CONTRIBUTIONS TO QUATERNARY GEOLOGY OF THE COLORADO PLATEAU

By G.E. Christenson, C.G. Oviatt, J.F. Shroder, and R.E. Sewell

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PREFACE

By G.E. Christenson and C.G. Oviatt

The Colorado Plateau is a unique area for the study of Quaternary deposits and processes since it has been the site of continuous stream downcutting, cliff retreat, and landscape denudation during much of this period. As a result, the Quaternary record is incomplete and the history must be pieced together from detailed study of the isolated outcrops that remain. The record consists chiefly of glacial, periglacial, and mass-wasting deposits in high plateau areas and in mountain ranges that rise above the plateau surface, alluvial-fan and pediment gravels surrounding areas of high elevation, loess and eolian sand on plateau surfaces, and alluvium along stream courses and in canyons. Reconnaissance mapping of these deposits and detailed studies of certain specific localities have been completed in the Colorado Plateau, but many areas remain unstudied or await re-evaluation based on current concepts and utilizing new analytical techniques. Papers presented in this volume include both detailed and reconnaissance studies representing new information in areas not previously studied and reinterpretations in areas previously studied in detail. Quaternary deposits of all types are addressed to some extent in these papers, with emphasis on alluvial and eolian deposits along the northern tributaries to the San Juan River and mass movement/glacial deposits in the La Sal Mountains.

The papers by Christenson and Oviatt discuss problems in alluvial stratigraphy, particularly for the late Holocene. Although there is some suggestion that alluvial sequences in this area can be correlated on a regional scale, implying climatic controls on fluvial deposition and erosion, local geomorphic factors cause considerable variability from one drainage basin to another. Detailed studies of alluvial stratigraphy in individual drainage basins such as those presented here are increasing our understanding of ephemeral-stream systems and aiding our ability to distinguish between the effects of local geomorphic factors and the effects of climate changes. The paper by Shroder and Sewell presents new interpretations of diamictons in the La Sal Mountains that raise questions about the limits of glaciation in that mountain range. Their work demonstrates the difficulties of distinguishing between glacial and mass-wasting deposits, and many of their methods are likely to be useful in other high-mountain areas of the Colorado Plateau where the effects of glacial and mass-wasting processes coincide.

The Quaternary deposits of the Colorado Plateau are also of practical importance in assessing geologic hazards (flooding, slope instability, avalanche potential) affecting construction and in making long-term assessments of landscape stability and tectonic setting for critical facilities such as the proposed Paradox Basin high-level nuclear-waste repository. Additional detailed studies of Quaternary deposits on the Colorado Plateau have been recently completed for the U.S. Department of Energy as part of the geotechnical investigation for siting of the nuclear waste repository. These studies, along with those presented in this volume, have contributed greatly to our knowledge of the Quaternary history of the Colorado Plateau in Utah. However, much detailed work remains to be done and many intriguing and critical questions remain unanswered.

Utah Geological and Mineral Survey Special Studies 64, 1985

QUATERNARY GEOLOGY OF THE MONTEZUMA CREEK-LOWER RECAPTURE CREEK AREA, SAN JUAN COUNTY UTAH

By Gary E. Christenson

QUATERNARY GEOLOGY OF THE MONTEZUMA CREEK-LOWER RECAPTURE CREEK AREA, SAN JUAN COUNTY, UTAH

By GARY E. CHRISTENSON [1]

ABSTRACT

Montezuma and Recapture Creeks drain 1410 mi^2 (3650 km^2) of southeastern Utah from the Abajo Mountains and Great Sage Plain south to the San Juan River. Jurassic and Cretaceous sedimentary rocks underlie the Colorado Plateau here and are exposed in canyons of both creeks. Quaternary deposits include gravels surrounding the Abajo Mountains, loess and eolian sand on plateau surfaces, and alluvium in stream valleys.

Incision of Montezuma Creek into the plateau surface began following deposition of the alluvial-fan and pediment gravels now found on upland surfaces around the Abajo Mountains before 700,000 B.P. Gravel-covered rock-cut terraces were formed at various levels as the canyon was cut, and calcic soil development in lower terraces indicates downcutting to near the present stream level by Late Pleistocene time. Early Holocene alluvial deposits are lacking, but alluvial deposition began in upper Montezuma Creek before 5130 B.P., filling the canyon locally with more than 50 feet (15 m) of sand, silt, and clay. This fill was subsequently incised 20-30 feet (6-9 m) and alluviation of this new channel began around 1430 B.P. Filling to the previous level with subsequent channel stability lasted until about 1900 A.D. when Montezuma Creek re-incised to its present level 20-40 feet (6-12 m) below the fill surface.

In lower Recapture Canyon, soil development in terrace gravels indicates downcutting to a level 115 feet (35 m) above stream grade by 150,000 B.P. A fill terrace of probable Holocene age is presently 20-25 feet (6-8 m) above channel level.

Rock-cut terraces in both canyons indicate net downcutting during the Pleistocene. Terrace development was locally influenced by lithology and few prominent or paired terrace levels are found. The Holocene fill sequence in Montezuma Creek indicates aggradation which was in part synchronous with alluviation elsewhere on the Colorado Plateau, although time control is not sufficient to attempt detailed correlation.

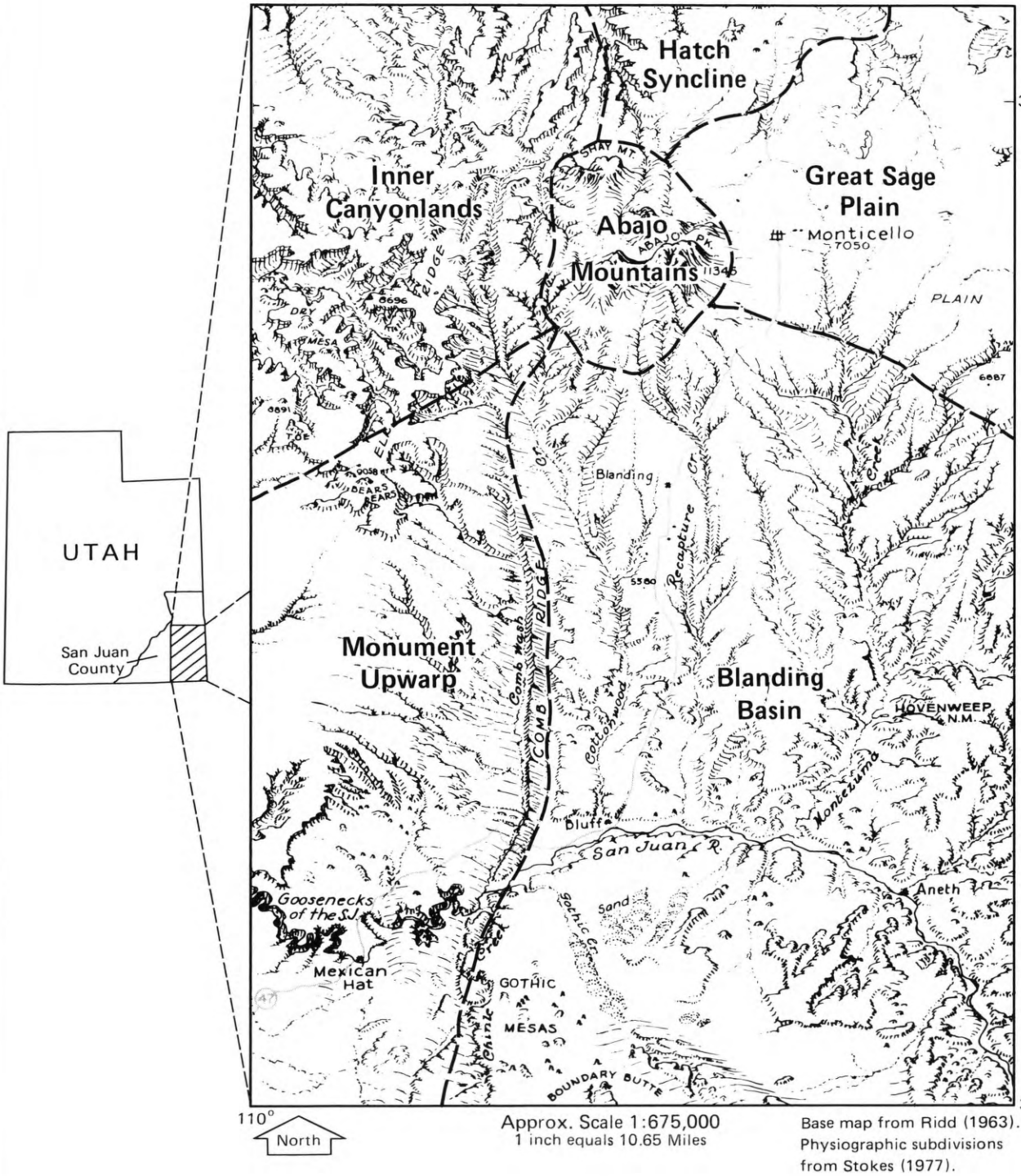
INTRODUCTION

PURPOSE AND SCOPE

This project was undertaken as part of the Utah Geological and Mineral Survey review of the Department of Energy (DOE) geotechnical studies in the Paradox Basin of Utah for siting of a nuclear waste repository. The principal objective of the study is to gain a further understanding of Quaternary processes in the area through a study of Quaternary deposits along Montezuma and lower Recapture Creeks, San Juan County, Utah. Both creeks are tributaries of the San Juan River and flow southward from headwaters in the Abajo Mountains (fig. 1). From the reconstructed Quaternary history, an assessment can be made of past erosional and depositional processes, including rates of activity, and of possible factors controlling these processes and rates. It also provides a basis for predicting future effects of surficial geologic processes in this part of the Paradox Basin.

Quaternary deposits were mapped from 1:24,000 scale air photos with field inspection of alluvial and soil stratigraphy in existing exposures. The most accurate existing topographic control consists of USGS 15' topographic quadrangle maps with 40-foot (12.2 m) contour intervals. Heights of upper terraces were estimated from these maps while those of lower terraces were measured with a hand level. Samples from various Holocene alluvial deposits were taken

[1]Site Investigations Section, Utah Geological and Mineral Survey.



Approx. Scale 1:675,000
1 inch equals 10.65 Miles

Base map from Ridd (1963). Physiographic subdivisions from Stokes (1977).

FIGURE 1. Location and physiography of the Montezuma Creek-Recapture Creek area.

from stream cuts for radiocarbon age determinations. Approximate ages of rock-cut terraces were estimated by comparison of soil profile development in the associated gravel deposits with dated profiles in similar deposits in nearby areas.

Work on Holocene deposits was concentrated in upper Montezuma Canyon where alluvial units were more distinctive and stratigraphic relationships less ambiguous. However, Pleistocene gravels and associated soils are best preserved and exposed in lower Recapture and lower Montezuma Canyons.

DESCRIPTION

Montezuma and Recapture Creeks are in the Blanding-Monticello area of San Juan County in southeastern Utah (fig. 1). Montezuma Creek drains the east flank of the Abajo Mountains, the southern part of the Great Sage Plain, and the plateaus and canyons between the Great Sage Plain and the San Juan River (fig. 2). Recapture Creek has a much smaller drainage basin rising on the south flank of the Abajo Mountains and draining a narrow strip of canyonland between the Abajo Mountains and the San Juan River (figs. 1 and 2). Elevations in both drainage basins range from about 4400 feet (1341 m) to over 11,000 feet (3353 m). However, average elevations on upland plateau surfaces which characterize the upper parts of both drainage basins are 6000-7000 feet (1829-2134 m). Montezuma Creek is incised 1000 feet (305 m) or more into this surface while Recapture Creek is generally less than 500 feet (152 m) below the plateau surface. Low relief badland-type topography is present in the lower parts of both drainages where rocks forming the resistant plateau surface have been removed by erosion to expose softer underlying units.

Because of the great elevation range, climatic conditions are quite variable within each drainage basin. Average annual precipitation decreases from about 16 inches (40.6 cm) near Monticello to 8 inches (20.3 cm) along the San Juan River (Iorns and others, 1965). Average annual precipitation at higher elevations in the Abajo Mountains exceeds 30 inches (76.2 cm), maintaining Montezuma and Recapture Creeks as perennial streams in their upper parts. In lower reaches, both are intermittent streams, flowing only during periods of snowmelt and heavy precipitation. Most precipitation is in the form of winter snowfall and intense summer thunderstorms. Average annual temperature at Blanding is about 50°F (10°C; Iorns and others, 1965).

Montezuma Canyon is rich in archeological resources and has been extensively studied by the Anthropology Department at Brigham Young University. Anasazi people occupied the canyon from around 500 A.D. to 1300 A.D. Early cultures were of the Basketmaker III stage, progressing into Pueblo stages beginning around 700 A.D. (de Haan, 1972). Settlement phases (Basketmaker III and Pueblo I) lasted until about 900 A.D., with major expansion and greatest population density from 900-1100 A.D. (Pueblo II stage). Populations declined during the Pueblo III stage beginning around 1100 A.D. with total abandonment of the canyon by 1300 A.D. (de Haan, 1972).

PREVIOUS WORK

Many regional studies of the geology of the Colorado Plateau have included observations of significance to this study. The first which addressed the physiography and Quaternary geology were by Baker (1936) and Gregory (1938). Hunt (1956, 1969) discussed Quaternary geology and drainage development of the Colorado Plateau in some detail although with no specific references to the study area. Likewise, Cooley and others (1969), Stokes (1977), and Sumsion (1975) have summarized geomorphic events and hydrology in the Navajo Indian Reservation area, principally south of the San Juan River.

Little work specifically on the Quaternary geology of the Montezuma Canyon and lower Recapture Canyon area has been done. Witkind (1964a) discussed the geology of the headwaters area and described Quaternary deposits in and around the Abajo Mountains. Probably the most detailed work prior to this study was by Huff and Lesure (1965) in relation to uranium resources in which they include a discussion of the Quaternary alluvium and other deposits of upper Montezuma Canyon. Three soil survey reports by the USDA Soil Conservation Service cover parts of the area. The San Juan area survey (Olsen and others, 1962) covers parts of Montezuma and Recapture Canyons near Monticello and Blanding. The Aneth area survey (Nielson and others, 1970) and Navajo Indian Reservation survey (Nielson and Erickson, 1980) cover lower Montezuma Canyon from about Hatch Trading Post south.

GENERAL GEOLOGY

PHYSIOGRAPHY

Montezuma and Recapture Creeks lie in the Canyonlands section of the Colorado Plateau Physio-

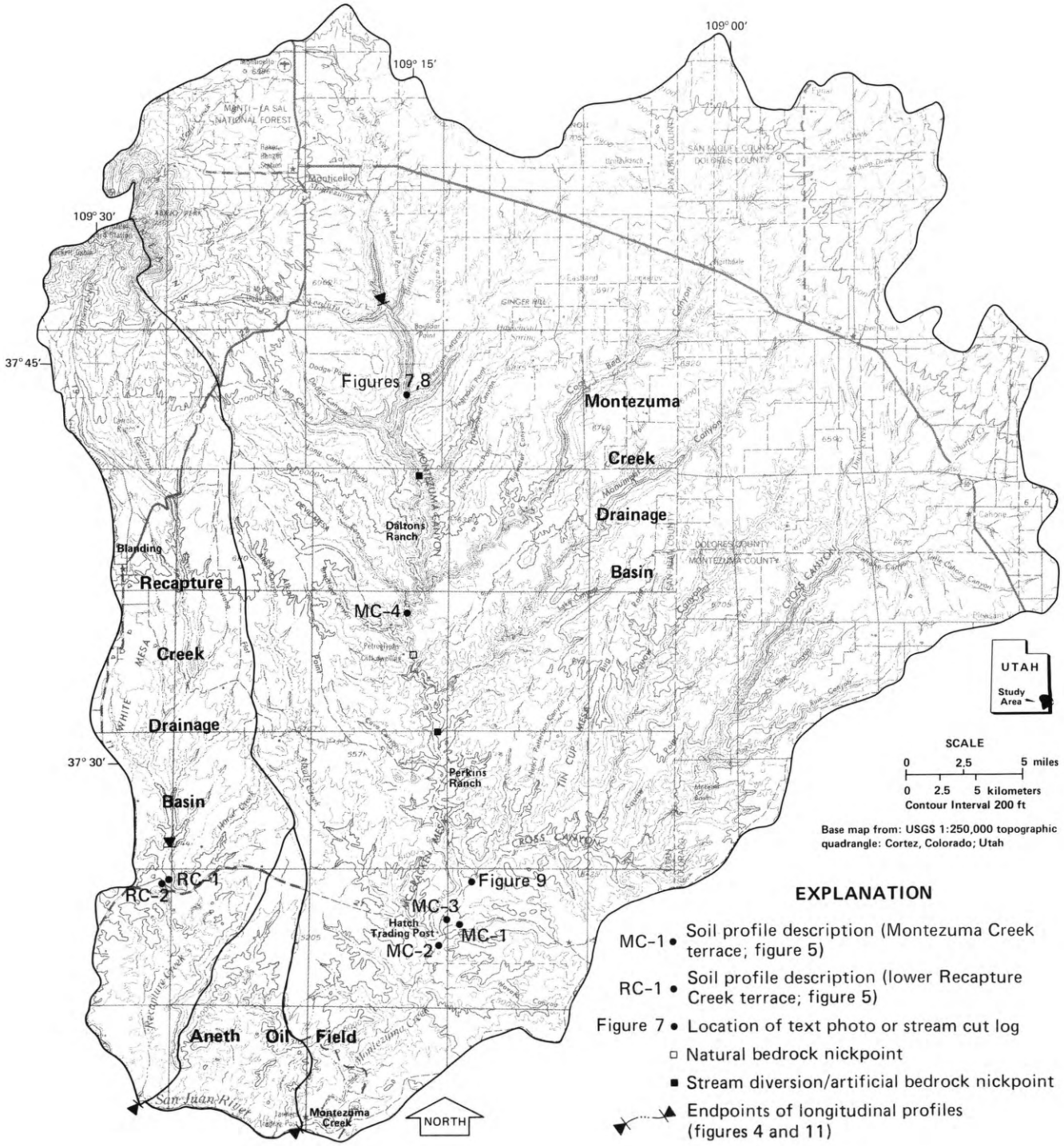


FIGURE 2. Drainage basin outlines, drainage features, and study localities, Montezuma Creek-Recapture Creek area.

graphic Province (Fenneman, 1946). Stokes (1977) has delineated physiographic subdivisions within the major physiographic provinces, and the study area lies in parts of the Abajo Mountains, Great Sage Plain, and Blanding Basin subdivisions (fig. 1). The headwaters of Recapture and Montezuma Creeks are in the Abajo Mountains, a Tertiary-age igneous complex intruded into the sedimentary rocks of the plateau. Part of the flow in Montezuma Creek is also derived from the Great Sage Plain east of the Abajo Mountains. The Great Sage Plain is an upland surface which extends into Colorado and is essentially a stripped surface on the underlying Cretaceous Dakota and Burro Canyon Formations (Stokes, 1977). Most of the Montezuma and Recapture Creek drainages are in the Blanding Basin, a broad structural basin underlain by gently dipping Mesozoic sedimentary rocks. These rocks differ greatly in erosion resistance as can be seen in canyon wall profiles and downstream changes in canyon depth, width, and steepness of walls. Resistant sandstones predominate in upper ends of both canyons, producing narrow, deep, steep-walled canyons. At the lower (southern) ends, less resistant, finer-grained rocks predominate and canyons are wide, shallow, and lack steep walls.

