris, E. C., 1953, Geology of the Big Piney area, Summit County, Utah [M.S. thesis]: University of Utah, Salt Lake City.

Mulvey, W. E., 1985, Reconstruction and interpretation of late Pliestocene equilibrium-line altitudes in the Lake Bonneville region [M.S. thesis]: Slat Lake City, University of Utah, 65 p.

Pierce, K. L., Obradovich, J. D., and Friedman, I., 1976, Obsidian hydration dating and correlation of Bull Lake and Pinedale Glaciations near West Yellowstone, Montana: Geological Society of America Bulletin, v. 87, p. 703-710.

Porter, S. C., Pierce, K. L., and Hamilton, T. D., 1983, Late Wisconsin mountain glaciation in the western United States, <u>in</u> Porter, S. C., ed., Late-Quaternary environments of the United States: Volume 1, The late Pleistocene: Minneapolis, University of Minnesota Press, p. 71-111.

Root, R. C., 1952, Geology of the Smith and Morehouse-Hayden Fork area, Summit County, Utah [M.S. thesis]: University of Utah, Salt Lake City, 58 p.

Shroba, R. R., 1988, Clay accumulation and soil B horizon development in late Pleistocene tills, eastern Uinta Mountains, Utah: American Quaternary Association Abstracts, v. 10, p. 152.

Spreng, W. C., 1979, Upper Devonian and Lower Mississippian strata on the flanks of the western Uinta Mountains, Utah: Brigham Young University Geology Studies, v. 26, pt. 2, p. 67-78.

Sugden, D. E., and John, B. S., 1976, Glaciers and landscape: New York, John Wiley and Sons, 376 p.

Weeks, F. B., 1907, Structure and Stratigraphy of the Uinta Range: Geological Society of America Bulletin, v. 18, p. 427-448.

Wallace, C. A., 1972, A basin analysis of the upper Precambrian Uinta Mountain Group: Santa Barbara, University of California Ph.D. thesis, 412 p.

Williams, N. C., 1953, Late pre-Cambrian and early Paleozoic geology of western Uinta Mountains, Utah: American Association of Petroleum Geologists Bulletin, v. 37, p. 2734-2742.

Zielinski, G. A., and McCoy, W. D., 1987, Paleoclimatic implications of the relationship between modern snowpack and late Pleistocene equilibrium-line altitudes in the mountains of the Great Basin, western U.S.A.: Arctic and Alpine Research, v. 19, p. 127-134.

Figure Captions

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Figure 5. Photograph of the swallow hole (arrow) in the Madison Limestone on the upstream side of the rock slide in South Fork. The vertical exposure of fluvial gravel and organic debris in the foreground is about 2 m (6 ft) high.

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head of the Middle Fork of the Weber River.

Figure 7. Bog near the drainage divide north of Kamas Lake (north of Bald Mountain). The bog has almost completely filled the depression with moss and sedge peat.

Figure 8. Striations, crescentic fractures, and lunate gouges on quartzite. The ice flowed away from the viewer. Note that the lunate gouges are convex in an up-ice direction, and the crescentic fractures are convex in a down-ice direction. The scale is 15 cm (6 in.) long.

Figures 9A and 9B. Striated and polished quartzite surfaces showing possible cavitation pits on the down-ice sides (right) of low rises in the quartzite. The scale is 15 cm (6 in.) long in each photograph. Cavitation pits form where melt water flowing at high velocity in a confined space at the base of the glacier suddenly enters an enlarged cavity at the base of the ice (Drewry, 1986) (such as the areas on the right sides of the photographs). At high velocity, pressure is decreased, and tiny airless bubbles form in the water. Within the cavity, the velocity decreases and the pressure increases causing the bubbles to collapse. When bubbles collapse next to the rock they exert a very high but localized stress on the rock, and over a long period of flow the rock is pitted. Figure 9A shows a large area of small pits, and Figure 9B shows larger pits that have formed a pattern that looks like ripple marks.