GEOLOGIC UNITS

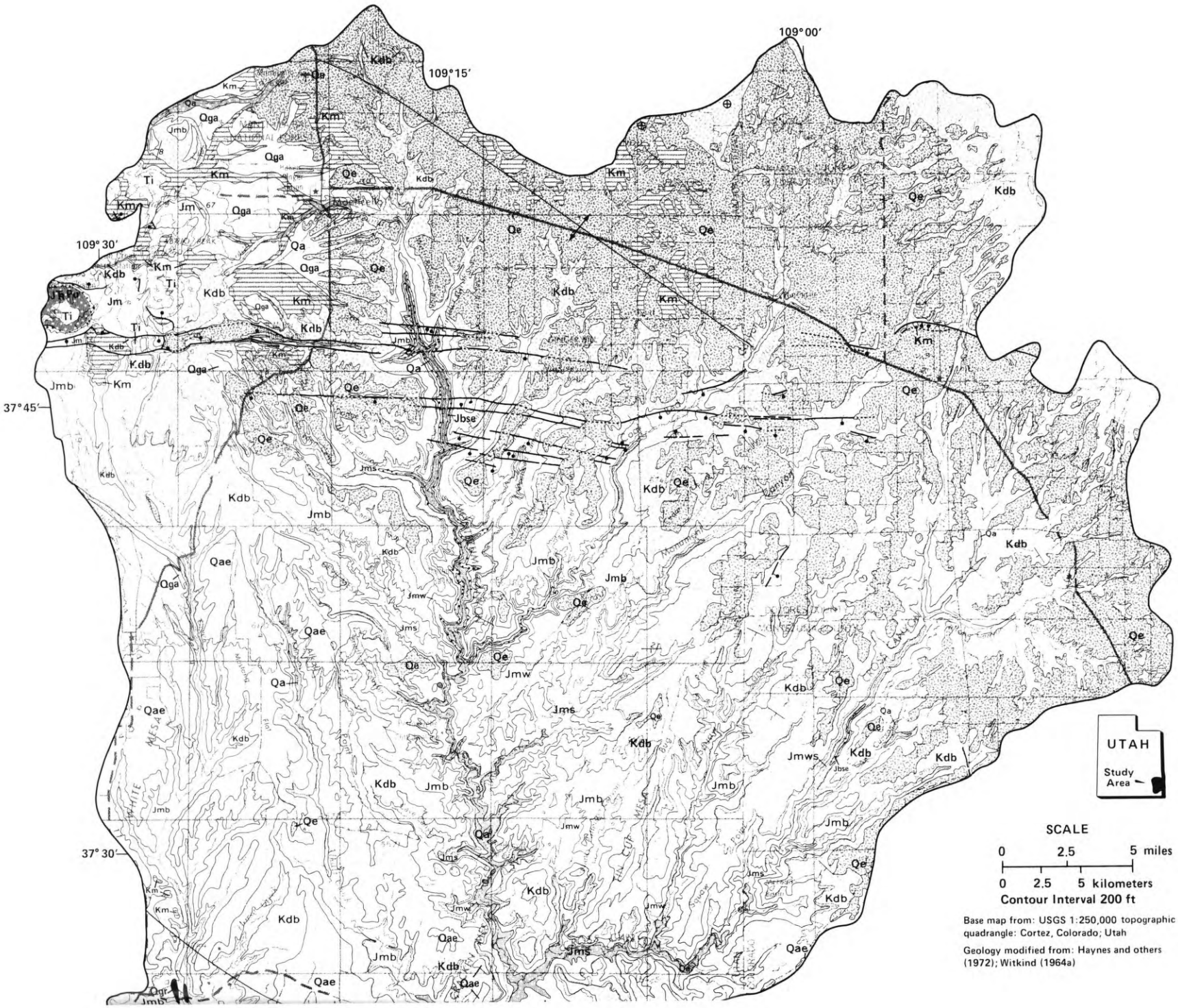
Geologic units ranging in age from Permian to Holocene crop out in the drainage basins of Montezuma and Recapture Creeks (fig. 3). The Permian Cutler and Triassic Kayenta, Wingate, Chinle, and Moenkopi Formations are exposed high in the Abajo Mountains, but only those units found in the Montezuma Canyon and lower Recapture Canyon areas will be described below. The Upper Jurassic Morrison Formation is the principal unit exposed in canyon walls throughout most of the area. The lowermost rocks in upper Montezuma Canyon include the Triassic(?)-Jurassic Navajo Sandstone and the Jurassic Carmel(?) Formation, Entrada Sandstone, and Summerville Formation which underlie the Morrison Formation. In lower Recapture Canyon, the oldest unit is the Jurassic Bluff Sandstone which directly underlies the Morrison Formation in this area. A nonresistant upper member of the Morrison Formation forms a bench at the top of upper Montezuma Canyon separating cliffs of the lower Morrison from cliffs of the overlying Cretaceous rocks. Cretaceous rocks consist of the Burro Canyon Formation and Dakota Sandstone. Remnants of Cretaceous Mancos Shale overlie the Dakota Sandstone along the base of the Abajo Mountains and on the Great Sage Plain. Igneous rocks of the Abajo Mountains do not crop out within the area studied but are important to the general geologic setting in the area because they locally affect the structure of sedimentary rocks and provide distinctive lithologies useful in studying Quaternary units. Quaternary deposits include pediment gravels, loess, eolian sand, alluvium, colluvium, and landslide deposits.

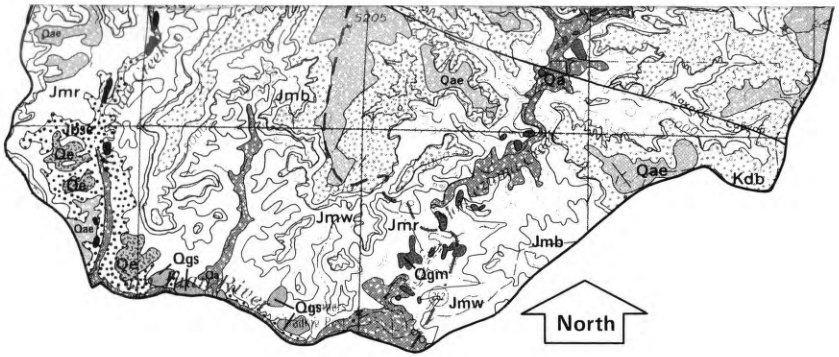
Mesozoic Sedimentary Rocks

Sedimentary rocks exposed in the area range in age from Triassic(?)-Early Jurassic to Late Cretaceous (Huff and Lesure, 1965) (fig. 3). The oldest unit is the Triassic(?)-Lower Jurassic Navajo Sandstone which is present only in a few isolated creek level exposures in Montezuma Canyon (Huff and Lesure, 1965). The Navajo is a light-colored, massive to crossbedded, eolian sandstone. Overlying the Navajo Sandstone are the Middle Jurassic Carmel(?) Formation and Entrada Sandstone. The Entrada ranges from 100-150 feet (30-46 m) thick and is very similar to the Navajo, consisting of a light-colored, massive to crossbedded, eolian sandstone forming the slickrock cliff along the bottom of Montezuma Canyon from near Dalton's Ranch northward (fig. 3). The red mudstone and siltstone of the Carmel(?) Formation is about 40 feet (12 m) thick and underlies the Entrada Sandstone in the bottom of Montezuma Canyon (Huff and Lesure, 1965). The Upper Jurassic Summerville Formation overlies the Entrada and is exposed north of Coal Bed Creek in Montezuma Canyon. This formation is chiefly an even-bedded sandstone with interbeds of reddish-brown siltstone and shale about 90-130 feet (27-40 m) thick.

In lower Recapture Canyon, the oldest unit exposed is the Upper Jurassic Bluff Sandstone which overlies the Summerville Formation. The Bluff Sandstone is a light-colored, crossbedded, eolian sandstone (Haynes and others, 1972) about 300 feet (91 m) thick which thins out to the north and east and is not present in Montezuma Canyon.

The terrestrial Upper Jurassic Morrison Formation overlies the Bluff Sandstone in Recapture Canyon and the Summerville Formation in Montezuma Canyon. The Morrison Formation has been subdivided into four members which, from oldest to youngest, are the 1) Salt Wash Member, 2) Recapture Member, 3) Westwater Canyon Member, and 4) Brushy Basin Member. The Salt Wash Member is exposed in upper Montezuma Canyon from about Perkin's Ranch northward and ranges from 320 to 520 feet (98-158 m) in thickness (fig. 3). It is a light-colored, thick-bedded sandstone which forms the upper walls of the inner canyon. Interbeds of red


109°00′
109°15′
109°30′
37°45′
37°30′
UTAH
Study Area
SCALE
0 2.5 5 miles
0 2.5 5 kilometers
Contour Interval 200 ft
Base map from: USGS 1:250,000 topographic quadrangle: Cortez, Colorado; Utah
Geology modified from: Haynes and others (1972); Witkind (1964a)



EXPLANATION

Contact

Fault

Dashed where approximately located, dotted where concealed. Bar and ball on downthrown side.

2/ inclined ⊕ horizontal

Strike and dip of beds

Syncline
Showing trace of axial plane

Anticline
Showing trace of axial plane

QUATERNARY

Holocene

Pleistocene

Qa Qe Qae Qgr Qgm Qgs Qga

- Qa – alluvium; gravel, sand, silt, and clay
- Qe – eolian sand and loess
- Qae – alluvial and eolian deposits, undifferentiated
- Qgr – terrace gravels of Recapture Creek
- Qgm – terrace gravels of Montezuma Creek
- Qgs – terrace gravels of the San Juan River
- Qga – alluvial fan and pediment gravels of the Abajo Mountains

TERTIARY

Ti

Rocks of the Abajo Mountains igneous center
Quartz diorite porphyry

CRETACEOUS

Km

Mancos Shale
Clay shale

Kdb

Dakota and Burro Canyon Formations
Sandstone and conglomerate

JURASSIC

Jm: Jmb, Jmw, Jmr, Jms, Jmws

Morrison Formation

- Jmb – Brushy Basin Member, bentonitic mudstone
- Jmw – Westwater Canyon Member, sandstone with interbedded shale and mudstone
- Jmr – Recapture Member, sandstone with interbedded siltstone and mudstone
- Jms – Salt Wash Member, sandstone with interbedded mudstone
- Jmws – Westwater Canyon and Salt Wash Members, undifferentiated

Jbse

Bluff Sandstone, Summerville Formation, Entrada Sandstone, and Carmel (?) Formation
Sandstone, siltstone, and shale

JURASSIC(?), TRIASSIC, and PERMIAN

JTRPu

Navajo Sandstone and older rocks, includes Triassic Kayenta, Wingate, Chinle, and Moenkopi Formations and Permian Cutler Formation
Sandstone, siltstone, mudstone, and shale

FIGURE 3. Geologic map, Montezuma Creek-Recapture Creek area.

mudstone form the breaks in slope between sandstone layers. The Salt Wash Member is absent in both lower Recapture and lower Montezuma Canyons, where the Recapture Member is extensively exposed. In lower Recapture Canyon, the Recapture Member lies directly on the Bluff Sandstone. In Montezuma Canyon, the Recapture Member is absent north of Hatch Trading Post but is present in lower canyon walls down to river level from about Hatch Trading Post south to the San Juan River. The Recapture Member is about 200 feet (61 m) thick and consists of interbedded reddish-gray, white, and brown sandstone and reddish-gray siltstone and mudstone (Haynes and others, 1972). It is a non-resistant unit and where exposed forms a wide, shallow canyon characterized by badland-type topography.

The Westwater Canyon Member is present in the upper walls of Montezuma and lower Recapture Canyons. It drops to stream level in Montezuma Canyon from about Hatch Trading Post to Perkin's Ranch with a maximum thickness of 180 feet (55 m) and rises steadily northward overlying the Salt Wash Member until it pinches out at Horsehead Canyon (fig. 3). The Westwater Canyon Member is an interbedded arkosic sandstone and mudstone which is generally non-resistant. The uppermost Brushy Basin Member of the Morrison Formation ranges from 150 to 700 feet (46-213 m) thick, is chiefly bentonitic mudstone with some sandstone and conglomerate lenses, and is also a non-resistant unit. In upper Montezuma Canyon these members form a bench between cliffs of the Salt Wash Member of the inner canyon and cliffs of overlying Cretaceous rocks of the outer canyon and plateau rim. The Brushy Basin Member is highly erodible and subject to slope instability and is undercutting the resistant Cretaceous Burro Canyon Formation and Dakota Sandstone which overlie it (Huff and Lesure, 1965).

The Lower Cretaceous Burro Canyon Formation is a light-colored conglomeratic sandstone with local interbedded green mudstone and limestone, and the overlying Dakota Sandstone is chiefly a yellowish brown to gray sandstone, locally conglomeratic, with interbedded gray claystone and coal (Huff and Lesure, 1965). The Burro Canyon Formation averages about 150 feet (46 m) in thickness and the Dakota about 100 feet (30 m) (Haynes and others, 1972). The uppermost sedimentary rock in the area is the Upper Cretaceous Mancos Shale. This is a gray marine shale which overlies the Dakota Formation and is exposed only locally on the flanks of the Abajo Mountains and in the Great Sage Plain where about 100 feet (30 m) of the formation is preserved beneath loess (Huff and Lesure, 1965).

Tertiary Igneous Rocks

The Mesozoic sedimentary rocks of the plateau have been intruded by a series of stocks, dikes, sills, and laccoliths forming the Abajo Mountains. The main rock type is a quartz diorite porphyry (Witkind, 1964a). A zone of shattered and tilted sedimentary rock surrounds each intrusion and has been removed to varying degrees by erosion exposing the igneous rocks. The time of intrusion has been established through potassium-argon dating as 27-28 million years ago during Oligocene time (Armstrong, 1969).

Quaternary Deposits

The oldest Quaternary deposits in the area are the alluvial-fan and pediment gravels surrounding the Abajo Mountains and extending southward as far as Dodge and Long Points (Huff and Lesure, 1965). The gravels are composed chiefly of clasts of porphyry and contain abundant subrounded boulders and cobbles, particularly near the mountains. Deposits in the mountains are up to 100 feet (30 m) thick, but generally do not exceed 20 feet (6 m) on the plateau south of Monticello (Witkind, 1964a; Huff and Lesure, 1965). Gravels were deposited on a beveled surface eroded on the Mancos Shale, Dakota Sandstone, and Burro Canyon Formation. Samples taken from sand lenses in the gravel deposits near Blanding have reversed paleomagnetic polarity, indicating deposition sometime during the Matuyama reversed epoch 0.7-2.3 m.y. ago (Biggar and others, 1982; Woodward-Clyde Consultants, 1982c). A paleosol with argillic B horizon and strongly developed calcium carbonate horizon is found at the top of the gravels which also indicates a probable Early Pleistocene age. Initial downcutting of streams into the plateau followed deposition of the gravels on which this paleosol is developed (Biggar and others, 1982; Huff and Lesure, 1965).

Loess and eolian sand cover much of the upland plateau surface throughout the area. The loess consists predominantly of very fine-grained sand with lesser amounts of silt and fine- to medium-grained sand (Huff and Lesure, 1965). Several buried soils with carbonate horizons and textural B horizons occur within loess deposits indicating episodal deposition which may extend back to Middle Pleistocene time or older (Biggar and others, 1982; Woodward-Clyde Consultants, 1982c).

Sand dunes occur extensively on the Bluff Bench near the mouth of Recapture Creek and on benches

within lower Recapture Canyon (fig. 3). Longitudinal dunes predominate and most are vegetated. Some dunes are active and are presently migrating over the canyon rim into Recapture Canyon at several localities. The sand is derived chiefly from weathering of the Bluff Sandstone. Several periods of sand movement and stabilization during Quaternary time are indicated by truncated, buried soils and by the occurrence of sand at various levels in the canyon and along the San Juan River (Oviatt, this volume).

Quaternary stream channel, flood-plain, and alluvial-fan deposits are present throughout the area. Stream gravels deposited during canyon downcutting occur at various levels in all canyons. Many of these have relict soils developed in gravels and in loess which commonly overlies the gravel. Most are believed to be of Pleistocene age. Fine-grained alluvium is present in varying thicknesses in the bottom of most canyons representing fluvial aggradation in canyons during Holocene time. Alluvial fans deposited from side-canyons constitute a part of this fill.

Quaternary colluvium and mass-wasting deposits are found along canyon walls and at the base of cliffs. Landslides are most extensive in the Brushy Basin Member of the Morrison Formation where slope failures have undermined blocks of the overlying Dakota and Burro Canyon Formations (Huff and Lesure, 1965). Rock-fall debris and talus are present at the base of most cliffs and steep canyon walls. Many recent rock-fall and debris-flow scars and deposits are present in Montezuma Canyon, particulary in the Salt Wash Member of the Morrison Formation.

STRUCTURE

In this part of the Colorado Plateau beds are nearly flat (generally less than 1° dip), folds are broad and shallow, and faults are few with small displacements (Huff and Lesure, 1965). Steep dips are present in anticlinal folds associated with the Abajo Mountains dome where bedding has been warped upward and rocks fractured by the intrusions (Witkind, 1964a). This folding occurred concurrently with intrusion in Oligocene time (Armstrong, 1969). To the south is the Blanding Basin, a broad syncline in which the arcuate axis trends generally east to intersect Montezuma Creek near Hatch Trading Post and Recapture Creek near the Highway 262 crossing (fig. 3). An extension of the east-trending Dove Creek anticline may pass through northern Montezuma Canyon beneath the Great Sage Plain and connect to the Abajo Mountains dome (Huff and Lesure, 1965). The age of this folding is not known but the folding was most likely concomitant with the Monument upwarp during early Tertiary time (Hunt, 1956). Huff and Lesure (1965) cite evidence that folding may have begun during Early Cretaceous time.

An east-west zone of block-faulting south of the Abajo Mountains crosses northern Montezuma Canyon and includes the Verdure graben and several smaller grabens and normal faults. The north Verdure fault in Montezuma Canyon has a maximum throw of 182 feet (55 m) (Huff and Lesure, 1965). The age of the faulting is not known but intrusions along the north Verdure fault indicate that it predates the Oligocene Abajo Mountains intrusives (Witkind, 1964b; Huff and Lesure, 1965; Witkind, 1975).

LATE QUATERNARY GEOLOGY

MONTEZUMA CREEK

Description

Montezuma Creek, which drains about 1200 mi^2 (3108 km^2) of southeastern Utah and southwestern Colorado, is a major tributary of the San Juan River. The cities of Monticello, Utah and Dove Creek, Colorado are in the headwaters area and the settlement of Montezuma Creek and the Aneth oil field on the Navajo Indian Reservation are at the mouth on the San Juan River. The highest point in the basin is Abajo Peak (elevation 11,360 feet-3463 m) in the Abajo Mountains west of Monticello. Montezuma Creek flows into the San Juan River about elevation 4400 feet (1341 m). Except for the east flank of Abajo Peak, the entire drainage basin is underlain by relatively flat-lying sedimentary rocks. Montezuma Creek has been divided into an upper (above Perkin's Ranch) and lower (below Perkin's Ranch) part (fig. 2) for this discussion.

Upper Montezuma Creek has cut a deep canyon about 1100 feet (335 m) below the plateau surface, here underlain by the Dakota and Burro Canyon Formations. These resistant cliff-forming sandstone units are underlain by the bench-forming Brushy Basin and Westwater Canyon Members of the Morrison Formation, forming an upper bench and outer rim of the canyon. Inner canyon walls are formed principally by the bedded sandstones of the Salt Wash Member of the Morrison Formation and the Summerville Formation and by the massive Entrada Sandstone. The canyon continues to widen by mass-wasting processes as evidenced by numerous fresh rock-fall and debris-flow scars, particularly in the Salt Wash Member. Erosion of the massive Entrada Sandstone is more by granular disintegration than mass movement.

Lower Montezuma Canyon is generally less than 500 feet (152 m) deep. Non-resistant rocks of the Recapture, Westwater Canyon, and Brushy Basin Members of the Morrison Formation form the canyon walls giving it a wide bottom with no narrow inner canyon.

The average width of Montezuma Canyon as measured between the Dakota-Burro Canyon escarpments on either side varies from 2 miles (3.2 km) north of Dalton's Ranch up to about 8 miles (12.8 km) near Coal Bed Canyon and back to about 1.5 miles (2.4 km) from Perkin's Ranch south (fig. 3). The great width in the central part of the canyon is a result of the removal of the Dakota-Burro Canyon Formations on divides between the several major tributaries in this area. Tributaries in the headwaters and lowermost parts of the canyon are smaller and do not contribute as significantly to the retreat of the rim escarpment.

A continuous alluvial fill extends the length of Montezuma Canyon, and the creek flows in a channel incised from 15 to 50 feet (5-15 m) into this fill. Total thickness of the fill is not known, but it generally exceeds the depth of incision. Bedrock is locally exposed in the creek bottom but only crosses the creek forming a natural nickpoint at one locality between Bradford and Tank Canyons (fig. 2). Alluvium is about 17 feet (5 m) thick above the nickpoint and 34 feet (10 m) thick below it. If the original channel of Montezuma Creek followed the same course it does today, then the canyon in this area was never deeper than at present. However, the valley bottom is wide here and alluvium may bury a deeper channel to the west. In two places artificial nickpoints have been formed where the stream has been diverted against a bedrock canyon wall to be channeled into irrigation ditches (fig. 2).

Upper Montezuma Canyon follows a meandering course in which the radii of curvature of meanders and width of the meander belt are larger than those of the modern stream. Canyon meanders are presumably inherited from those of ancestral Montezuma Creek as it flowed across the plateau surface. Stream discharge has decreased since that time. Lower Montezuma Canyon does not exhibit this pronounced underfit character, perhaps indicating lithologic control over meander preservation. The presence of the resistant Salt Wash Member, Summerville Formation, and Entrada Sandstone in upper Montezuma Canyon have preserved the inherited meandering course while the less resistant rocks (Recapture, Westwater Canyon, and Brushy Basin Members) in lower Montezuma Canyon have not. Montezuma Creek presently flows in a set of irregular meanders entrenched into the alluvial fill in the canyon bottom. A longitudinal profile showing the modern stream gradient and height of various terraces above modern stream grade is shown in figure 4. The modern stream gradient of Montezuma Creek averages about 22 feet/mile (4.2 m/km). The average gradient of upper Montezuma Creek is 25 feet/mile (4.8 m/km) while that of the lower portion is 18 feet/mile (3.4 m/km). The depth of the modern arroyo in upper Montezuma Canyon averages about 30 to 40 feet (9-12 m) with local depths as much as 50 feet (15 m). Incision depth decreases in lower Montezuma Canyon to about 15-20 feet (5-6 m). The profile shown in figure 4 for the upper terrace fill does not represent a paleogradient since measurements were made along the modern channel. While perhaps not greatly different, the stream length at the time of deposition of this fill is not known.

Deposits

Late Quaternary deposits in Montezuma Canyon consist of alluvial gravel, sand, silt, and clay and eolian sand, silt, and clay. Quaternary landslide deposits and various alluvial-fan, colluvial, and terrace cover (slope wash) deposits are also present but were not studied for this report.

Older alluvium Older alluvial deposits consist chiefly of thin gravels on rock-cut terraces in lower Montezuma Canyon. Gravels are relatively resistant and cap bedrock knobs and benches at various levels above the stream. In upper Montezuma Canyon, the walls are steep and actively eroding, and conditions are not conducive to development or preservation of gravel terraces.

Gravels consist predominantly of rounded clasts of intrusive rocks from the Abajo Mountains (80%) and of various sandstones exposed along the canyon (20%). The highest level above stream grade in Montezuma Canyon at which Montezuma Creek gravels were found is about 280 feet (85 m) in a terrace near the mouth. Gravels generally occur from about 20 feet (6 m) to 150 feet (46 m) above river level. Except for a 40-foot (12 m) terrace near the mouth of Alkali Creek, continuous paired terrace levels are absent (fig. 4). Remnants of the 40-foot terrace extend only for about a mile south of Hatch Trading Post and are not in evidence elsewhere. Gravels at 90-100 feet (27-30 m) are found in several isolated localities. Thin terrace gravels mantel much of the badland terrain of lower Montezuma Canyon and this, along with poor topographic control and lack of soil exposures, makes correlations of terrace

levels very difficult. North of Hatch Trading Post most gravel deposits have been removed for road construction and original gravel thicknesses are unknown. Many gravels overlie resistant sandstone beds, and downcutting appears to have been largely controlled by lithology and erosion resistance of underlying rocks.

Soil profiles exposed in terrace gravels at 20 feet (6 m), 40 feet (12 m), 70 feet (21 m), and 120 feet (37 m) were studied to estimate the minimum age of the deposits (fig. 5). The soil profile of the lowest terrace is exposed in a canal cut near stream level and the profiles of the upper three are exposed in gravel pits or road cuts. In all cases, exposures are poor and profiles are incomplete and exhibit considerable lateral variation. Descriptions shown in figure 5 are thus highly generalized. Upper A and B horizons are generally removed or disturbed in these profiles. Secondary calcium carbonate horizons are best preserved and the morphology and thickness of these horizons are considered the most diagnostic characteristics of these soils. Diagnostic morphologies of the various stages of development of pedogenic calcium carbonate horizons used in this report are given in table 1.

Gravel terraces generally exhibit the expected increase in soil profile development with age and terrace height. The 120-foot (37 m) terrace contains a relatively thin K horizon (stage IV carbonate) with an upper laminar layer about 1 inch (3 cm) thick underlain by a plugged zone about 5 inches (13 cm) thick. This is underlain by a thick Ck horizon with zones of manganese and iron oxide staining near the base. The A and B horizons have largely been removed (fig. 5). The lack of a thick plugged zone beneath the laminar layer indicates that the laminar layer may have developed prematurely as a result of erosion of the upper parts of the carbonate horizon and is not a true indication of stage IV pedogenic carbonate (Gile and others, 1966; Lattman, 1973).

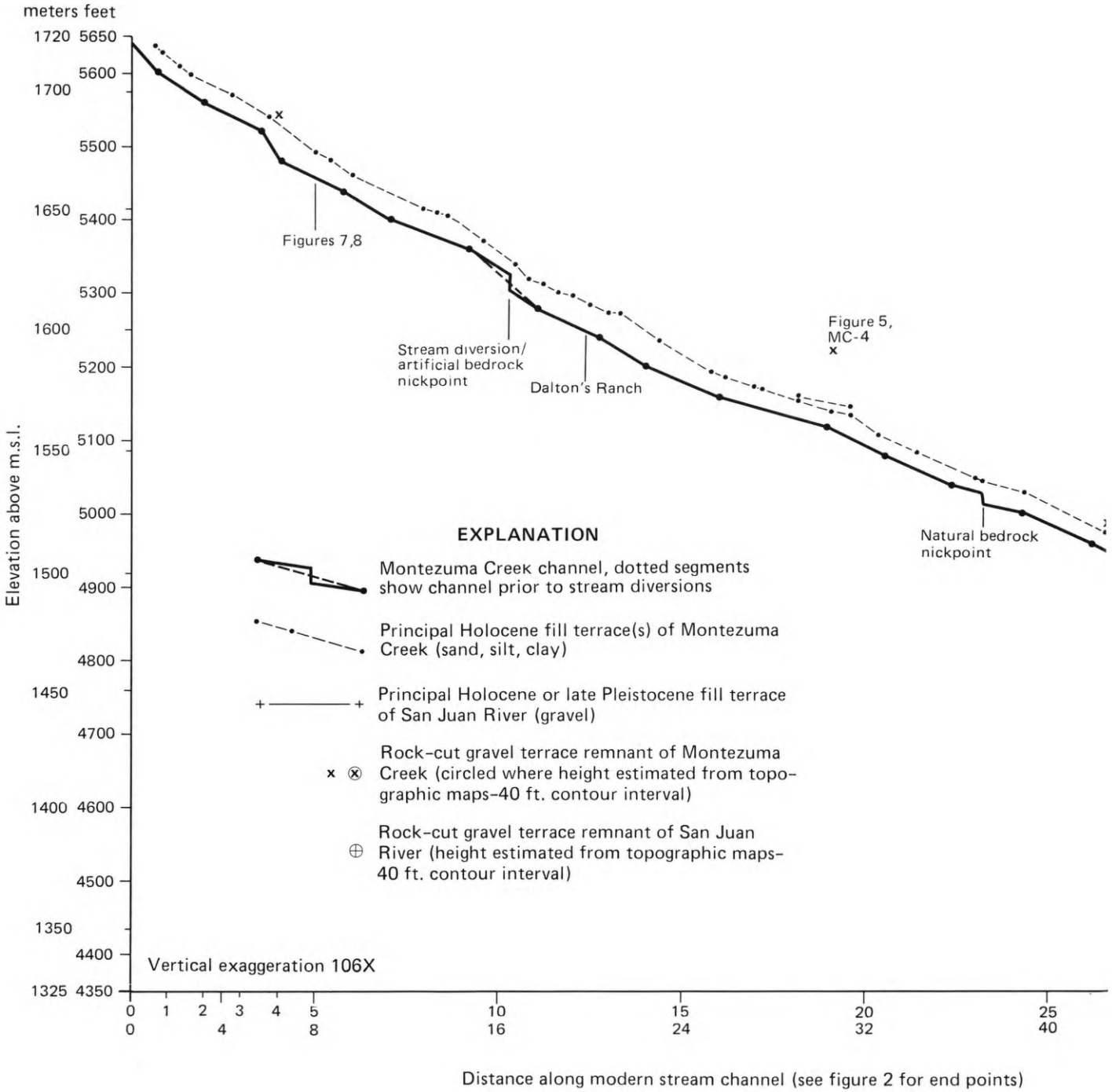
TABLE 1. *Stages in the morphogenetic sequence of carbonate deposition in calcic soils (from Gile and others, 1966, as modified by Bachman and Machette, 1977).*

STAGE	DIAGNOSTIC CARBONATE MORPHOLOGY
I	Filaments or faint coatings. Thin, discontinuous coatings on lower surface of pebbles.
II	Firm carbonate nodules few to common but isolated from one another. Matrix may include friable interstitial accumulations of carbonate. Pebble coatings continuous.
III	Coalesced nodules and pebble coatings in disseminated carbonate matrix.
IV	Platy, massive indurated matrix. Relict nodules may be visible in places. Plugged. May have weak incipient laminae in upper surface. Case hardening common on vertical exposures.
V	Platy to tabular, dense and firmly cemented. Well-developed laminar layer on upper surface. May have scattered incipient pisoliths in laminar zone. Case hardening common.
VI	Massive, multilaminar and brecciated, with pisoliths common. Case hardening common.

Profiles on lower terraces are also undergoing rapid erosion. A soil in terrace materials 70 feet (21 m) above Montezuma Creek contains a carbonate horizon with stage II-III carbonate development in the upper 24 inches (60 cm) and stage I-II development in the lower part. Soils in a 40-foot (12 m) terrace contain a stage II-III carbonate horizon (fig. 5). Calcium carbonate root casts and a laminar calcium carbonate layer overlying shale bedrock at a depth of 63 inches (160 cm) are present locally in the profile.

Soil development in a terrace 20 feet (6 m) above modern stream grade is exposed in a canal cut north of Hatch Trading Post (fig. 2). Carbonate development appears anomalous in this profile, being generally thicker and more advanced in terms of stage of development than that in some higher terraces. The soil is developed in gravels overlain by sand of eolian and alluvial origin. A stage III K horizon about 12 inches (30 cm) thick, predominantly in sandy deposits, overlies a 81-inch (205 cm) thick carbonate horizon (Ck) in sands and gravels with stage I-III carbonate development. This exposure is at and below the level of the younger fine-grained alluvium in Montezuma Creek, whereas other terraces are above this younger fill. The deposit was periodically influenced by ground water and stream flow during the Holocene, and the seemingly anomalous carbonate content may be attributable to ground-water and stream-deposited carbonate, with case-hardening of the exposed face (Lattman and Simonberg, 1971; Lattman, 1973).

Richmond (1962), Harden and others (1982), and Woodward-Clyde Consultants (1982a) have studied soil profile development versus age in gravelly alluvium in Spanish Valley near Moab, Utah. Lithology of the gravels (predominantly porphyritic intrusive rocks with some sandstone clasts) and weathering environments in the Spanish Valley area are very similar to those in the Montezuma-Recapture Creek area. Average annual rainfall is slightly greater (15


meters feet
1720 5650
5600
1700
5500
1650 5400
5300
1600
5200
1550 5100
5000
1500 4900
4800
1450
4700
1400 4600
4500
1350
4400
1325 4350
Elevation above m.s.l.
Figures 7,8
Stream diversion/
artificial bedrock
nickpoint
Dalton's Ranch
Figure 5,
MC-4
Natural bedrock
nickpoint
EXPLANATION
Montezuma Creek channel, dotted segments show channel prior to stream diversions
Principal Holocene fill terrace(s) of Montezuma Creek (sand, silt, clay)
Principal Holocene or late Pleistocene fill terrace of San Juan River (gravel)
Rock-cut gravel terrace remnant of Montezuma Creek (circled where height estimated from topographic maps-40 ft. contour interval)
Rock-cut gravel terrace remnant of San Juan River (height estimated from topographic maps-40 ft. contour interval)
Vertical exaggeration 106X
0 1 2 3 4 5
0 4 8
10 15 20 25
16 24 32 40
Distance along modern stream channel (see figure 2 for end points)

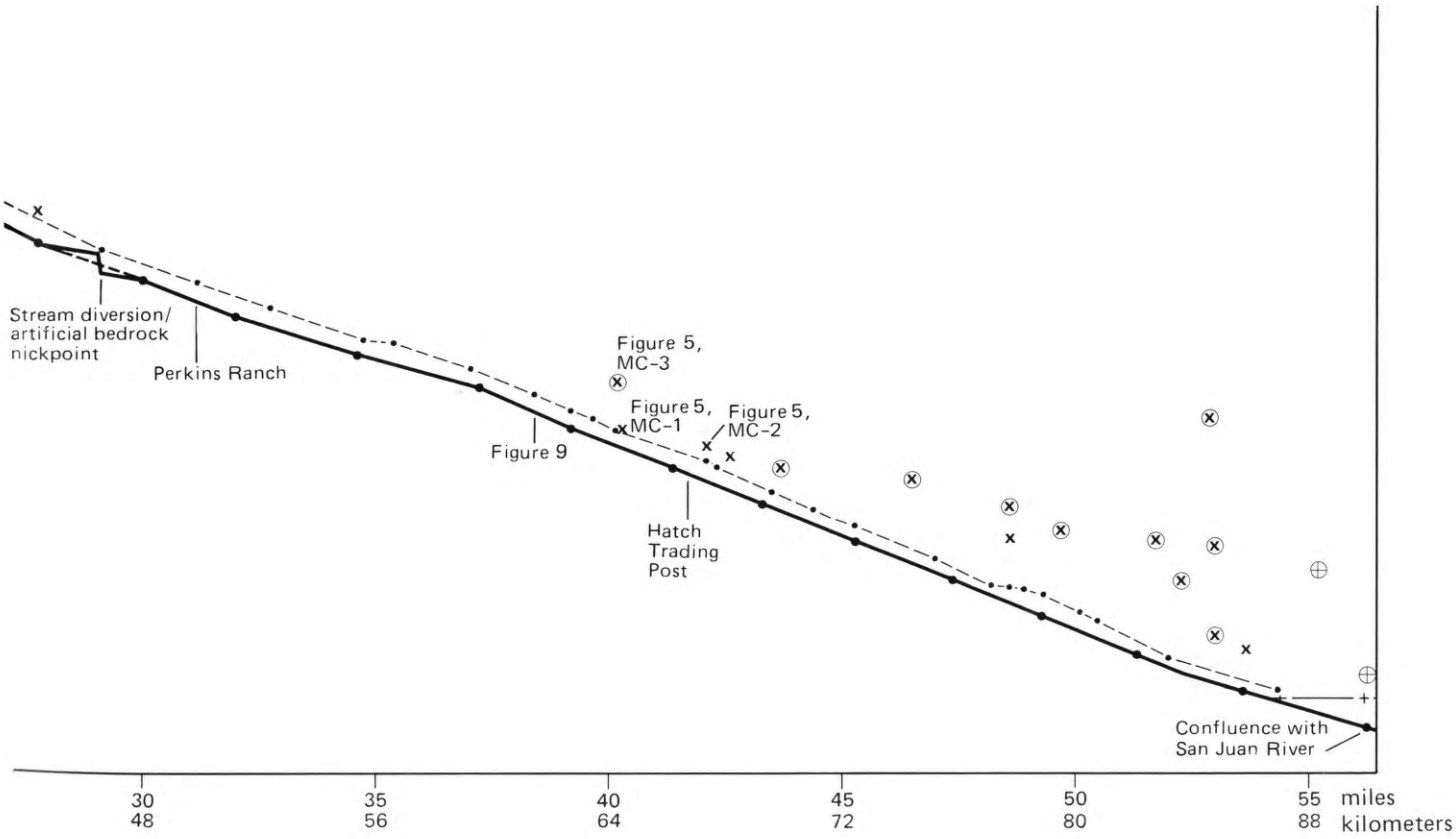


FIGURE 4. Longitudinal profile of Montezuma Creek showing terrace heights.

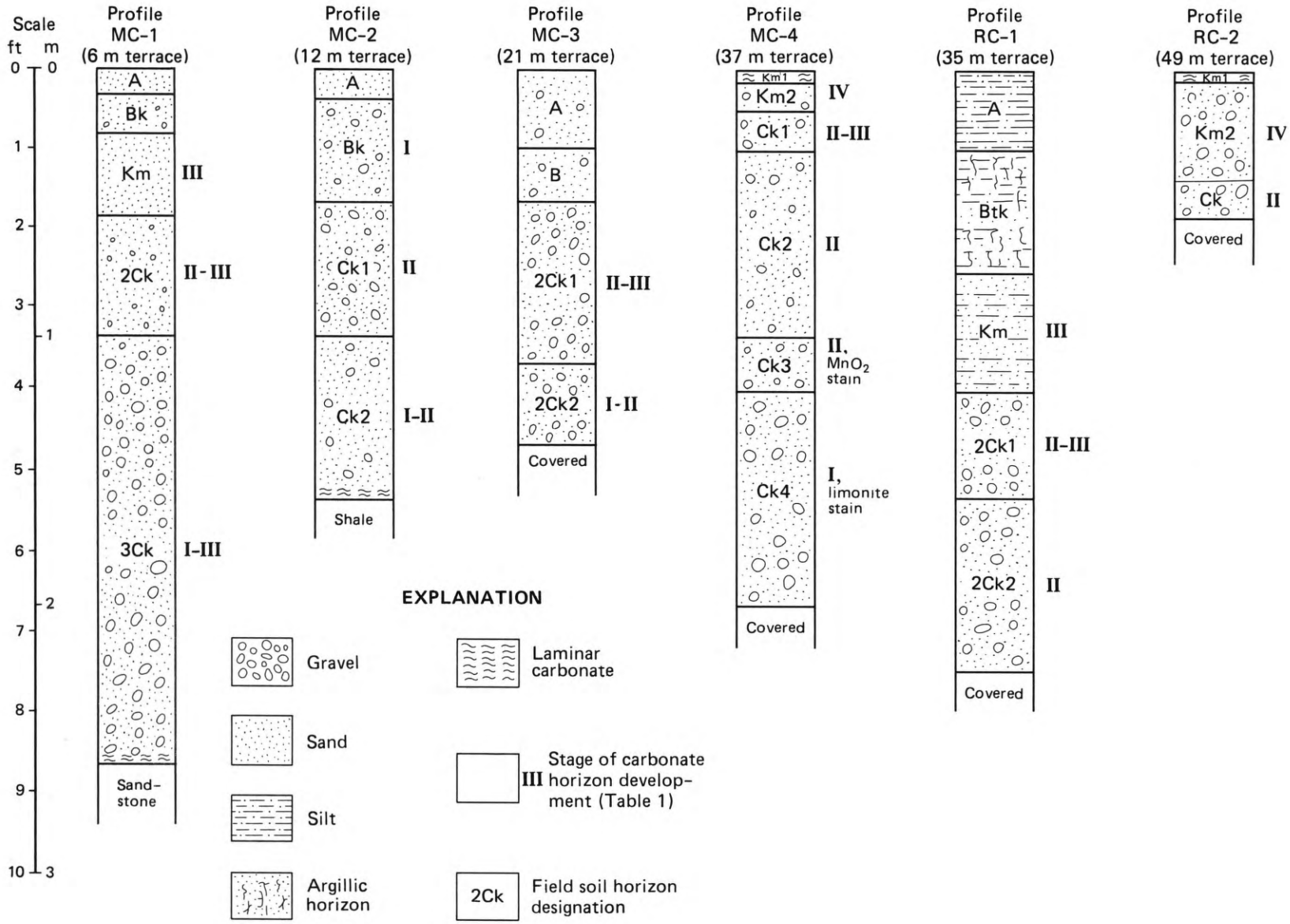


FIGURE 5. Soil profiles in gravel terraces, Montezuma and Recapture Creeks.

inches-38 cm) than that in the area of lower Montezuma and Recapture Canyons (10 inches-25 cm) where most soil profiles were described for this study. However, both areas probably underwent similar climatic changes during the Quaternary. Because of this, comparisons of soil profile development in the two areas provide approximate ages for terrace gravels in the Montezuma-Recapture Creek area.

Because upper soil horizons are missing in many exposures in Montezuma-Recapture Canyons, carbonate horizon development was relied on most in comparisons. Soils in terrace gravels from 20 to 120 feet (6-37 m) above modern creek level appear to be equivalent in development to soils of the Beaver Basin and Placer Creek Formations of Richmond (1962), Harden and others (1982), and Woodward-Clyde Consultants (1982a) indicating a Late to Middle Pleistocene age. Because of inconsistencies between relative profile development and geomorphic position, the general poor quality of exposures, and the incomplete and irregular nature of profiles exposed, no attempt was made to assign a more precise age for these deposits.

Younger Alluvium The alluvial fill terrace which extends the length of Montezuma Canyon is composed of deposits representing several periods of deposition. Little surface expression of these various deposits is present except for a single terrace which extends for about a mile (1.6 km) between Coal Bed and Devil Canyons about 5 feet (1.5 m) above the level of the principal fill (fig. 6B). Otherwise, the only terraces are those within the modern arroyo left as the stream cut down to its present position. In general, long distance correlations of the various deposits exposed in stream cuts in the fill can only be made with assurance through correlation of terrace surfaces or through absolute age dating. In Montezuma Canyon, distinct terrace levels are not associated with each deposit and dates are few. Correlations made in this report are based on local lithologic similarities and stratigraphic relationships, and they must be considered tentative until more dates are available. Generalized cross sections showing stratigraphic relationships and making tentative correlations of units at various localities along Montezuma Creek are shown in figure 6.

Radiocarbon dates on detrital charcoal samples in these units have been obtained from one locality (figs. 7 and 8). Radiocarbon ages were determined by Beta Analytic, Inc. (Libby half-life 5568 years) and have been converted to calendar years according to Klein and others (1982) as shown in figure 8. Where discussed in the text, however, radiocarbon ages are used for ease of reference and because most of the referenced works use them.

At least two periods of aggradation separated by incision and downcutting of the stream are apparent in stream cuts in upper Montezuma Canyon. A photograph and log of a stream cut showing typical deposits of these two units are shown in figures 7 and 8, respectively. The earlier fill (Unit 1) represents the first period of major canyon alluviation during Holocene time. This fill is best exposed in upper Montezuma Canyon, particularly between Verdure Creek and Dodge Creek (fig. 6A). It is quite uniform lithologically, consisting in its lower part of thick interbeds of black to blue-gray clays and red to yellow sands. Sandy layers contain varying amounts of silt and clay but locally are devoid of fines. Deposits grade upward into more thinly bedded materials of similar lithology (fig. 7). Cut and fill structures and channel sands and gravels are generally not found in Unit 1, and it appears to represent continuous aggradation in a low-energy environment, probably a broad flood plain which extended for great distances along the stream. Deposits contain considerable charcoal and a radiocarbon date about 28 feet (8.5 m) below the terrace surface (8 feet-2.4 m above creek level) yielded an age of 5130 $\pm$ 200 B.P. (fig. 8). Following deposition of Unit 1, the stream cut down to near the level of the modern channel. In upper Montezuma Canyon, this downcutting was to within 3 feet (1 m) of present stream level (fig. 6A). In lower Montezuma Canyon, beds with lithologies typical of Unit 1 have not been identified.