Figure 10. Diagram showing a plot of the equilibrium-line altitudes (ELA) of Pinedale glaciers in the Weber River drainage basin versus distance east of an arbitrary north-south line west of Pinyon Canyon (data from Table 1). Refer to Plate 1 for locations of glaciers. P = Pinyon Canyon; S =

Swifts Canyon; BB = Bear Basin; WP = White Pine; SF = South Fork; N = Nobletts Creek; SM = Shingle Mill; SM-e = Shingle Mill east; BT-n = Bear Trap north; S&M = Smith and Morehouse; SR-w = Slader Ridge west; BC = Broad Canyon; SC = Slader Creek; W = main Weber Valley glacier.

Table 1. ELA estimates for Pinedale glaciers in the Weber River drainage basin

glacier[1]	Zt[2] (m)	Zh[3] (m)	ELA[4] (m)	dist.[5] (km)
Pinyon Canyon	2440	2860	2610	0.9
Swifts Canyon	2200	3000	2520	1.8
Bear Basin	2300	2900	2540	2.9
White Pine Creek	2250	3050	2570	4.4
South Fork	2250	3150	2610	6.5
Nobletts Creek	2500	2950	2680	7.7
Shingle Mill Creek	2480	2950	2670	8.7
Shingle Mill east	2500	2990	2700	9.6
Bear Trap north	2650	2950	2770	9.9
Smith & Morehouse	2250	3400	2710	12.7
Slader Ridge west	2500	3000	2700	13.3
Broad Canyon	2600	3100	2800	14.8
Slader Creek	2550	3100	2770	15.4
main Weber	2300	3500	2780	21.5

[1]see Plate 1 for locations of valleys

[2]altitude of lower limit of glacier

[3]altitude of upper limit of glacier

[4]calculated by the THAR method, see text

[5]distance east of an arbitrary north-south line west of Pinyon Canyon

Legend for Plate I
("Q" omitted on map)

MAP UNITS

QUATERNARY DEPOSITS

Qal Alluvium (Holocene) -- Coarse- to fine-grained alluvium along flood plains and channels of major streams. Maximum thickness probably less than 20 m.

Qac Alluvium and colluvium (Holocene) -- Coarse- to fine-grained alluvium and colluvium of lower-order streams and hillslopes. Maximum thickness probably less than 20 m.

Qab Bog deposits (Holocene) -- Fine-grained alluvium and peat in bogs filling shallow depressions at higher elevations in glaciated valleys. Maximum thickness probably less than 20 m.

Qat Terrace alluvium (late Pleistocene) -- Coarse-grained outwash alluvium forming terraces along the Weber River valley downstream from terminal moraines. Maximum thickness probably less than 50 m.

Qgt Glacial till (late Pleistocene) -- Poorly sorted bouldery till that forms lateral, terminal, and recessional moraines; most likely of Pinedale (late Wisconsin) age. Maximum thickness probably less than 75 m.

Qbg Thin, discontinuous glacial till overlying bedrock (late Pleistocene) -- Poorly sorted bouldery till; includes areas of erratics and striated bedrock. Maximum thickness unknown.

Qms Landslide deposits (Holocene and Pleistocene) -- Deposits of large landslides, slumps, debris slides, and rock slides; many are post-glacial, but some large slides existed prior to glaciation. Maximum thickness unknown.

Qmt Talus deposits (Holocene) -- Deposits of angular boulders on steep slopes below cliffs, especially in glaciated valleys; includes thickened talus in protalus ramparts and avalanche boulder tongues. Maximum thickness unknown.

Qnd Diamicton (Pleistocene ?) -- Deposits of rounded boulders in a fine-grained matrix; depositional environment unknown (probably glacial or fluvial). Maximum thickness probably less than 30 m.