Following incision of Unit 1, alluviation of the channel began. This fill (Unit 2) is characterized by a basal channel sand and/or gravel locally containing mud balls derived from Unit 1. Basal deposits of Unit 2 have been dated at 1430 $\pm$ 80 B.P. (fig. 8). These basal deposits are overlain by massive to bedded silty and clayey sand with minor clay layers, particularly near the top. Cut and fill channel sands are common throughout Unit 2, which filled back to or slightly below the level of Unit 1. A date on the upper part of Unit 2 is 1410 $\pm$ 80 B.P. (fig. 8) indicating rapid filling of the channel. Eolian sand or interbedded eolian and alluvial sand about 3 feet (1 m) thick or less commonly occurs at the top. Locally in upper Montezuma Canyon a prominent unconformity within Unit 2 is apparent (fig. 6B) indicating a period of stability and erosion within this unit. However, channel cuts into Unit 2a are generally less than 6 feet (2 m) deep, and the unconformity elsewhere is parallel to bedding in both units. This break indicates a local hiatus in depo-

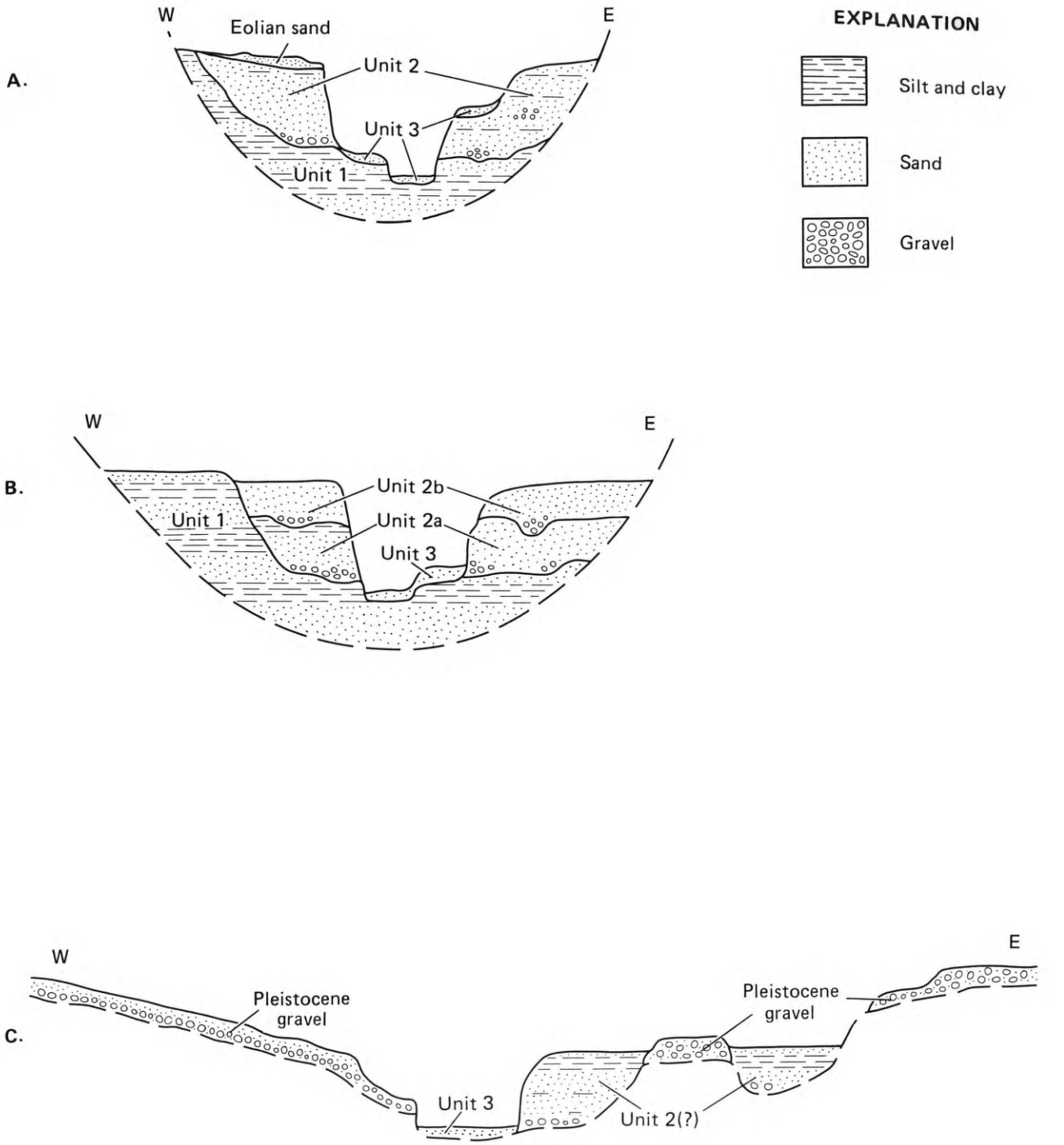


FIGURE 6. Diagrammatic cross sections showing Quaternary stratigraphic relationships in Montezuma Creek alluvium: a) between Verdure and Dodge Canyons, b) near Coal Bed Canyon, and c) near Hatch Trading Post.

FIGURE 7. Stream cut north of Pearsons Canyon showing Units 1 and 2.

sition between Unit 2a (below unconformity) and Unit 2b (above unconformity) which was not associated with deep stream incision or soil formation.

In lower Montezuma Canyon, deposits are much less distinctive and cut and fill relationships are not apparent. The thick clay beds of Unit 1 disappear downstream, probably as a result of source-area influences. The Mancos Shale, believed to be the source of these clays, is present in the headwaters of tributaries in upper Montezuma Creek but is generally absent in tributaries south of Coal Bed Creek (fig. 3). These southern tributaries contribute chiefly sand to Montezuma Creek, resulting in a relative decrease in clay content downstream. As clay content decreases, deposits lose their distinctive character, and correlation of deposits in lower Montezuma Canyon with either Unit 1 or 2 in upper Montezuma Canyon based on lithology is generally not possible. A common sequence observed in stream cuts in lower Montezuma Canyon consists of a channel sand and gravel, locally containing mud balls at the base, overlain by massive silty and clayey sands, in turn overlain by a dark clay bed (fig. 9). Eolian dune and sheet sands up to 8 feet (2.4 m) thick locally overlie the clay bed. The mud balls found in basal deposits may have been derived from Unit 1 deposits upstream, and it is thought that most deposits in lower Montezuma Canyon are younger than Unit 1 and may correlate with Unit 2. However, evidence is poor and no dates are available. Huff and Lesure (1965) report fresh-water clams and snails and land snails in this fill along Alkali Wash in lower Montezuma Canyon. Shells were not dated, but all species still exist although under conditions slightly more moist than at present at this locality.

Unit 3 was deposited during modern downcutting which followed deposition of Unit 2 and the overlying eolian materials. A series of terraces was formed within the modern arroyo as the stream became incised to the present level (fig. 6). These terraces consist of clean, loose, coarse-grained sand (Unit 3), generally less than 10 feet (3 m) thick, on surfaces cut in the older alluvial fill. Unit 3 includes modern channel sands and gravels and flood-plain deposits. The time of inception of arroyo cutting and deposition of Unit 3 is not known, although photographs taken in 1875 at Coal Bed Canyon indicate that incision of the modern arroyo had not yet begun. Williams (1925) states that incision began about 1900 A.D. and in 3 years had cut the full length of the stream (Huff and Lesure, 1965). Matheny (1962) quotes a report from near Long Canyon that "when white man began to settle in Montezuma Canyon in 1922, the stream was gentle and clear and only a foot or so below the surface of the land." Reports of the exact time of incision thus conflict but it probably occurred in the early 1900's and deposits of Unit 3 (terraces in the inner arroyo and modern channel) are thus less than 100 years old. An irrigation diversion dam north of Pearson's Canyon reportedly built in 1936 and breached shortly thereafter is presently about 20 feet (6 m) above the modern channel level,

EXPLANATION

SOIL TYPES

- Silt and clay
- Sand
- Gravel

RADIOCARBON SAMPLES

	Radiocarbon age*	Calibrated age**
△△△△△△△	1410 ± 80 B.P.	1165 - 1515 B.P.
+++++++	1430 ± 80 B.P.	1175 - 1530 B.P.
xxxxxx	5130 ± 200 B.P.	5585 - 6345 B.P.

*Dates by Beta Analytic, Inc., Coral Gables, Florida

**Radiocarbon years converted to calendar years according to method of Klein and others (1982)

Vertical Scale: ft 0, 3, 6, 9, 12, 15; m 0, 0.9, 1.8, 2.7, 3.7, 4.6

Horizontal Scale: ft 0 3 6 9 12 15; m 0 0.9 1.8 2.7 3.7 4.6

No vertical exaggeration

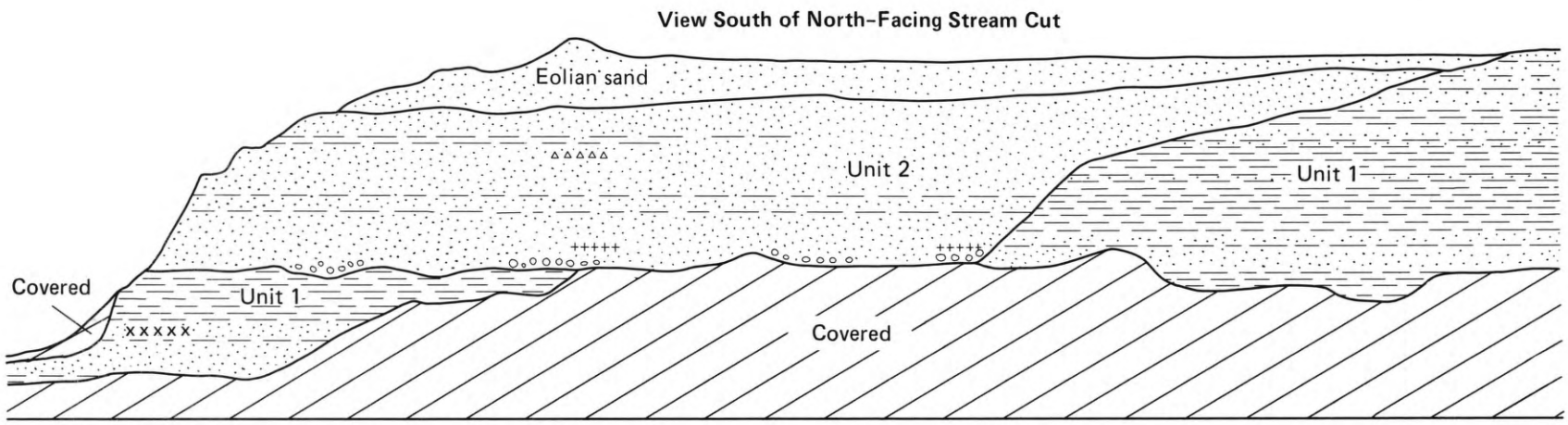


FIGURE 8. Log of stream cut north of Pearsons Canyon (left half of cut in fig. 5) showing typical deposits of Units 1 and 2 and radiocarbon sample locations.

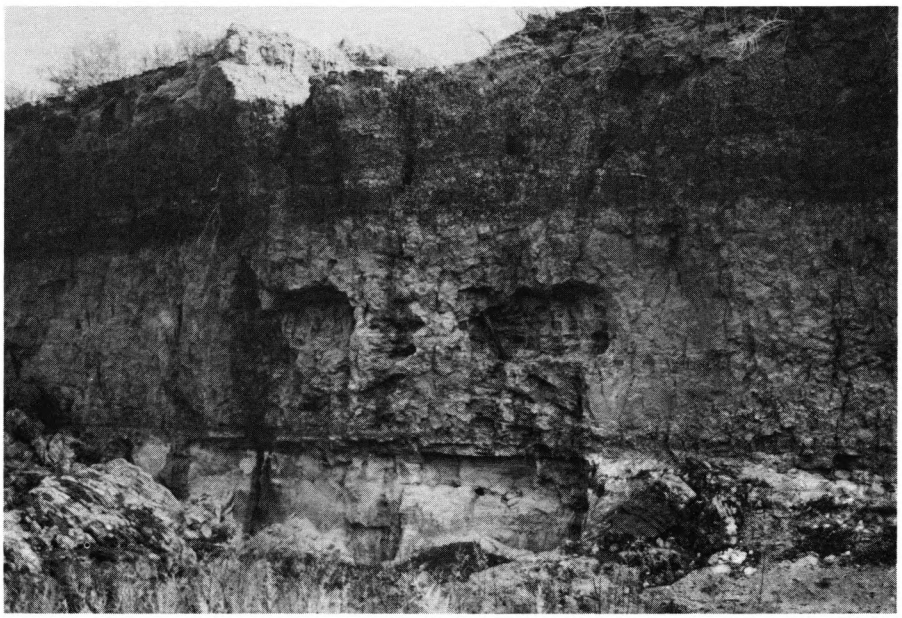


FIGURE 9. Stream cut at Cross Canyon showing typical alluvial deposits in lower Montezuma Creek.

indicating downcutting of 20 feet (6 m) in the last 40-45 years. A long-time resident reported that up to 6 feet (2 m) of downcutting occurred during heavy runoff in the mid-1970's. Accelerated downcutting in response to floods in 1977 is documented elsewhere, although generally not of this magnitude (Cooley, 1979).

Geomorphic History

Downcutting of Montezuma Canyon began following deposition of the alluvial fan and pediment gravels now found on upland surfaces around the Abajo Mountains in Early Pleistocene time more than 700,000 years ago. The oldest Montezuma Creek gravels are found about 280 feet (85 m) above the present river level in lower Montezuma Canyon near the San Juan River. The age of these gravels is not known, and no exposures of soil profiles were found. Soil profiles are exposed in lower terraces 20 feet to 120 feet (6-37 m) above creek level. Exposures are poor and data are insufficient to assign ages except to say all are Late to Middle Pleistocene.

Based on a minimum age of 700,000 years for the Abajo Mountains alluvial-fan and pediment gravels (Biggar and others, 1982; Woodward-Clyde Consultants, 1982c) with incision of Montezuma Canyon beginning at this time and an average present depth of upper Montezuma Canyon of 1100 feet (335 m), an average maxiumum downcutting rate of 1.6 feet (0.5 m)/1000 years is indicated for this part of the canyon. Rates have probably varied greatly through geologic time. The occurrence of strongly developed soil profiles in lower terraces indicates greatly reduced rates during late Quaternary time. The Holocene was characterized by alternating degradation and aggradation with little additional bedrock downcutting. Rates of retreat of the Dakota-Burro Canyon escarpment along Montezuma Creek as indicated by canyon widths in areas not affected by closely spaced tributaries average between 5.7 to 7.5 feet (1.7-2.3 m)/1000 years).

The sequence of events following initial downcutting of the canyon is shown in figure 10. A period of aggradation began shortly before 5130 B.P., the date on the lowermost deposits exposed. The thickness of these deposits is unknown but bedrock occurs locally in the modern creek bed indicating the thickness near the center of the valley may not greatly exceed the depth of incision. Original alluviation of the canyon (deposition of Unit 1) ended sometime before 1430 B.P. Incision of this fill was complete by 1430 B.P. with alluviation (deposition of Unit 2) beginning about that time and continuing until after 1410 B.P. The age and time span of the local unconformity within Unit 2 is not known. Montezuma Creek flowed at near the level of the principal terrace from sometime after 1410 B.P. until around 1900 A.D. During this time, eolian sand dunes and sheet sands were deposited on the flood plain. These deposits probably represent eolian reworking of local chan-

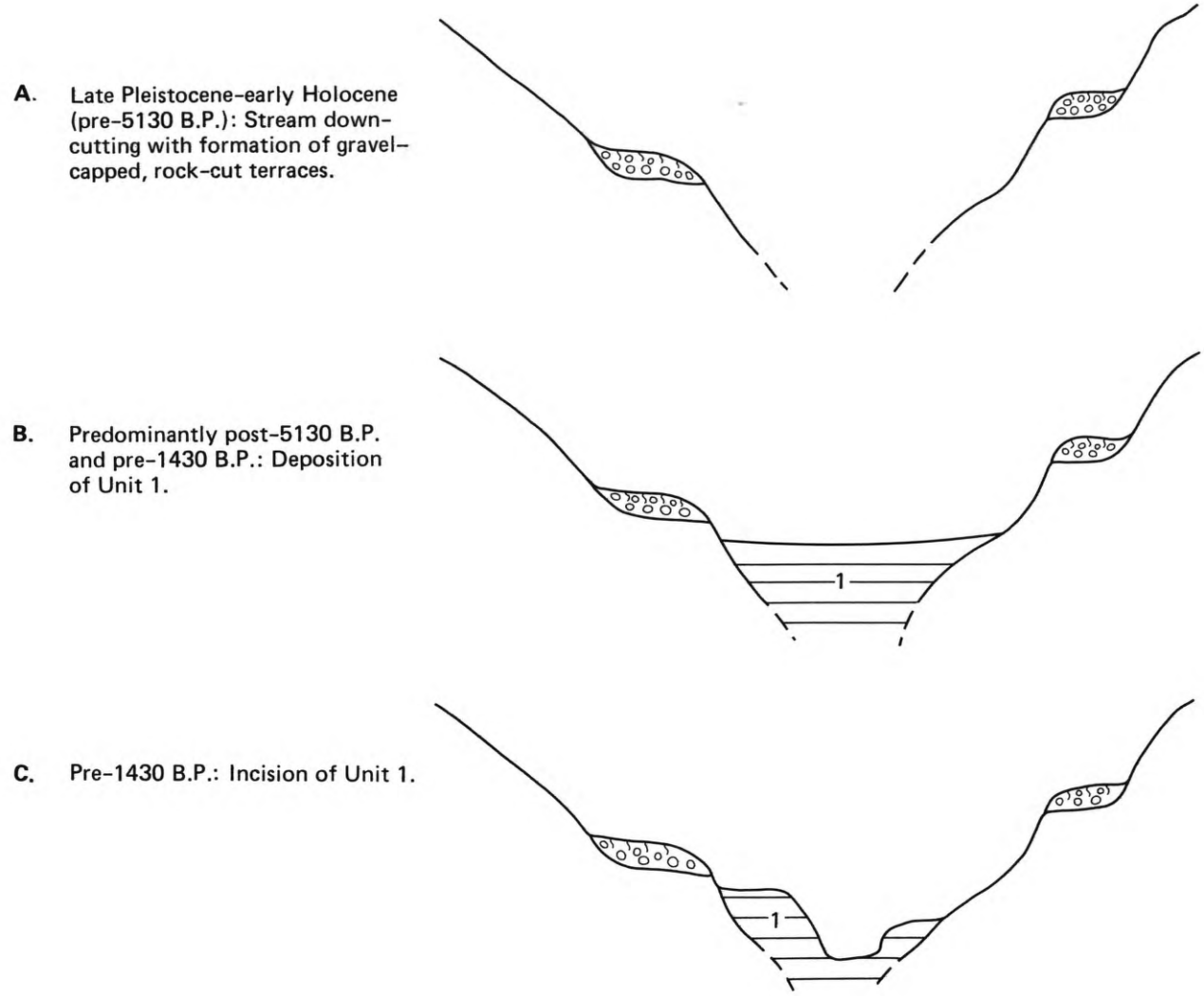


FIGURE 10. Schematic cross sections showing probable sequence of Holocene events in upper Montezuma Canyon.

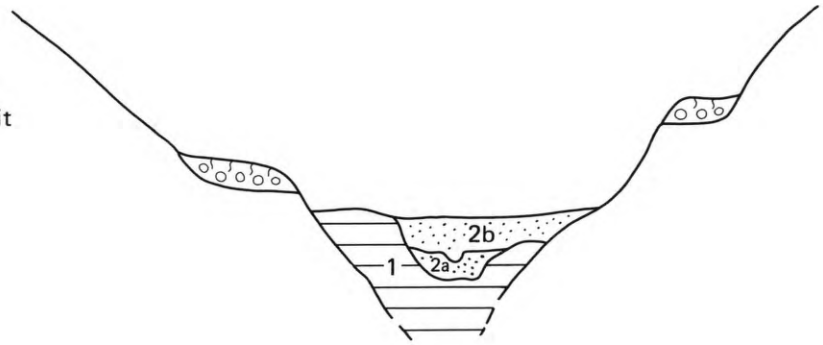
nel and flood-plain sands.

Montezuma Creek has cut to the present level of the San Juan River at its mouth. Upstream, one natural bedrock nickpoint and two man-made diversions forming artificial bedrock nickpoints control downcutting (fig. 2). However, lateral bank cutting occurs at rapid rates during periods of high flow, undercutting roads and destroying cropland, principally on the outside of meander bends. Erosion of the alluvial fill also occurs through extensive soil piping and subsequent gully development following the collapse of soil pipes.

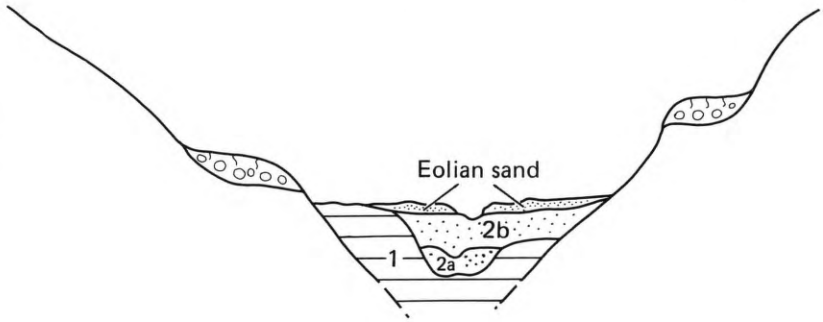
Archeology and Impact of Man

Arroyo cutting during the late 1800's and early 1900's has occurred throughout the American southwest in response to a variety of factors (Cooke and Reeves, 1976). The influence of man, particularly in terms of land use, is considered one of the important causes. While it was not a purpose of this study to evaluate causes of modern arroyo cutting, Matheny (oral communication, April, 1981) and de Haan (1972) consider it to have been at least in part a result of grazing and cultivation, both on the plateau surface and the canyon floor. Whatever the cause, the effects of arroyo cutting are adverse with respect to agriculture as fertile cropland is eroded and the water table is lowered. In some cases, farmland has been abandoned because of modern arroyo cutting, and Hall (1977) cites evidence that earlier arroyo-cutting cycles may have been responsible for abandonment of Chaco Canyon in northwestern New Mexico by the Anasazi around 1150 A.D.

D. Post-1430 B.P.: Deposition of Unit 2, locally consisting of several cut and fill sequences.



E. Post-1410 B.P. to about 100 B.P.: Stability of surface, accumulation of eolian sand.



F. Post-100 B.P.: Incision of Units 1 and 2, deposition of Unit 3.

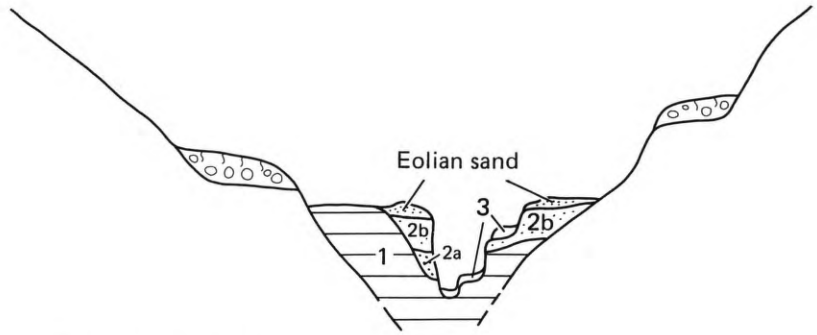


FIGURE 10. (Continued)

Previous erosion cycles in Montezuma Canyon may have affected Anasazi settlement patterns, although no evidence was found to indicate an arroyo-cutting episode near 1300 A.D. when the canyon was abandoned. Deposition of Unit 1 predates settlement by the Anasazi in Montezuma Canyon. Incision of Unit 1 was probably complete prior to their arrival as well. Deposition of Unit 2 after 1430 B.P. (555 A.D.) may coincide roughly with the arrival of the Anasazi in Montezuma Canyon around 500 A.D. Basketmaker III (500-700 A.D.) sites are found on the Holocene alluvial fill surface in upper Montezuma Canyon near Dodge Canyon (Matheny, 1962) where it is believed to be underlain by Unit 2 deposits. However, in a survey downstream from Tank Canyon to Hatch Trading Post, de Haan (1972) notes that no Basketmaker III and very few Pueblo I sites (700-900 A.D.) are found on the Holocene fill surface (de Haan, 1972). In contrast, Pueblo II (900-1100 A.D.) sites are concentrated on this surface and agriculture was practiced. Basketmaker III sites in this area are found only on the Pleistocene gravel stream terraces.

This distribution of sites may be solely a result of the cultural settlement patterns noted by de Haan (1972). However, it may also indicate that the majority of the Holocene fill in lower Montezuma Creek is a Unit 2 or younger fill and that deposition was not complete during Basketmaker III time, either preventing settlement or burying evidence of settlements here. Plotting of early Anasazi sites and their relationships to the various Holocene fills throughout the canyon, as well as a detailed search of the alluvial fill for archeological materials reworked and deposited with the sediment or buried by it, may aid in eval-

uating any possible influences alluviation may have had on settlement patterns. More corroborative radiocarbon dates on alluvial stratigraphic units, particularly in lower Montezuma Canyon, are required before any firm conclusions can be drawn.

LOWER RECAPTURE CREEK

Description

Recapture Creek drains about 210 mi^2 (544 km^2) from the south slope of Abajo Peak to the San Juan River. The drainage basin is long and narrow, generally only 6-7 miles (10-11 km) wide. The highest point in the basin is over 11,200 feet (3414 m) above sea level in the Abajo Mountains. The confluence of Recapture Creek and the San Juan River is at elevation 4340 feet (1323 m).

This study covered only the lower segment of Recapture Canyon from the Highway 262 crossing south to the San Juan River (fig. 2). Bedrock units exposed in this part of the canyon are the Brushy Basin, Westwater Canyon, and Recapture Members of the Morrison Formation and the Bluff Sandstone. The canyon lacks steep cliffs except in the lowermost part where low cliffs of Bluff Sandstone flank Recapture Creek. Although lower Recapture Creek flows at depths up to 900 feet (274 m) below the adjacent plateau surface, the cliffs of the Dakota and Burro Canyon formations have retreated a mile or more from the stream and canyon relief along Recapture Creek is generally less than 250 feet (76 m).

Recapture Canyon generally lacks the large radii canyon meanders found in upper Montezuma Canyon, and follows a much straighter course. Rock types exposed in both areas are similar, but the modern stream gradient of lower Recapture Creek is 36 feet/mile (6.9 m/km), about twice as steep as lower Montezuma Creek (fig. 11). This difference in gradient is probably a major factor controlling the differences in sinuosity of the two streams (Harden, 1982), giving Recapture Creek a much straighter course lacking incised meanders.

Deposits

Late Quaternary deposits in lower Recapture Canyon consist of alluvial gravel, sand, silt, and clay and eolian sand, silt, and clay. Eolian deposits are more extensive than in Montezuma Canyon, and large areas of active and inactive dunes are found in the Bluff Bench area.

Older alluvium Gravels are found throughout lower Recapture Canyon on rock-cut terraces at various levels above the modern stream (fig. 11). The highest Recapture Creek terrace gravels occur at heights of about 220-230 feet (67-70 m) above modern stream level. An extensive group of terraces marks the contact between the Brushy Basin and underlying Westwater Canyon Members of the Morrison Formation along Highway 262 at heights of 110-160 feet (34-49 m) above modern stream level. Here, an upper sandstone of the Westwater Canyon Member is extensively mantled with gravels, indicating a lithologic control over terrace development. Differential preservation may also play a role, because terraces cut in shale beds would be removed much faster than those cut in sandstone. Gravels near the mouth of Recapture Creek occur at various levels within the massive Bluff Sandstone, indicating that lithologic controls are not the only factors influencing terrace development. Consistent gravel terrace levels were not noted and the cutting of most terraces, particularly in the Bluff Sandstone, may have occurred randomly during downcutting.

Gravel clasts in terrace deposits are predominantly intrusive rocks of the Abajo Mountains and various sandstones exposed along the canyon. Soil profile development is quite advanced in some of the higher terraces. Descriptions of soils developed on gravel and loess on terraces 160 feet (49 m) and 115 feet (35 m) above the modern channel are shown in figure 5. The profile at 160 feet (49 m) has been truncated by erosion and gravel pit operations. It contains a 16-inch (40 cm) stage IV K horizon consisting of a thin laminar layer overlying a plugged zone. The situation is similar to that of the 120-foot (36 m) terrace in Montezuma Canyon, and the laminar layer may have formed prematurely during erosion of the upper part of the carbonate horizon. The soil profile on the 115 feet (35 m) terrace is the most uniform and complete profile found in the Montezuma-lower Recapture Canyon area. Parent material consists of about 3 feet (1 m) of sandy loess overlying Recapture Creek gravels. An 18-inch (45 cm) argillic B horizon and 18-inch (45 cm) stage III K horizon occur chiefly in loess, grading downward into cemented gravels (Ck horizon with stage II-III carbonate). Significant soil development in gravel prior to deposition of loess is not apparent. The profile is roughly equivalent in development to those on Placer Creek deposits of Richmond (1962), Harden and others (1962), and Woodward-Clyde Consultants (1982a), believed to be roughly 150,000 B.P. Gravels would be somewhat older since time must be allowed for deposition of loess prior to soil formation.

Younger alluvium The lowermost gravel ter-

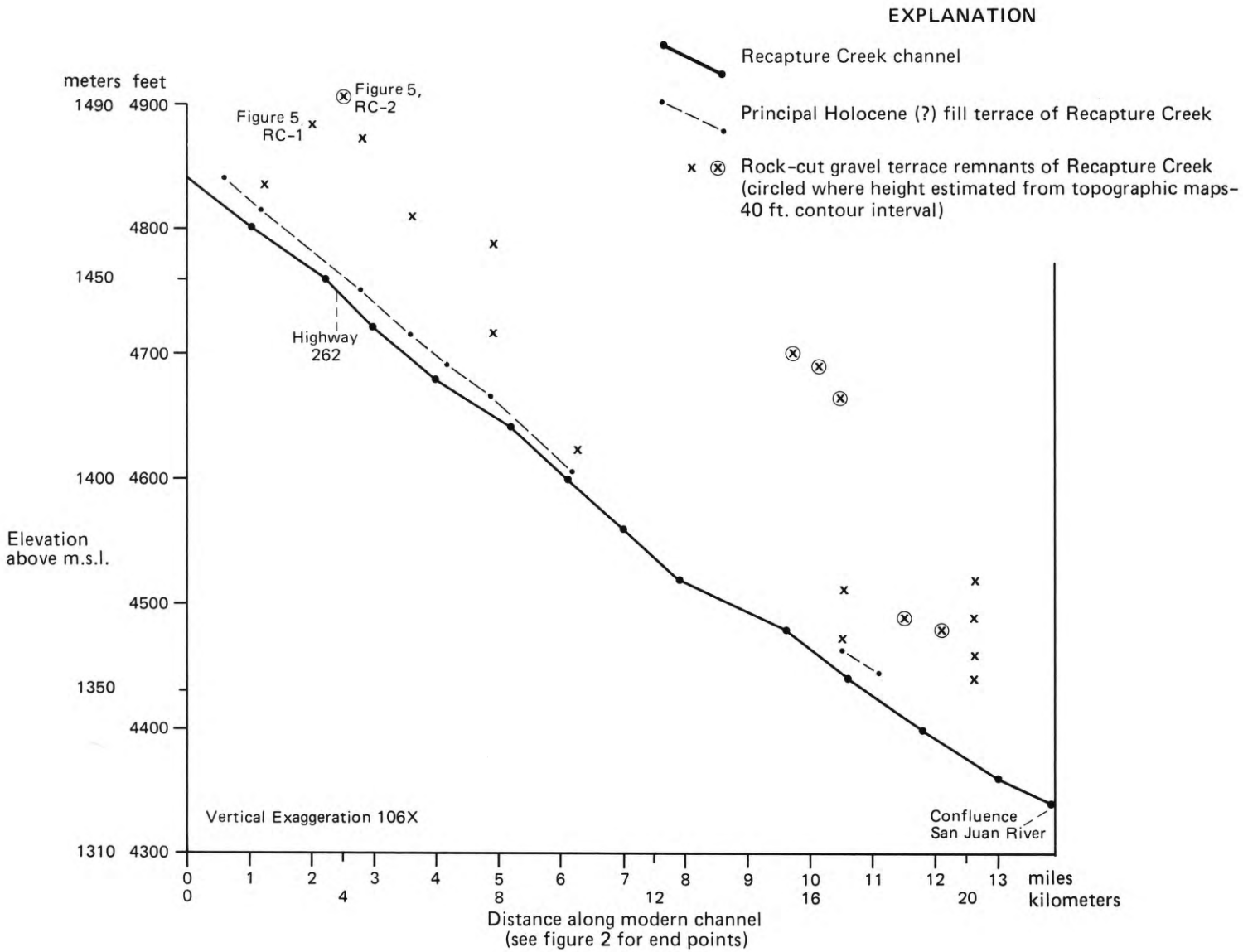


FIGURE 11. Longitudinal profile of lower Recapture Creek showing terrace heights.

races in lower Recapture Canyon are at heights from 20 (6 m) to 30 feet (9 m) above stream level. Sandstone bedrock occurs beneath these gravels indicating that they are on rock-cut terraces. A soil profile exposed in a gravel deposit about 22 feet (7 m) above the channel is weakly developed with stage I carbonate coatings beneath clasts, suggesting a probable early Holocene age.

Recapture Canyon exhibits a single finer-grained late Holocene(?) fill similar to that in Montezuma Canyon. This fill was not studied in detail, but deposits generally lack stratification and no cut and fill relationships were observed. Deposits consist of sand, silt, and clay with gravel lenses which extend 17-22 feet (5.2-6.7 m) above stream level. Holocene deposits are coarser than their counterparts in lower Montezuma Canyon, perhaps due to the steeper gradient of Recapture Creek.

Eolian. Eolian sand and loess are found extensively in lower Recapture Canyon. The largest area of eolian sand caps the Bluff Sandstone in the Bluff Bench area at the mouth of Recapture Creek. Sand on the bench is derived from alluvium, older eolian deposits, and weathering of the Bluff Sandstone. Broad areas of sheet sands and longitudinal dunes are found here. Several periods of eolian activity are indicated by the presence of a truncated and buried paleosol in older alluvium and eolian deposits underlying an upper, loose, stabilized sand layer. Eolian sand and loess also occur within Recapture Canyon on mesas and ridges underlain by the Morrison Formation. Most deposits are stabilized by vegetation but some are still active, drifting across roads and over cliffs into Recapture Canyon. Gravel terraces are commonly overlain by loess or eolian sand. The soil profile description of the 115-foot (35 m) terrace in figure 5 shows 4 feet (1.2m) of eolian material overlying gravels. These eolian materials are strongly cemented with carbonate, exhibit a strongly developed argillic B horizon, and post-date deposition of the gravels. Still older eolian deposits are found capping the plateau surface and higher benches, exhibiting various buried paleosols indicative of several eolian episodes as discussed earlier (Woodward-Clyde Consultants, 1982a).

Geomorphic History

Downcutting of Recapture Creek probably began about the same time as in Montezuma Creek, following deposition of alluvial-fan and pediment gravels surrounding the Abajo Mountains. In lower Recapture Canyon, the oldest gravels in the canyon are found 220-230 feet (67-70 m) above stream grade. A complete soil profile in deposits 115 feet (35 m) above Recapture Creek indicates an age of perhaps 150,000 years. This would yield an average downcutting rate of about 0.8 feet (0.24 m)/1000 years. Downcutting continued with development of gravel-capped rock-cut terraces at various levels down to about 20-30 feet (6-9 m) above modern stream level. These lowermost gravel terraces are perhaps early Holocene in age. Following formation of these terraces, incision continued to a level below that of the modern stream. Deposition of finer-grained, presumably Holocene material followed this downcutting, filling the channel to a level 17-22 feet (5-7 m) above the modern stream or nearly to that of the lowermost gravel terrace. The time of incision of this Holocene fill is unknown, but the degree of preservation of stream cuts and the similarity to those of lower Montezuma Creek indicate it probably represents the late 1800-early 1900 arroyo-cutting episode better documented in Montezuma Canyon. No bedrock nickpoints were observed in lower Recapture Creek, and the total depth of fill is not known.

Eolian sand is present at various levels within Recapture Canyon. Some, such as that on the Bluff Bench, exhibit truncated carbonate horizons buried by uncemented sand and indicate an early stage of eolian activity, perhaps soon after the canyon was cut to that level. The overlying uncemented sands are generally stabilized with vegetation, but some are presently active. The older cemented sands are probably Late Pleistocene or early Holocene in age, whereas the overlying stabilized dunes are probably mid to late Holocene.

REGIONAL CORRELATION OF ALLUVIAL DEPOSITS

The Quaternary record of Montezuma and lower Recapture Creeks indicates a history generally similar to that of other streams on the Colorado Plateau. Studies of late Quaternary alluvium on the Colorado Plateau have been done: 1) the San Juan River (Oviatt, this volume), Cedar Mesa (Agenbroad, 1975; Salkin, 1975), and Cottonwood Creek (Woodward-Clyde Consultants, 1982c) areas of Utah to the west, 2) the Indian Creek (Woodward-Clyde Consultants, 1982b) and Spanish Valley (Richmond, 1962; Harden and others, 1982; Woodward-Clyde Consultants, 1982a) areas of Utah to the north, 3) Animas Valley (Gillam, 1982) and Chaco Canyon (Bryan, 1954; Hall, 1977; Love, 1977; Wells and others, 1982) of southwestern Colorado and northwestern New Mexico to the east, and 4) Jeddito Wash and Tsegi Canyon (Hack, 1942; Karlstrom, 1982) and the Little Colorado River (Hereford,

1984) in northeastern Arizona to the south. Table 2 shows dates and chronologies worked out for these areas where available.

The generally parallel alluvial histories in these areas are apparent from table 2. Middle and Late Pleistocene time was characterized by active stream erosion in this part of the Colorado Plateau with formation of rock-cut terraces as rivers entrenched their channels into bedrock. This overall erosional regime was reversed during at least part of the Holocene as streams aggraded. Much of the Holocene is typified by alternating deposition and erosion of non-gravelly alluvium with little additional bedrock downcutting.

The similarities in geomorphic histories of many Colorado Plateau streams raises a question as to the extent to which individual deposits are correlative. A long-standing theory that climatic change is the dominant factor controlling erosional and depositional episodes and that such episodes are synchronous over large areas is presently under debate. A number of studies have been done in which long-distance correlations over parts of the western U.S. are made (Leopold and Snyder, 1951; Leopold and Miller, 1954; Miller and Wendorf, 1958; Haynes, 1968; Karlstrom, 1982). Studies by Ritter (1974), Patton and Schumm (1981), Wells and others (1982), McDowell (1983), Knox (1983), and Begin and Schumm (1984) have indicated a complex response by alluvial systems both to climate change and the influences of other factors such as vegetation, lithology and base level. They suggest that perhaps long-distance correlations are not valid and that, even given synchroneity of climate change over periods of a thousand years or less, different drainage basins and even different segments of a particular drainage may not react at the same time or in the same manner.

In order to evaluate the various hypotheses, accurate dating of alluvial deposits is required. Dating of Pleistocene gravel terraces, generally through soil profile development, is generally of insufficient accuracy to attempt to make or evaluate the validity of long-distance correlations. Correlations of alluvial gravel deposits with glacial deposits have been made in streams draining glaciated areas, but the relationship between alluvial deposits and glacial stages in streams draining unglaciated areas is less well known. Most streams draining unglaciated areas of the Colorado Plateau exhibit a series of gravel-capped terraces cut during Pleistocene time which are controlled as much by bedrock lithology as climate (Woodward-Clyde Consultants, 1982a, b, c). This appears to be the case in the Montezuma-Recapture Creek area as well. A nearly continuous series of terraces with few well-developed, prominent levels is present. Where present, paired or prominent terrace levels are usually controlled by lithology and cap resistant sandstone layers. However, some areas with multiple terrace levels such as near the mouth of Recapture Creek in the massive, homogeneous Bluff Sandstone exhibit no apparent local lithologic control. These terraces may have formed in response to regional climatic, tectonic, or base level control. However, they may also represent random, localized events during downcutting. The numerous terraces of lower Montezuma Creek occur in non-resistant rocks and in many cases have been removed by man or are evidenced only by a remnant gravel mantle. This makes correlation and evaluation of controlling influences very difficult since it is possible that prominent levels did exist at one time but are now obscured.

The change from Pleistocene gravel deposition on rock-cut surfaces to Holocene aggradation of sand, silt, and clay in previously cut channels is apparent here as elsewhere in the southwestern U.S. The lowermost gravels observed are about 20 feet (6 m) above modern stream grade, or about the same level as the top of the Holocene fill in the same areas. Haynes (1982) has suggested that channel deposits of sand and gravel which commonly mark the transition from abandonment of the last Pleistocene terrace to vertical accretion during the Holocene in many parts of the western U.S. may be considered a lithostratigraphic boundry between the Pleistocene and Holocene Epochs. These channel sands and gravels conformably underlie fine-grained Holocene deposits (Haynes, 1982), and presumably overlie (unconformably) bedrock in the channel bottom. This basal channel sand and gravel unit in Montezuma Creek, if present, underlies Unit 1 of the Holocene fill and thus predates 5130 B.P. It would post-date the lowermost gravel terrace which is of Late Pleistocene age in Montezuma Canyon. Thus, in Montezuma Canyon, as elsewhere, the transition from the Pleistocene into the Holocene was marked by a change from predominantly gravel deposition in an erosional regime to deposition of sand, silt, and clay in an aggradational regime. The exact time of the change is unknown, but it may have taken place during early Holocene time since the 5130 year date is believed to be near the base of the Holocene sequence. In lower Recapture Canyon, limited soil profile data indicate that gravel deposition on rock-cut terraces may have continued into the early Holocene.

The Holocene depositional sequence in Montezuma Canyon shows certain similarities with sequences

worked out elsewhere (table 2). Dates on Units 1 and 2 in Montezuma Canyon fall within the time ranges given for deposition of the Tsegi formation in northern Arizona (Karlstrom, 1982). As in Montezuma Canyon, deposits of the Tsegi Formation and deposits of that time period elsewhere commonly include two separate depositional episodes (Hack, 1942; Leopold and Miller, 1954; Miller and Wendorf, 1958; Haynes, 1968). The Gallo and Chaco units in Chaco Canyon also cover the same general time period as the two-fold Tsegi sequence (Hall, 1977). A detailed sequence of events remains to be worked out in other areas and, in general, only isolated dates on alluvium from various streams are available from the other localities on the Colorado Plateau shown in table 2. From the dates available, however, deposition in these areas in some cases was not synchronous with deposition in Montezuma Canyon, Chaco Canyon, and northeastern Arizona.

As can be seen in table 2, erosional episodes are considered to be relatively short compared to depositional episodes. Euler and others (1979) have compiled a paleoclimatic scale for the last 2250 years on the Colorado Plateau in which they define four wet periods during which alluvial deposition took place and five dry periods of erosion. Erosional periods were thought to last from 50-100 years, although more recent work has indicated longer erosional periods of 150-300 years (T. Karlstrom, oral commun., November 1982). In either case, they were short compared to depositional episodes and it is not surprising that deposition was occurring in many areas at the same time, whether controlled by regional climate change or not. It is more instructive in terms of evaluating factors controlling stream regimen to attempt correlation of the shorter-term erosional episodes. Data from Montezuma Creek indicate erosional episodes from 0-100 B.P. (Unit 3), shortly before 1430 B.P. (Unit 1/Unit 2), and before 5130 B.P. (pre-Unit 1). The modern erosional period is well documented in many areas. The pre-1430 B.P. episode in Montezuma Creek may correlate with erosional episodes shortly pre-dating this time in several areas, but sequences are not sufficiently well dated to make firm conclusions regarding synchroneity.

Early Holocene deposits appear to be absent in Montezuma Canyon, although it is not known whether this is due to non-deposition, erosion, or burial. Non-gravelly alluvial deposits of the time period from about 7000-10,000 B.P. have been found in nearby Elk Ridge (Woodward-Clyde Consultants, 1982c) and Gibson Dome (Woodward-Clyde Consultants, 1982b). Elsewhere, the oldest non-gravelly deposits are 5000-7000 years old, and the early Holocene is marked perhaps by continued gravel deposition or by erosion and soil formation (Hack, 1942; Hall, 1977; Karlstrom, 1982).