PRE-QUATERNARY ROCKS

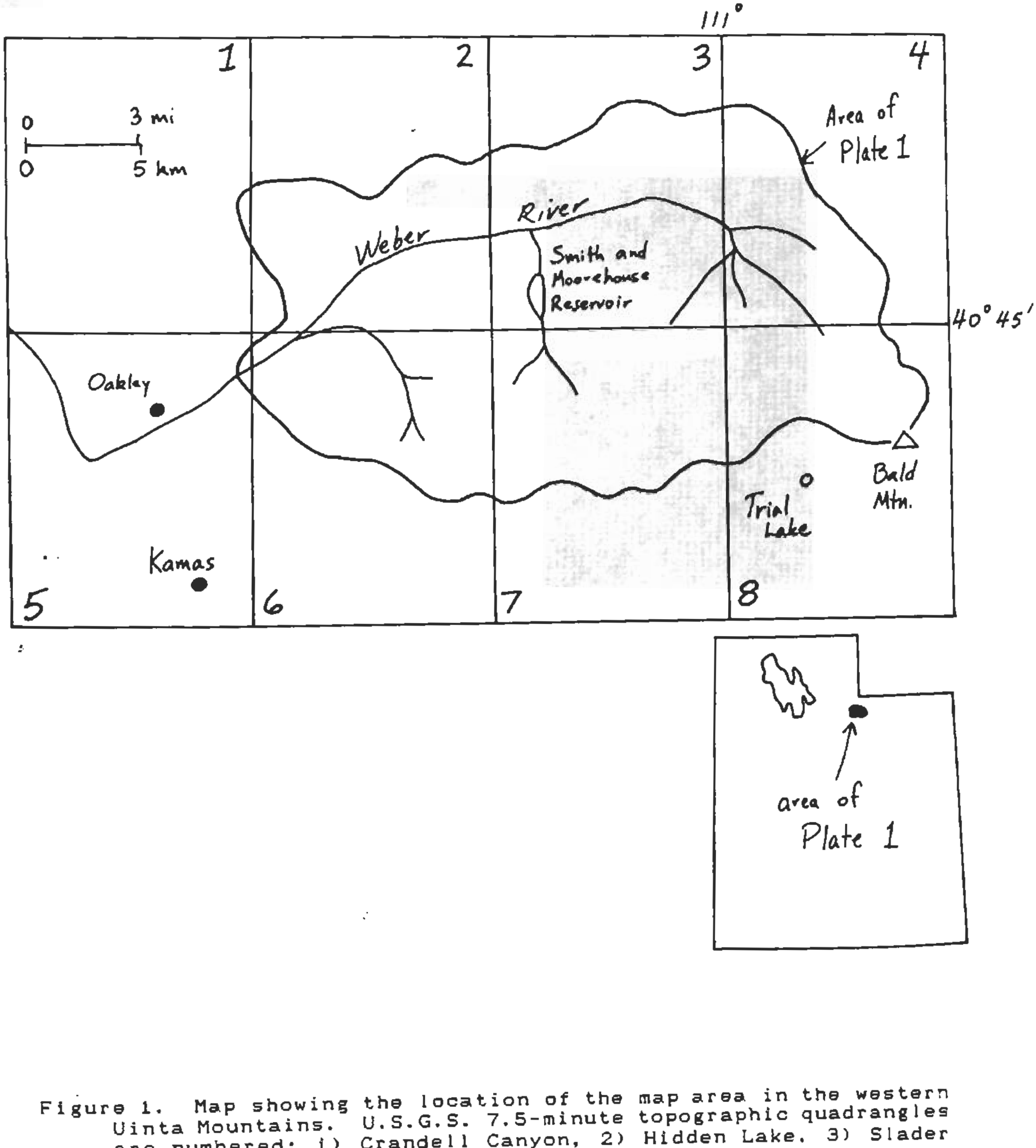
b Precambrian, Paleozoic, Mesozoic, and Tertiary Rocks -- Bedrock (see Bryant, 1990).

SYMBOLS

Landslide head scarp

Direction of glacial ice flow as indicated by crescentic fractures or lunate gouges in bedrock.