SUMMARY

The late Quaternary history of Montezuma Creek and Recapture Creek generally parallels that of other streams on the Colorado Plateau. Downcutting was active during most of Quaternary time. The maximum average rate of downcutting indicated for upper Montezuma Creek during the last 700,000 years is about 1.6 feet (0.5 m)/1000 years. The average rate of downcutting in lower Recapture Creek during the last 150,000 years has been 0.8 feet (0.24 m)/1000 years. Downcutting during Pleistocene time was accompanied by formation of rock-cut terraces which appear to have been in large part controlled by lithology, and gravels commonly cap resistant sandstone beds in canyon walls. Deposition of finer-grained sediment began in upper Montezuma Creek before 5130 B.P. The canyon bottom was filled to a depth locally exceeding 50 feet (15 m) with bedded sand, silt, and clay (Unit 1). The uniform bedding, lack of channel cut and fill deposits, and thick organic clay beds in Unit 1 indicate a broad flood-plain environment generally absent today. A period of incision of these sediments occurred sometime before 1430 B.P., followed by alluviation of incised channels lasting until after 1410 B.P. This channel-fill material is very similar to that found in the modern channel. It consists of a basal sand and/or gravel overlain by multiple channel cut and fill deposits (Unit 2). Evidence for a hiatus in deposition with minor reincision during channel filling is present locally (Unit 2a/Unit 2b), but ultimately the channel was filled to or slightly below the previous level of Unit 1 and remained stable until about 100 years ago. During this period, eolian dune and sheet sands covered part of the flood plain and became locally interbedded with alluvial deposits. Sometime after 1900 A.D., Montezuma Creek and probably Recapture Creek cut down to their present level.

During the last 5000 years, essentially no additional bedrock incision has occurred. The canyons have continued to widen, with retreat of the rim of Montezuma Canyon proceeding at a maximum average rate since about 700,000 years ago of 5.7-7.5 feet (1.7-2.3 m)/1000 years. Eolian deposits are present atop the rim and on various benches within the canyons. These deposits vary in age from Early or Middle Pleistocene to the present. Active dunes are

TABLE 2. *Age and correlation of Quaternary alluvial deposits, Colorado Plateau*

Age		Montezuma Creek Southeastern Utah (this study)		Gravel Pit Wash, near Bluff Southeastern Utah (Oviatt, this volume)		Gibson Dome Area Southeastern Utah (Woodward-Clyde Consultants, 1982b)		Elk Ridge Area Southeastern Utah (Woodward-Clyde Consultants, 1982c; Salkin, 1975)		Spanish Valley Southeastern Utah (Harden and others, 1982; Woodward-Clyde Consultants, 1982a)		Chaco Canyon Northwestern New Mexico (Hall, 1977)		Jeddito Wash-Tsegi Canyon Northeastern Arizona (Karlstrom, 1982)	
Epoch	Years before present	Age	Unit (height above stream in m)	Age	Unit (height above stream in m)	Age	Unit (height above stream in m)	Age	Unit (height above stream in m)	Age	Unit	Age	Unit	Age	Unit
HOLOCENE			Unit 3	?	Deposition 4								Historic Post-Bonito	250; 500-600; 650-700	Naha (Naha/ Tsegi boundary)
				?	Deposition 3B	720±550(?)	Base of cut and fill, Cottonwood Creek								
	1000			650±70	Deposition 3A										
		1410±80; 1430±80	Unit 2	1350-1450	Deposition 2	1270±100; 1600±100	Continuous fill, Cottonwood Creek			1280±55	Gold Basin		Chaco		
	2000							1750±60	Continuous fill, Cottonwood Wash						
	3000							3070±325							Tsegi
	4000			? late to early Holocene ?									Gallo		
	5000	5130±200	Unit 1		Deposition 1	5160±55	Alluvium, Hart's Draw							5300	Closely predates soil on Jeddito Fm. (Tsegi/Jeddito boundary)
	6000													6500	Soil on Jeddito Fm
	7000														
	8000					7760±155	Alluvium, lower Hart's Draw	7385±80; 8100±345	Continuous fill, Kane Gulch					?	
	9000			? early Holocene to latest Pleistocene ?	San Juan River Sand (1.5)			9550±80	Alluvium, Mule Canyon						Jeddito (gravel)
	10,000						Indian Creek		Cottonwood Wash			?			
Not To Scale				Pinedale(?)	Gravel (20-25)	Pinedale	Gravel (10)		Gravel (9)	Pinedale	Beaver Basin	Late Pleistocene ?	Fajada (gravel)	Late Pleistocene ?	
PLEISTOCENE	LATE	?	Gravel (6)	?	Gravel		Gravel (8)								
			Gravel (12)	?			Gravel (12)								
	125,000	Late to Middle Pleistocene	Gravel (21)	Late to Middle Pleistocene	Gravels (130-150)	Bull Lake	Gravel (20)		Various gravels	Bull Lake	Placer Creek				
	MIDDLE		Gravel (35-37)				Gravel (32)								
		?	Gravel (49)	?			Various gravels (53)	?	Gravel (67)		Middle Harpole Mesa				
	700,000	Matuyama reversed epoch				?		Matuyama reversed epoch	No Man's Island gravels (164)	Matuyama reversed epoch	Older Harpole Mesa				
	EARLY		Abajo Mountains gravels						Abajo Mountains gravels						

found chiefly in the Bluff Bench area of lower Recapture Canyon.

Correlations with alluvial sequences in adjacent areas indicate that deposition in Montezuma Canyon was occurring at the same time as deposition elsewhere during the last 5000 years. However, bracketing dates for these depositional episodes in Montezuma Canyon are not sufficiently well-known to draw conclusions regarding their synchroneity with depositional episodes in other streams on the Colorado Plateau.

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LATE QUATERNARY GEOMORPHIC CHANGES ALONG THE SAN JUAN RIVER AND ITS TRIBUTARIES NEAR BLUFF, UTAH

By Charles G. Oviatt

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By CHARLES G. OVIATT [1]

ABSTRACT

Three small ephemeral washes graded to the San Juan River near Bluff, Utah, have produced fine-grained coalescing alluvial fans and have experienced at least four periods of Holocene deposition and erosion. Sediments of the first depositional period possess a moderately developed soil at their upper surface that predates Basketmaker III human occupation (ceramic- and radiocarbon-dated at A.D. 500 to 600). Sediments of the second period of deposition enclose Basketmaker III occupation sites in their upper part. A hearth near the middle of the third sedimentary sequence provided a radiocarbon date of 650 ± 70 yrs. B.P. (A.D. 1300). Horse bones in alluvium near the top of the fourth depositional sequence were most likely deposited after A.D. 1880, the date of settlement at Bluff. Eolian deposition has been continuous throughout the Holocene, and eolian deposits locally interfinger with alluvium.

Climate is undoubtedly an important extrinsic variable in both the high-order San Juan River and the low-order ephemeral washes, but the two systems probably do not respond in phase to climate changes. Conditions on the San Juan River dictate whether its ephemeral tributaries can deposit or erode. Both large and small stream systems, however, are controlled by extrinsic factors other than climate, particularly human activity. The San Juan River has eroded its banks at Bluff during catastrophic floods to a greater degree during the past 100 years of historic agriculture than it has in more than 1400 years prior to this. Prehistoric human populations may have exerted a similar influence, although probably of a lesser magnitude.

The intrinsic and extrinsic variables in the ephemeral stream systems at Bluff interact in complex ways through time resulting in a local stratigraphic sequence that has significance for environmental reconstructions and environmental planning. From a regional perspective, sediments of the first three depositional episodes at Bluff can be correlated with the Tsegi deposits, and sediments of the fourth episode with the Naha deposits of the generalized alluvial model for the southwestern U.S. These correlations require the assumption that the Tsegi and Naha are diachronous lithostratigraphic units.

INTRODUCTION

This study was undertaken in conjunction with archeological investigations along a re-alignment of U.S. Highway 163 near Bluff, Utah. The area discussed in this paper is in the San Juan River valley approximately 1.5 km west of the town of Bluff (fig. 1). Archeological excavations were conducted by the Antiquities Section of the Utah Division of State History. Five Basketmaker III (Anasazi) sites were excavated and dated by ceramic and radiocarbon methods to the period A.D. 500 to 600. All of the sites were in sand dunes overlying and intertonguing with alluvium of three ephemeral tributaries of the San Juan River. Archeological sites discussed in this paper are designated by their Smithsonian numbers (e.g., 42Sa8821). The complete archeological report for this project is published elsewhere (Neily, 1985).

Bedrock units in the study area are the Carmel Formation, Entrada Sandstone, Summerville Formation, and Bluff Sandstone (O'Sullivan, 1965). Surficial deposits consist of Pleistocene cobble and gravel alluvium, Holocene sandy alluvium, coarse-grained colluvium, and sandy eolian deposits (figs. 2 and 3). Altitudes in the study area range from about 1310 to 1460 m.

Three ephemeral washes, graded to the San Juan River during the Holocene, have formed the broad,

[1]Geologist, Mapping Section, Utah Geological and Mineral Survey.

sandy, alluvial bench (haA in fig. 2), which the highway traverses in the study area. The bench appears to be a terrace of the San Juan River when viewed from a distance but is actually a truncated alluvial-fan complex produced by the three washes. The bench is bounded on the north by the arcuate cut of an old San Juan River bend, and on the south by the northern edge of the modern San Juan River floodplain. The stratigraphy and environmental history of the sediments underlying this bench are the major foci of this paper.

The modern climate at Bluff is warm and dry with a mean annual temperature of 12.5°C and mean annual precipitation of 19.2 cm. Paleoclimate during the late Holocene in the Four Corners area is discussed by Andrews and others (1975), Euler and others (1979), Fritts and others (1965), Hall (1977), Martin (1963), Mehringer (1967), and Schoenwetter and Eddy (1964). Modern vegetation near Bluff is dominated by salt desert species including greasewood, four-wing saltbrush, snakeweed, shadscale, Indian ricegrass, and galleta grass (Oviatt, 1985).

PLEISTOCENE ALLUVIUM

The oldest Quaternary deposits in the Bluff area consist of isolated dissected mounds of cobbles, gravel, and sand perched on Bluff Bench, Tank Mesa, and Casa del Eco Mesa 130 to 150 m above the San Juan River. The cobbles consist predominantly of resistant igneous and metamorphic rocks having sources upstream along the San Juan River and its tributaries. Abundant quartzite cobbles, as well as

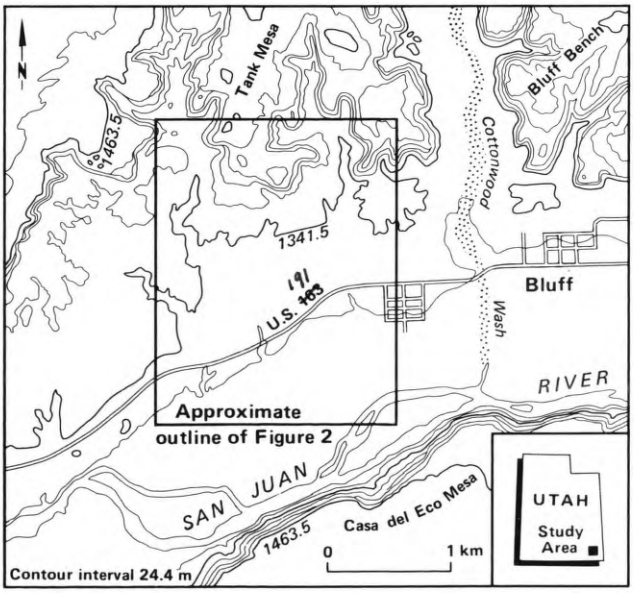


FIGURE 1. Location and topography of the study area.

FIGURE 2 EXPLANATION *(See opposite page)*

e1 Thick eolian sand deposits overlying Pleistocene alluvial deposits, bedrock knolls, and colluvial slopes. Includes sheet sands, and climbing and falling dunes.

e2 Thick eolian sand deposits overlying and interfingering with Holocene sandy alluvium. Includes parabolic dunes containing Basketmaker III sites.

af Fine-grained San Juan River floodplain deposits, and coarse to fine-grained modern channel deposits of ephemeral washes.

haC Late Holocene sandy to gravelly alluvial deposits of Depositions 3 and 4 of Gravel Pit Wash. Composed of locally derived weathered bedrock and reworked alluvial and eolian deposits. Eolian sand 0-2 m thick interfingers with or overlies the alluvial deposits in places.

haB **haA** Holocene sandy to silty or gravelly alluvial deposits, undifferentiated. haA is composed of locally derived weathered bedrock and reworked alluvial and eolian deposits. haA includes Depositions 1 and 2 of Gravel Pit Wash and similar deposits of the two adjacent washes. haB is derived from the Cottonwood Wash drainage. The stratigraphic relationship between haA and haB is unknown. Eolian sheet sands overlie the alluvial deposits in places.

pa2 **pa1** Pleistocene cobble, gravel, and sand deposits of the San Juan River. Clasts are predominantly resistant igneous and metamorphic rocks. pa1 occurs as dissected mounds of coarse alluvium perched on the bluffs above the valley. pa2 caps bedrock benches and forms a distinct terrace within the valley.

bca Bedrock, colluvium, and alluvium including sandstone and shale outcrops of the Jurassic Summerville, Entrada, and Carmel formations; colluvium composed of weathered bedrock, rockfall debris, and reworked Pleistocene alluvial deposits; thin Holocene alluvium and modern channel deposits along washes.

b Jurassic Bluff Sandstone bedrock.

A, B, C, D, E: locations referred to in the text.

Base from ASCS aerial photographs taken in 1971 and 1980.

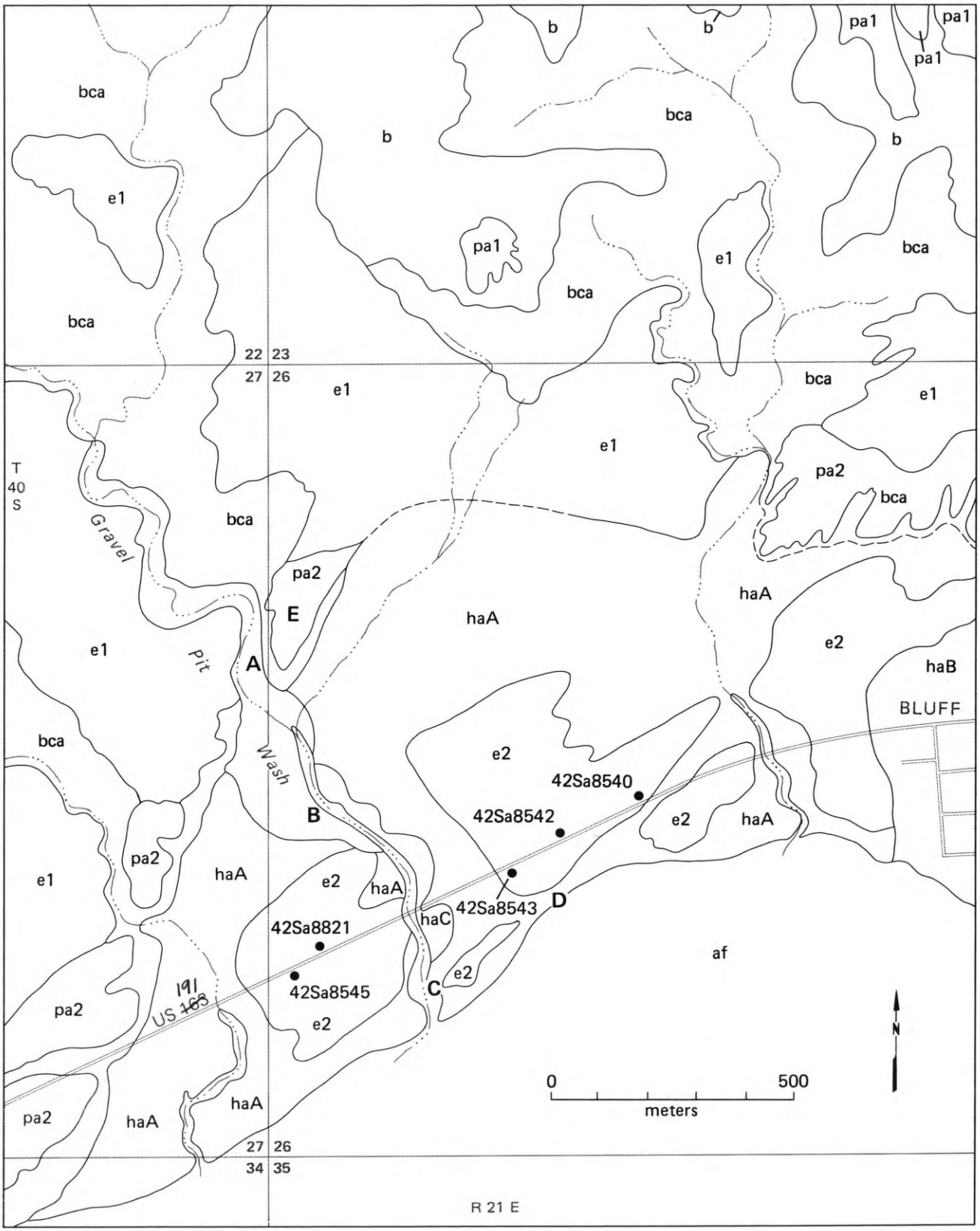


FIGURE 2. Surficial deposits in the study area (see Explanation on opposite page).

clasts of other metamorphic and igneous rock types, were transported by the San Juan River from the San Juan Mountains in Colorado (Cooley, 1960). Other igneous rocks, especially dioritic porphyries, were eroded from the intrusive Abajo Mountains (Cooley, 1960). Some clasts of fine-grained siliceous rocks eroded from sedimentary formations, particularly the Morrison Formation, are also present.

These high-level deposits no longer retain terrace morphology because the surrounding rocks and sediments have eroded more quickly. They represent deposition by the San Juan River before it had downcut through the Bluff Sandstone (pa1 in fig. 2). These deposits have not been dated, but they are probably middle to late Pleistocene in age if the San Juan River has downcut 250 to 450 m during all of Quaternary time (Cooley, 1962).

Younger Pleistocene deposits of similar texture and composition form a series of terraces within the river valley. There are at least two terraces, the lower of which is most prominent and is represented by pa2 in figure 2. The higher terrace(s) are obscured in the Bluff area by sand dunes, and were not mapped on figure 2. Cobbles, gravel and sand in the lowest terrace are well exposed in the abandoned gravel pit east of Gravel Pit Wash (location E, fig. 2). The gravel here is about 3 m thick and overlies red Entrada Sandstone.

A shallow pit dug on the undisturbed terrace surface north of the gravel pit disclosed a moderately developed soil. The profile has a thin (0.5 cm) A horizon, a 17-cm thick, reddish brown (5YR 4/4), pebbly B horizon, and a pale brown (10YR 6/3) Cca horizon. The degree of soil development in this profile is compatible with a late Pleistocene age for the deposit (see Richmond, 1962).

This lowest gravel terrace (pa2, fig. 2) is about 20 to 25 m above the San Juan River and therefore may be equivalent to the latest Pleistocene (Pinedale?) terrace(s), as defined by previous workers, upstream along the San Juan River (Antevs, 1939; Bandoian, 1968; Pastuszak, 1968).

No absolute dates have been obtained on the river terraces or on their suggested correlative glacial tills in the San Juan Mountains (Richmond, 1965). The assumption behind the glacial till/river terrace correlation is that coarse alluvium was deposited by the river during periods when streams had higher discharge rates and were overloaded with glacial outwash and periglacially derived sediments. This assump-

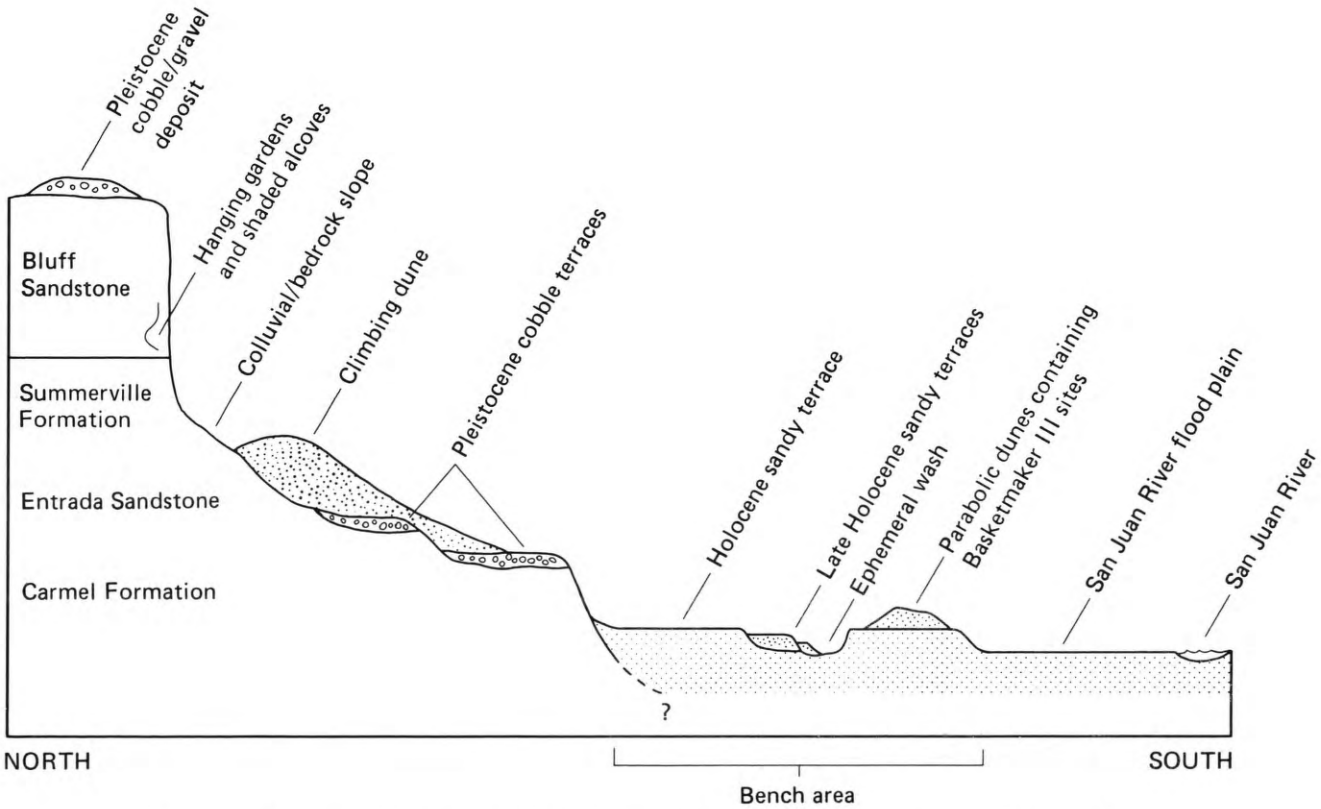


FIGURE 3. Schematic cross section showing general physiographic relationships in the Bluff area.

tion, although probably valid, has not been adequately tested by regional mapping and radiometric dating of alluvium and till in the San Juan drainage basin. Work by Gillam (1982) addresses this and other questions related to Animas River terraces in Colorado and New Mexico.

HOLOCENE ALLUVIUM

GENERAL CHARACTERISTICS

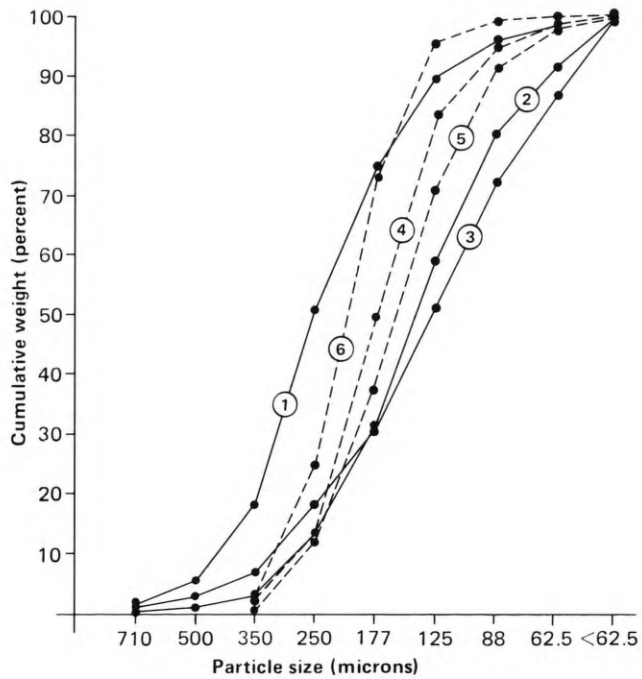
Holocene deposits in the study area consist of alluvium, colluvium, and eolian sand (fig. 2). The unstratified colluvial deposits were not studied in detail. The eolian and alluvial deposits are similar in general appearance, because they were derived from the same local sources. A careful examination of bedding, as well as texture (fig. 4) and composition (fig. 5), is therefore required to confirm the depositional environments of stratigraphic units that lack geomorphic expression or other evidence of their origin.

All the bedrock units in the area, especially the eolian Bluff Sandstone, contain a high proportion of rounded and frosted quartz sand grains. These grains are abundant in both the Holocene alluvial and eolian deposits (fig. 5). The alluvium, however, contains some iron-oxide and calcite-cemented siltstone and sandstone fragments which are not found in eolian sand samples (fig. 5). In the eolian environment the rock fragments disintegrate quickly from intense abrasion by the more resistant quartz grains.

All of the Holocene alluvial deposits are primarily sandy in texture, but they also contain coarse sand and gravel channel-fill deposits, and silty or clayey layers. Some rounded pebbles and cobbles reworked from the Pleistocene deposits are present, but most of the sediment is derived from weathered bedrock, reworked sand dunes, colluvium, and older sandy alluvium.

Sand deposited by the San Juan River is exposed in the lower parts of cuts in several locations and is light yellowish brown (2.5Y 6/4) to pale brown (10YR 6/3). Sand deposited by the local ephemeral washes is commonly reddish brown (5YR 5/4), but can range to light brown (7.5YR 6/4). The ephemeral wash sediments, therefore, appear markedly red as compared to the San Juan River sands. The San Juan River sands, moreover, contain relatively few rounded and frosted grains (fig. 5).

After downcutting below pa2 (fig. 2) at the end of the Pleistocene, the San Juan River deposited sediments in at least part of the bench area as high as 1.5 m above the surface of the modern floodplain. The



1. Gravel Pit Wash alluvium—historic terrace sand
2. Gravel Pit Wash alluvium—Deposition 2 sand
3. Gravel Pit Wash alluvium—Deposition 1 sand
4. Longitudinal dune crest
5. Parabolic dune crest
6. Climbing dune crest

FIGURE 4. Cumulative grain-size curves for various sediments from the Bluff area. Seive analysis following Folk (1974).

river subsequently shifted its channel to the southern part of the valley and has remained there for probably most of the Holocene. Virtually all the exposed sediments in the bench area (haA on fig. 2), therefore, were deposited by the three ephemeral washes that enter the area from the north.

Gravel Pit Wash, the largest of the three washes, drains an area of 5.0 km^2. Its drainage basin is bounded on the east and west by smaller basins of 1.7 km^2, and 1.5 km^2, respectively. Each of these washes drains a portion of the southern edge of Tank Mesa in addition to a small area below the cliffs. Sixty-five percent of the total drainage area of Gravel Pit Wash is on Tank Mesa, compared to 43 percent of the eastern basin and 24 percent of the western basin. These second or third order streams have deposited approximately 5 x 10^6 m^3 of sediment in overlapping and coalescing low-gradient alluvial fans.

GRAVEL PIT WASH TERRACES AND SEDIMENTS

Four terraces along Gravel Pit Wash (fig. 6) pro-

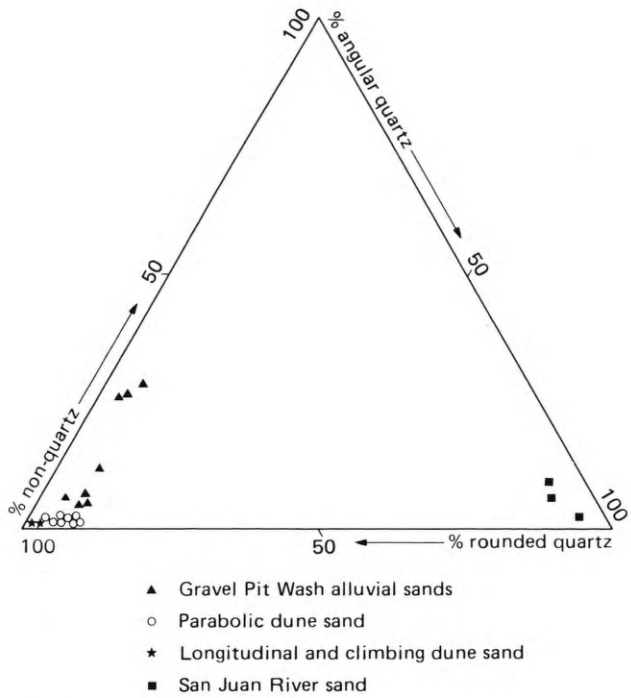


FIGURE 5. Triangular diagram showing the composition of various sediment samples from the Bluff area. Each point represents a separate sample (200 sand grains 177-250 micrometers counted). Non-quartz includes rock fragments and heavy mineral grains.

vide the geomorphic framework for a sequence of numbered depositional and erosional events. Terrace 1, for example, is underlain by alluvium of Deposition 1, and became a terrace with the initiation of Erosion 1. Some deposits have been dated by associated archeological materials or by radiocarbon, but other deposits and erosional events have been placed in the sequence in terms of their relative ages.

Terrace 1 extends laterally from Gravel Pit Wash as a broad, dissected surface covering more than 50 percent of the bench area (figures 2 and 6). Deposition 1 sediments, although not well exposed, are generally sandy with interbedded gravel lenses and clayey layers.

The age of Deposition 1 is unknown but it is bracketed by geomorphic events whose ages are grossly inferred. The sediments are younger than late Pleistocene or early Holocene downcutting by the San Juan River, and are older than a relatively long period of geomorphic stability and soil formation on the Terrace 1 surface. Eolian sand, which began accumulating shortly before the A.D. 500 to 600 Basketmaker III inhabitation of the area, overlies a soil at the surface of Terrace 1 (figs. 6 and 7). Thus Deposition 1 spans most of the early and middle Holocene. Although exposures are inadequate for proper testing, Deposition 1 probably consisted of multiple depositional and erosional events.

The soil at the surface of Terrace 1 was exposed in a 70 m long bulldozer cut at 42Sa8821 during highway construction. The well-developed argillic and oxidized B horizon of the buried soil (fig. 7) was discontinuous; some parts of the buried terrace surface were merely oxidized and lacked argillic horizon development. Discontinuous argillic horizons were also found buried in eolian sand and at the modern surface of dunes and terraces. The argillic horizons, or clay pans, are apparently related to greasewood bushes (*Sarcobatus vermiculatus*) which induce clay translocation by adding sodium cations to the soil from their decomposing foliage (Oviatt, 1985).

Greasewood-induced clay pans cover approximately 50 percent of the exposed surface of Terrace 1 north of 42Sa8545 and 42Sa8821 along a dirt road that leads from U.S. Highway 163 to the abandoned gravel pit (location E, fig. 2). Clay pans cover less than two percent of the surface of a younger terrace (3A or 3B) along the same road, and therefore provide a relative measure of the age of the terraces.

Terrace 2 is similar in height to Terrace 1 (fig. 6), although any existing topographic break between the two terraces is obscured by eolian sand. Deposition 2 sediments are generally finer grained than Deposition 1 sediments.

In cutbacks along Gravel Pit Wash, Deposition 2 sediments appear banded due to the sharp boundaries between sedimentation units. A sedimentation unit, as characterized here, is an upward-fining aggradational sequence that is slightly oxidized in the upper 10 cm due to a short period of subaerial exposure following deposition. A single exposure five m high may have six sedimentation units, averaging 80 cm in thickness, extending continuously along the length of the cut. The time represented by an individual sedimentation unit has not been determined. Some sedimentation units appear to be related to single floods, but others suggest compound depositional episodes.

Eolian sand within which a Basketmaker III occupation site is exposed is intercalated with Deposition 2 sediments near the mouth of Gravel Pit Wash (fig. 8; location C, fig. 2). At location D scattered Basketmaker III pottery fragments occur in alluvium about one m down from the bench surface. These occurrences suggest that Deposition 2 ended shortly after Basketmaker III occupation, A.D. 500 to 600 in this area.

During Basketmaker III occupation the ephemeral-

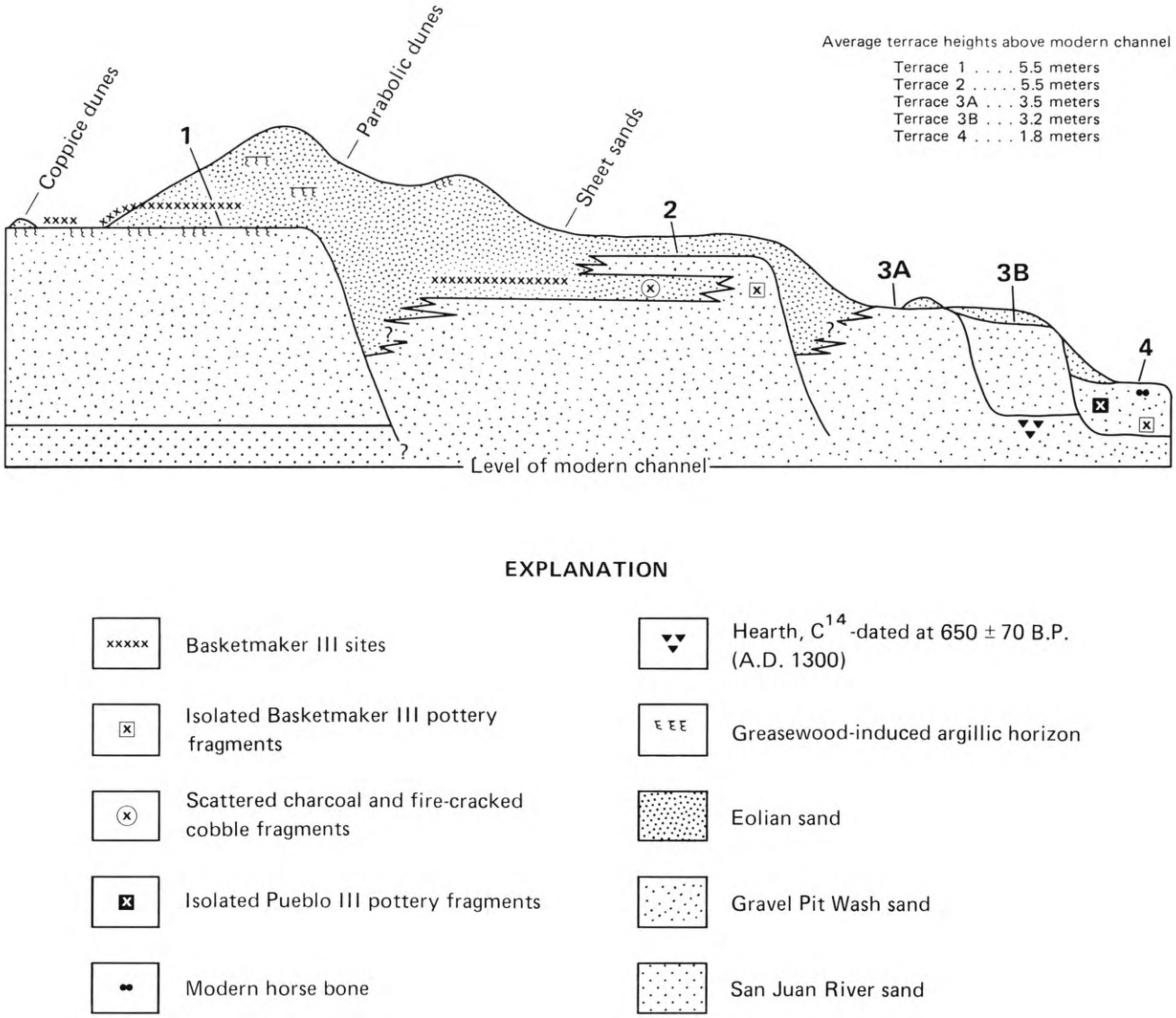


FIGURE 6. Composite schematic diagram of geomorphic and stratigraphic relationships along Gravel Pit Wash and nearby areas.

wash fan surfaces had a continuous convex-upward gradient down to the San Juan River floodplain. The landscape would have been a striking contrast to the present-day landscape in which a five-meter vertical scarp eroded by the San Juan River separates the floodplain from the bench surface. The fans were probably ideal places for flood-water farming in Basketmaker III times.

Fine-grained alluvium that was being supplied to the fans during Deposition 2 was reworked by the wind into sand dunes and sheet sands. Dunes were formed adjacent to distributary channels on the active fan surfaces and on older fan surfaces or terraces. At the backs of the aggrading fans, where they intersected the Deposition 1 sediments, eolian sand accumulated and interfingered with the alluvium (fig. 6). At 42Sa8542 (fig. 2) exploratory pits 2.5 to 3 m deep in and adjacent to the modern interdune hollow did not reach alluvial deposits. Based on plane table surveys, the surface of Terrace 2 should have been encountered beneath eolian sand at a shallow depth. Similar relationships exist at 42Sa8540 and 42Sa8543.

Deposition 3 sediments are very similar in texture, composition, and internal structure to Deposition 2 sediments. Except where datable materials are present, the two sedimentary sequences can be separated from each other only by the height of the terrace surfaces above the modern channel of Gravel Pit Wash (fig. 6). In eroded areas, however, or where

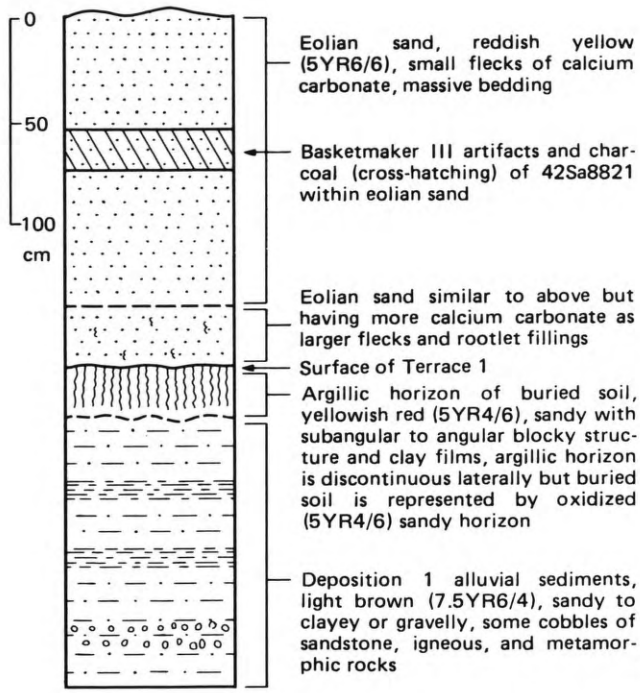


FIGURE 7. Stratigraphic relationships at 42Sa8821.

the eolian sand cover is thick, even the terrace height is not a useful distinguishing characteristic.

Deposition 3 consists of two aggradational sequences separated by a relatively minor unconformity (Erosion 3A). The difference in height between terraces 3A and 3B is not significant (fig. 6) and, at the one location where their sediments can be seen in stratigraphic section (location B, fig. 2), the two terraces cannot be distinguished on the surface. Because of this, and because the volume of Deposition 3B sediments is small relative to the 3A sediments, they are treated together as representing two depositional episodes within a single major aggradational period.

Charcoal from a hearth in Deposition 3A sediments yielded a radiocarbon date of 650 ± 70 yr B.P. (A.D. 1300; Beta-1945). The hearth was exposed in a cutbank of Gravel Pit Wash (location B, fig. 2) and was directly below the unconformity separating 3A from 3B sediments (fig. 6). There is no evidence, however, that the charcoal was emplaced after, or was disturbed by, the Erosion 3A event that produced the unconformity. The date is interpreted to be from the middle of the Deposition 3 (3A plus 3B) major aggradational period.

Geomorphic conditions during Deposition 3 were probably similar to those during Deposition 2. Fine-grained alluvial fan deposits probably interfingered with eolian deposits as they did during the Basketmaker III occupation. The total volume of sediments was less, however, and the fluvial and eolian landforms were somewhat constrained by the walls of the arroyos that formed during Erosion 2. Nevertheless, conditions were probably nearly as favorable for horticulture as they had been during Basketmaker III occupation.

Terrace 4 is confined within the narrow, constricted walls of the Erosion 3B arroyo. The modern channel of Gravel Pit Wash parallels this ancient arroyo. Sediments of Deposition 4 are generally coarser grained than the older alluvial deposits. Graded sedimentation units, which are prominent in Depositions 2 and 3, are not well-formed in Deposition 4. In the 1 to 1.5 m of sediments, channel-fill structures are common and horizontal silty or clayey layers are rare.

Deposition 4 sediments contain abraded Basketmaker III and Pueblo III pottery fragments, and the bones of a modern horse. The bones are near the surface of Terrace 4 and are overlain by 20 cm of sandy alluvium having undisturbed bedding (location A, fig. 2). Although white explorers passed through the San Juan County region in the mid 1800's (Gregory, 1938), the white population of the area increased markedly when the Mormon settlers arrived in Bluff in 1879 and 1880 (Perkins et al., 1968). Therefore, based on the probability of horses being in the area, the bones in Deposition 4 were most likely deposited after 1880. Aerial photographs taken in 1937 show the channel of Gravel Pit Wash entrenched approximately 1 meter below Terrace 4. Thus, the end of Deposition 4 probably came after 1880, but before 1937.

EOLIAN DEPOSITS

Sand dunes in the Bluff area are similar in many respects to the sand dunes studied by Hack (1941) in northeastern Arizona. In fact, the chain of sand dunes between Black Mesa, where Hack worked, and Bluff is broken only by minor gaps (Hack, 1941, p. 244; Cooley et al., 1969, fig. 14).