Ridge crest of moraine (in glacial till) or protalus rampart (in talus)



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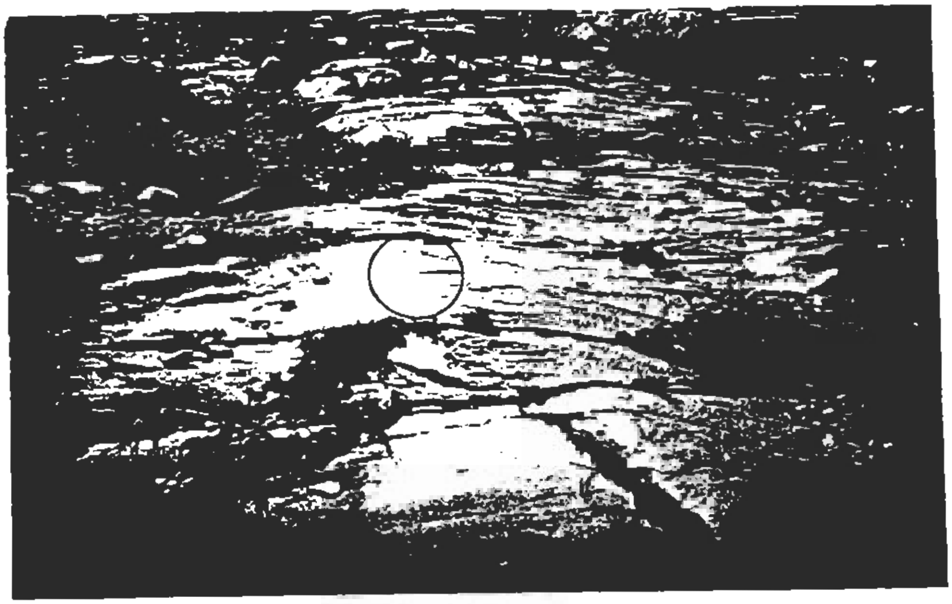
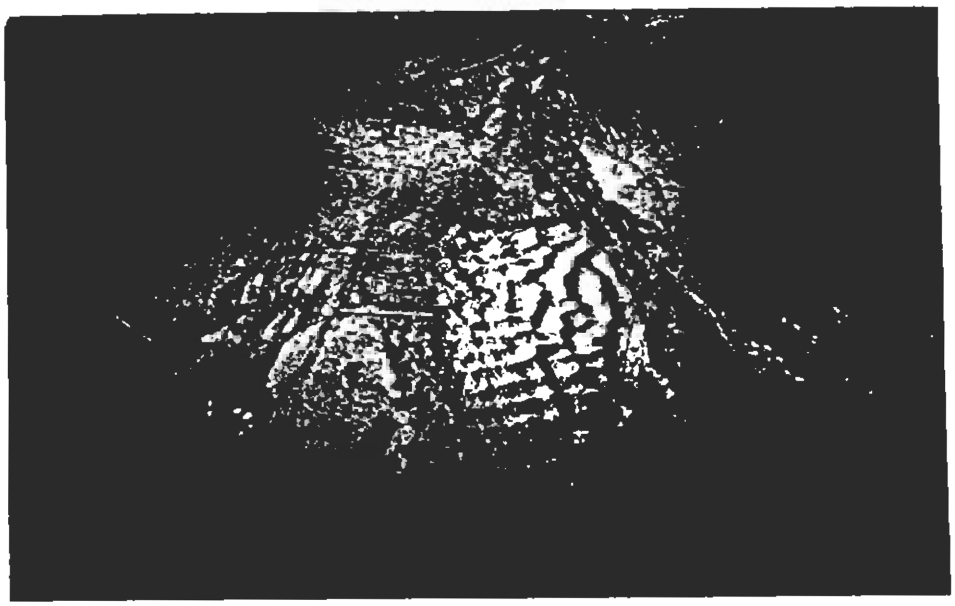


A.



B.

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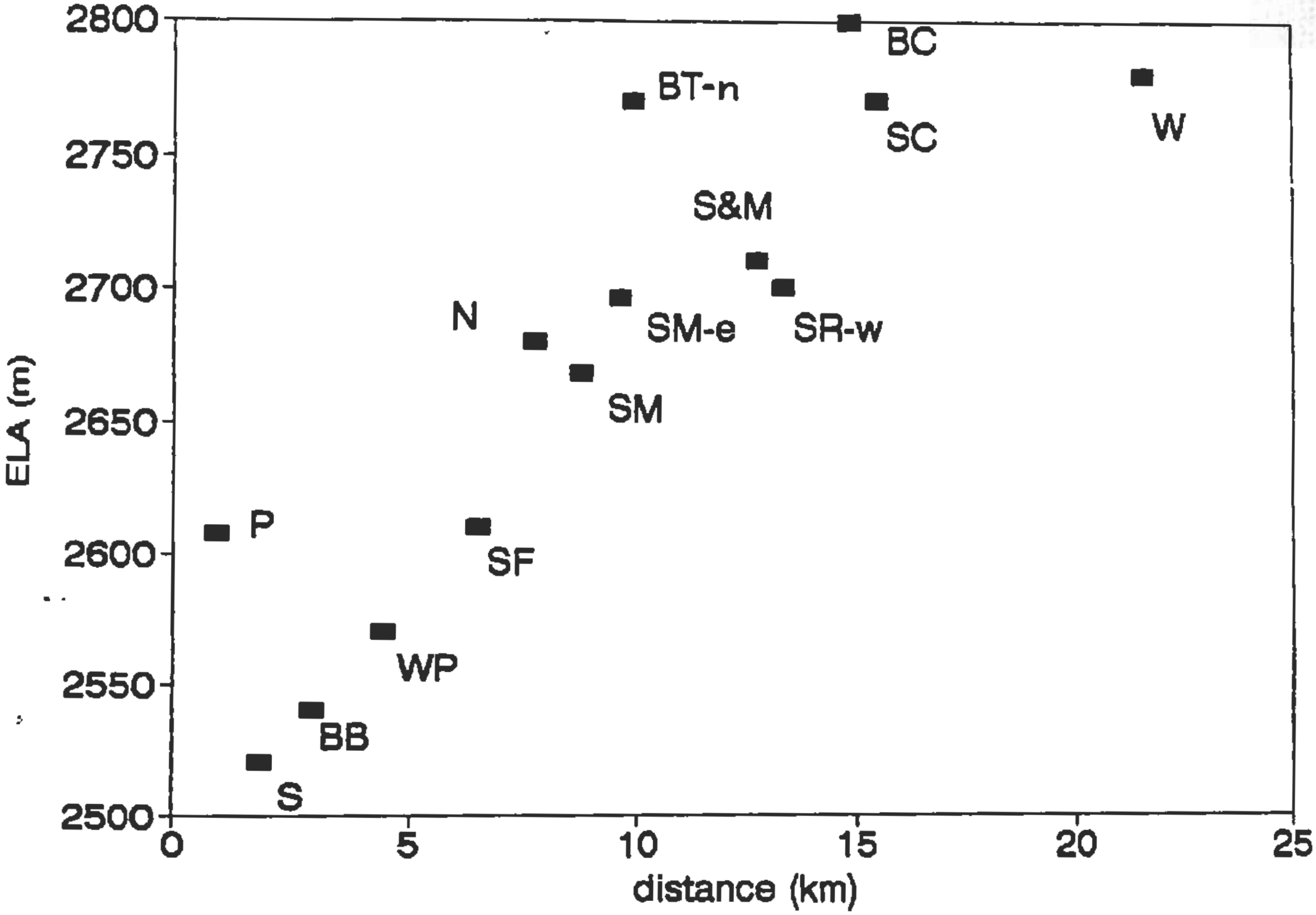

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Correlation of Map Units