Longitudinal dunes, whose crests parallel the wind direction, on Bluff Bench, Tank Mesa, and Casa del Eco Mesa are commonly three or more kilometers long and three or more meters high. The dunes have been formed by southwesterly winds, although there are slight variations in orientation from mesa to mesa (table 1). The longitudinal dunes generally have a moderate vegetation cover, but local blowouts and portions of the dune crests are free of vegetation. In road cuts on Bluff Bench the longitudinal dunes have a relatively well-developed calcic horizon at depth. This suggests that their initial deposition was considerably earlier than the initiation of the parabolic

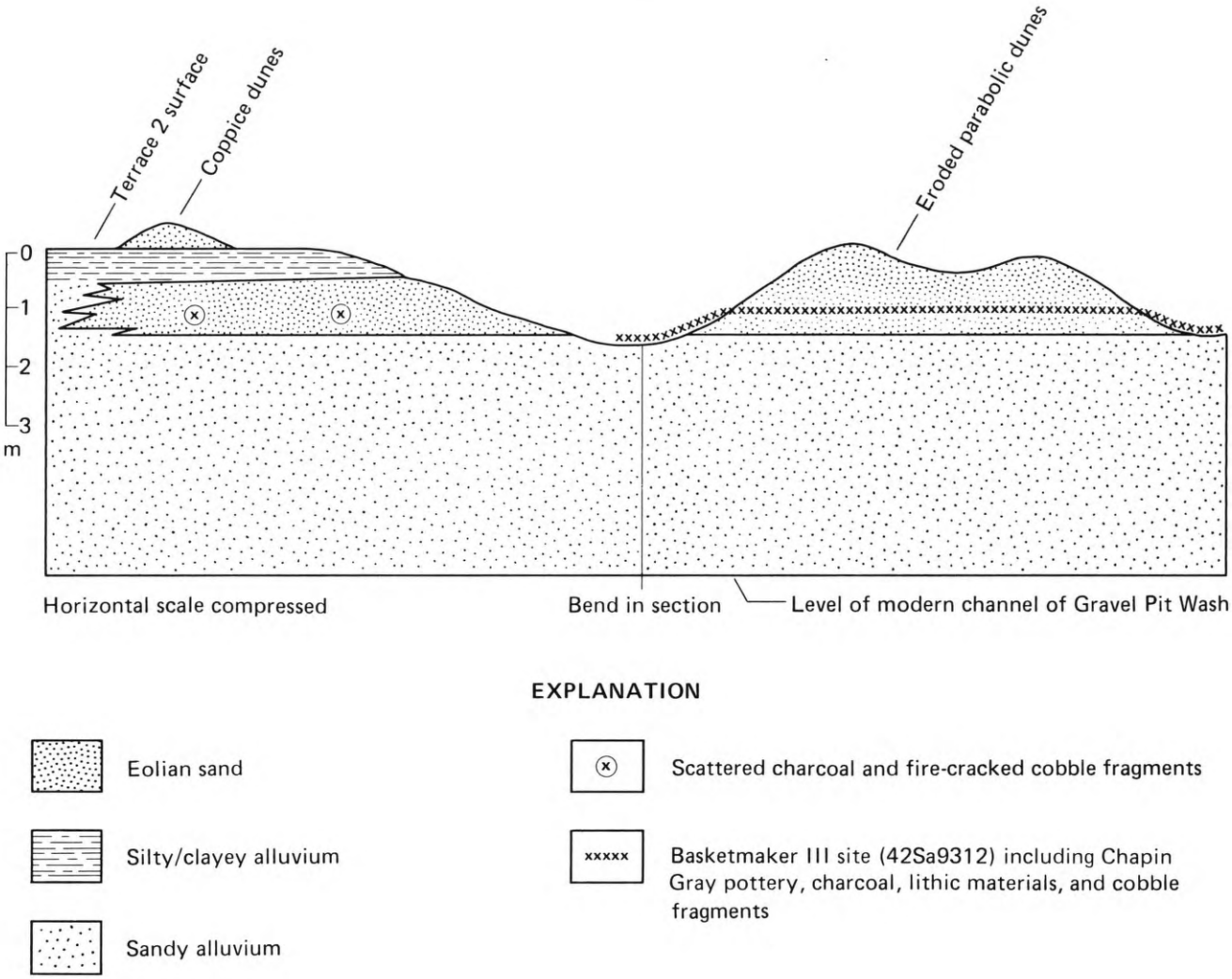


FIGURE 8. Stratigraphic relationships in a natural cut on the east (left) bank of Gravel Pit Wash near its juncture with the San Juan River floodplain (location C in fig. 2).

dunes in the bench area, which lack a well developed calcic horizon. The longitudinal dunes may be relict land forms produced during the glacial episodes of the Pleistocene (Hack, 1941).

Within the river valley, climbing and falling dunes are common on remnants of Pleistocene cobble terraces and on colluvial or bedrock slopes (e1 in fig. 2). Sand is deflated from the nearby washes and from exposed badland areas and is deposited on the tops of small hills. The sand cascades down the leeward (northeast) sides of the hills as falling dunes and is removed at the base of the slopes by flood waters in the ephemeral washes. In some places the sand is trapped along the high cliffs and builds up as massive climbing dunes covering the colluvial slopes. Where this happens sand is removed from the dune system by running water concentrated at the base of the cliffs. The climbing dunes have probably been continuously active since the Pleistocene, maintaining a relatively constant mass balance between sand inputs and outputs.

The bench area is dominated by parabolic dunes and sheet sands. The parabolic dunes closely resemble the "parabolic dunes of accumulation" of Hack (1941), and the "source-bordering lee dunes" of Melton (1940). Parabolic dune areas (e2 in fig. 2) are located downwind from the three ephemeral washes that supplied the sand for dune formation.

The parabolic dunes grade laterally into sheet sands that cover large areas of the Holocene and Pleistocene terraces. The sheet sands are generally less than one meter in thickness, but sheet sands that

TABLE 1. *Average sand dune orientations in the Bluff area.*

Longitudinal dunes - Casa del Eco Mesa	N. 42°E.
Longitudinal dunes - Tank Mesa	N. 56°E.
Longitudinal dunes - Bluff Bench	N. 49°E.
Parabolic dunes - bench area	N. 51°E.

grade into climbing dunes farther up on the slope partially obscure the 13-m high San Juan River meander-cut east of the abandoned gravel pit (fig. 2). The surface of the sheet sands appears streaked parallel to the dominant wind direction.

The parabolic dunes and sheet sands are young landforms, having developed in the last two millennia. Parabolic dune deposition was initiated sometime before Basketmaker III occupation, but probably after the washes had transported a considerable amount of sand onto the fans. There are no extensive buried soils or unconformities within the eolian deposits, suggesting that eolian deposition has been continuous at least since Deposition 2, regardless of whether erosion or deposition was occurring in the stream channels.

HISTORIC GEOMORPHIC CHANGES

Historic changes in the San Juan River valley are documented by the written and oral accounts of early settlers and by the reports of geologists and hydrologists.

> When the Hole-in-the-Rock expedition reached the site of Bluff in 1880,the soil was apparently quite good, but as the pioneers converted the brushland and reed swamps along the river into agricultural land, the San Juan, which had for some time flowed through the center of the valley, began to cut into the banks, changing its channel and flooding the farm land. South of Bluff it took at least 150 acres of good soil, leaving only sand and cottonwood trees (from an interview with Norman Nielson; Jensen, 1966, p. 23).

When the Mormon settlers arrived "the soil was rich, covered with cottonwood, and about six feet above the swift current of the river" (Perkins et al., 1968, p. 60).

> The river, at the time of the arrival of the settlers, ran almost due west through the center of the valley, making land available on both sides of the stream. The land south of the river was not used and has since been almost all flooded away (Jensen, 1966, p. 67).

> Within the last 30 years [1894-1924] the channel has widened . . . and has assumed a braided appearance (Miser, 1924, p. 67).

According to Albert R. Lyman (Jensen, 1966, p. 24) early surveys showed 960 acres of arable land on the north side of the river. By 1886, however, there were only 700 acres at Bluff (Gregory 1938, p. 33), and by the 1930s "the destruction of arable land, the result of changes in the river's current and the neglect of irrigation ditches, [had] reached a stage where barely 200 acres [remained] . . ." (Gregory 1938, p. 9).

Floods having exceptionally high discharges (equal to or greater than 50,000 ft^3s^{-1}) occurred in 1884 (Gregory 1938, p. 33; Jensen 1966, p. 30), 1911 (LaRue 1916, p. 213; Miser 1924, p. 56), 1927, 1941, and 1970 (fig. 9). Between 1911 and 1925 the San Juan River washed away from the bench area—south of 42Sa8540—about 425 acres of arable land that had been irrigated by an artesian well (Bryan and LaRue 1927, p. 256-257). The well casing, which had protruded less than a meter above the bench surface, was left standing 5 to 7 m above the broad, braided river channel (Bryan and LaRue 1927, fig. 8, p. 256).

Aerial photographs taken in 1937, 1950, 1955, 1961, 1971, and 1980 by the U.S. Soil Conservation Service show random changes in the configuration of the braided channel of the San Juan River. This rearrangement of sand bars and active channels probably occurs with every large discharge event.

There has been no significant erosion of the bench sediments by the San Juan River during the last 43 years. Even the large floods of October 14, 1941, and September 6, 1970 (fig. 9), had no noticeable effect on the bench scarp. The 1970 flood, however, destroyed a levee and damaged fields and irrigation facilities near Bluff (Roeske et al., 1978, p. 35). Only an extremely large flood (possibly greater than the September 10, 1927 flood; fig. 9) would significantly erode the bench scarp in its present position.

The channel of Gravel Pit Wash has deepened and straightened in the past 43 years as shown by the aerial photographs. A prominent entrenched meander directly south of the gravel pit was occupied by a single main channel on the 1937-1961 photographs. By 1971 the stream had cut off the meander in two places and had entrenched itself slightly, and by 1980 stream flow was concentrated in a single channel at the meander cutoff. During the 43 years of record the channel has entrenched approximately 0.8 m.

In 1947 a meander bend was artificially cut off and a berm was constructed adjacent to the straightened

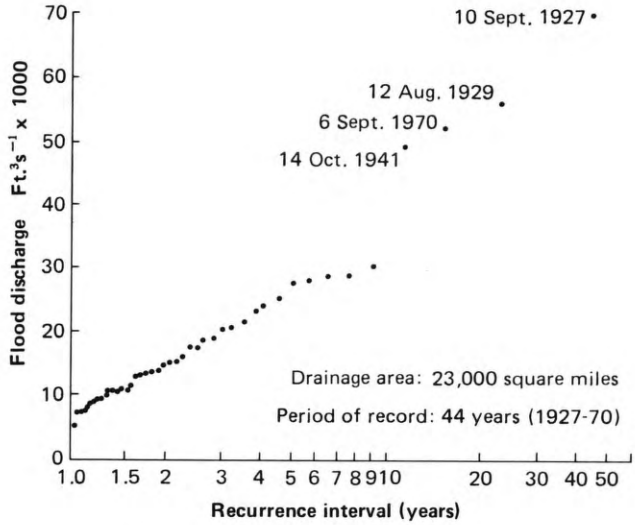


FIGURE 9. Recurrence interval of annual floods of specific magnitude, San Juan River near Bluff, Utah. Constructed following the method of Leopold (1974) using data from U.S.G.S. (1970), U.S.G.S. (1973), and Patterson and Somers (1966).

channel to control water flow into the culvert under Utah Highway 47 (later designated U.S. Highway 163). This highway was paved in 1957, and the channel was further straightened by an additional berm and meander cut-off.

A powerline was installed parallel to the highway, and the gravel pit (location E, fig. 2) was first developed between 1955 and 1961. The natural vegetation was removed and sand was redistributed on the parabolic dunes and Terrace 1 surface for an irrigated wheat field in the late 1970s. Additionally, several dirt roads were established in the drainage basin after 1961. All of these developments disrupted the natural runoff and sediment yield characteristics of the drainage basin and undoubtedly contributed to the gradient steepening and entrenchment of the Gravel Pit Wash channel. With more and more human activity of this kind in the area, it seems unlikely that the ephemeral streams could return to an aggradational regimen, even if the other variables involved (i.e., climate, San Juan River flooding, etc.) were favorable.

Sand is constantly redistributed by the wind in the Bluff area. In the early settlement "despite all precautions and housekeeping, sand sifted—even drifted—into the log cabins and tents, into the food and bedding. Sand storms were the order of the day in Bluff" (Perkins et al., 1968, p. 66). Some arroyo walls along Gravel Pit Wash, which are sharply defined on the 1937 and 1950 photos, are obscured by eolian sand in the 1971 and 1980 photos. A metal-post barb wire fence, constructed sometime between 1971 and 1980 northeast of 42Sa8540, is presently almost half buried in places with windblown sand. Sand is blown off the relatively bare ground of a neglected wheat field onto the dune complex where the fence is located.

In conclusion, it appears that humans have influenced the natural processes along the San Juan River and its tributaries during historic time. Prehistoric San Juan River floods were certainly of similar magnitudes and probably occasionally even greater than recorded historic floods (fig. 9), but they probably were less destructive. The disruption of the natural vegetation and floodplain ecology by the agricultural practices of the early settlers apparently significantly increased the erosive potential of the San Juan River. Prehistoric human populations may have had similar effects on the streams and the river at Bluff, but the destructive results were not as extreme. The San Juan River has eroded more of the bench during historic floods than it has in more than 1400 years. This is shown by Basketmaker III sites in Deposition 2 sediments that are now exposed in the bench scarp. Erosion of the banks of the floodplain induced headward erosion by the ephemeral washes, and the washes have subsequently been encouraged to maintain an entrenched condition by human intervention within their channels and drainage basins.

DISCUSSION

Alluvial stratigraphy and cycles of arroyo cutting and filling in valleys of the southwestern U.S. have been studied for many years. Attempts are often made to correlate a stratigraphic sequence from different areas with an alluvial stratigraphic model that is assumed to be applicable over the entire southwest and into neighboring regions (Miller 1958). The pioneering research of Bryan (1940) and Hack (1942) formed the basis of the regional model, but later work by Leopold (i.e., Leopold and Snyder 1951; Leopold and Miller 1954), Antevs (1955), Cooley (1962), Karlstrom (1982) and others has expanded and refined the chronology.

Haynes's (1968) data cast some doubt on the validity of inter-regional correlations, but he maintained the concept of climatically controlled order underlying alluvial stratigraphy. Cooke and Reeves (1976) have more recently attempted an integrated systems approach to the problem of arroyo cutting and filling. Euler et al. (1979) have presented an elaborate model, closely tied to climate, that implies long-range alluvial correlations across ecologi-

cally diverse regions.

The fluvial system is a complex natural system that is influenced by a number of intrinsic and extrinsic variables (Schumm, 1977). In valleys studied by Butzer (1980, p. 140-141), alluvial cycles "...reflect complex ecological readjustments that involve channel and floodplain geometry, rainfall seasonality, intensity, and periodicity, as well as runoff patterns, ground cover, and sediment calibre and amount. The critical, immediate variables are ground cover, runoff, and sediment supply. But the ultimate variables are climate and human activity." Love (1979) reports three scales of fluvial adjustment in Chaco Canyon, New Mexico, in response to changes in biologic, sedimentologic, morphologic, hydrologic, land use, and climatic variables.

Schumm and Parker (1973) and Womack and Schumm (1977) have demonstrated, both in a laboratory setting and in the field, that complex response of fluvial systems is effective in producing significant sedimentologic and geomorphic features. Studies by Patton and Schumm (1981) show that an ephemeral stream may undergo simultaneous erosion and deposition in different reaches independent of changes in extrinsic variables such as climate, vegetation, or land use.

Schumm and his colleagues feel that because of these, and other, complexities in the fluvial system, regional correlations between alluvial units may not be valid (Patton and Schumm, 1981). Butzer (1980), Knox (1983) and McDowell (1983) express similar viewpoints. If alluvial stratigraphic units are regarded as diachronous rock units, however, broad correlations that reflect pronounced or prolonged climatic episodes should be valid, at least within relatively homogeneous regions. Climate may, in fact, be the most important extrinsic variable in many drainages, but a direct one-to-one relationship between climate changes and alluvial cycles is unlikely (Butzer, 1980, p. 140).

By comparing well-dated stratigraphic sequences with independent climatic data Hereford (1984) and McDowell (1983) have shown that the streams they studied responded to changes in both climate and in intrinsic geomorphic controls.

Ephemeral streams in the Bluff area are affected by the same intrinsic and extrinsic variables that are important in most stream systems (Schumm, 1977; Love, 1979, p. 302-303). Variables that are especially important to the Bluff-area streams are discussed below.

Three kinds of San Juan River changes can have an effect on the base level of ephemeral washes graded to the river channel or floodplain. These are:

1. erosion of the toes of the alluvial fans by gradual lateral shifts in the position of the channel(s);
2. gradual vertical changes in channel or floodplain level (e.g., downcutting or aggradation); and
3. catastrophic modification of the floodplain or floodplain margin by floods.

For fine-grained alluvial fans to be built by Gravel Pit Wash and its neighboring washes, therefore, the San Juan River has to maintain a stable or slightly rising base level. During periods of fan deposition the river channel(s) are not swinging laterally to truncate the fan toes; the channel and floodplain remain stable or are aggrading; and the floods are of low magnitude, or are not destroying the banks of the floodplain. These conditions are controlled or influenced by such variables as climate, vegetation, human activity, local gradient changes, local changes in sediment load characteristics, and many other factors. The San Juan River is a complex system in itself and has a direct influence on its small tributaries.

The ephemeral washes are also complex systems having their own set of controls. Among the constraints intrinsic to the ephemeral washes is a geomorphic equilibrium threshold (Bull, 1979; Schumm, 1977) that prevents them from building fans above a certain level. The threshold is largely controlled by the stream gradient on the fan surface as discussed below.

As a fan aggrades, the stream gradients of the headward and central reaches of the fan gradually decrease. In contrast, the gradient on the distal reach of the fan steepens. This contrast is due to increases in self-enhancing feedbacks, such as flow width, infiltration capacity, and vegetation, which encourage deposition in the headward and central reaches (Bull, 1979). Stream velocity, channel width, sediment size, and sediment load in the distal reach are all affected by deposition higher on the fan. At some critical point (the geomorphic threshold) the gradient in the distal reach becomes oversteepened and the stream begins to downcut. The limit of this downward erosion is determined by base level.

If the sediment yield from the headward parts of the ephemeral drainage remains high, the stream will eventually begin to reaggrade and to produce a new threshold-controlled fan. The new fan will be built farther out on the flat-floored San Juan River valley and will have its apex at the mouth of the previous entrenched channel. In this way alluvial fans could be cyclically built and entrenched, and fluvial sediments redistributed on the valley floor. A similar mechan-

TABLE 2. *Summary of major Holocene geomorphic events along Gravel Pit Wash and adjacent areas.*

13. **Erosion 4** (historic) to level of modern channel; several minor depositional events below the level of Terrace 4; erosion of bench sediments by gullying, and extensive lateral erosion by San Juan River.
12. **Deposition 4** to 1.8 m above modern channel; modern horse bones in upper part; settlement at Bluff beginning A.D. 1880.
11. **Erosion 3B** to about one m above modern channel.
10. **Deposition 3B** aggradation to same level as Terrace 3a; continued eolian sand deposition.
9. **Erosion 3A** to about 1.3 m above modern channel.
8. **Deposition 3A** to about 3.5 m above level of modern channel; dated hearth (A.D. 1300 ± 70) from middle third of aggradational sequence; continued eolian sand deposition.
7. **Erosion 2** to below level of modern channel.
6. **Deposition 2** aggradation of fine-grained coalescing alluvial fans up to 5.5 m above level of modern channel; parabolic sand dune accumulation; Basketmaker III occupation in latter part of period.
5. **Erosion 1** along Gravel Pit Wash to at least the level of the modern channel; soil development on the surface of Terrace 1 predates Basketmaker III occupation.
4. **Deposition 1** in the bench area by ephemeral washes; early (?) to late (?) Holocene.
3. Lateral shifting of the San Juan River to the southern part of the valley; early Holocene (?).
2. Deposition by the San Juan River in the bench area as high as 1.5 m above the surface of the modern floodplain; early Holocene (?) or latest Pleistocene(?).
1. Downcutting by the San Juan River following the last major Pleistocene glaciation in the San Juan Mountains.

Depositions and Erosions are capitalized in the text for clarity but are not intended as formal stratigraphic units; numbers are arbitrary and do not imply stratigraphic correlations with any other published studies.

ism for the cyclic transport of sediments in higher-order ephemeral streams has been proposed by Schumm and Hadley (1957).

Because of landform singularity, however, the geomorphic threshold is gradational (Begin and Schumm, 1984). Therefore, cyclic aggradation and entrenchment along a single stream is unlikely to be spatially or temporally uniform, and alluvial cycles may not be synchronous, even in contiguous drainage basins.

The example above presents the ephemeral washes as independent of any extrinsic controls. In nature, however, this is not the case. The same geomorphic threshold is important in the stream system, but the threshold itself is influenced by changes in extrinsic variables (Bull, 1979).

Extrinsic variables affecting both the San Juan River and the ephemeral washes include climate, vegetation, and human activity. The washes and the river differ profoundly in order of magnitude, however, and it is unlikely that a change in some common factor, such as climate, would induce an identical response in each stream system. The interaction between the ephemeral washes and the San Juan River is, therefore, complex, and must result in dynamic geomorphic patterns. Each depositional and erosional event along Gravel Pit Wash could have been related to a somewhat different set of environmental conditions, including human acivities. A simple chronology of fluvial cutting and filling based on a single variable would, therefore, be difficult to construct.

CONCLUSION

The Bluff alluvial sequence is summarized in table 2, and in figures 2 and 6. The sequence can be viewed from two different stratigraphic perspectives, each of which has merit. On a regional scale, and in a rock-stratigraphic sense, sediments of Depositions 1, 2, and possibly 3 could be correlated with the Tsegi deposits, and sediments of Deposition 4 with the Naha deposits of Hack (1942; Karlstrom, 1982). This broad correlation is based on lithologic, morphologic, and chronologic characters, and assumes that the Tsegi and Naha are diachronous rock units. The magnitude and frequency of fluvial geomorphic processes at Bluff, however, cannot be extracted from a generalized regional model, and it is the local variation that is most critical in environmental reconstructions and environmental planning. Therefore, the details of the local stratigraphic sequence may provide information important in agricultural, urban, or industrial development or in archeological studies, but carefully documented regional patterns of erosion and deposition may help clarify the relationship between climate and fluvial processes.

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MASS MOVEMENT IN THE LA SAL MOUNTAINS, UTAH

By John F. Shroder and Robert E. Sewell

MASS MOVEMENT IN THE LA SAL MOUNTAINS, UTAH

By JOHN F. SHRODER [1] and ROBERT E. SEWELL [2]

ABSTRACT

The La Sal Mountains of central eastern Utah are largely three stock and laccolith complexes that intruded during Tertiary time into surrounding sedimentary rock of Mesozoic age. Much of the outward-dipping uppermost host rock is Morrison Formation and Mancos Shale, which are two of the most unstable, landslip-prone rock units of the Colorado Plateau. High altitude and this unstable substrate have combined to produce plentiful interrelated glacial and mass-movement landforms and diamicton deposits that have been described by diverse previous workers. Unsorted, unstratified, coarse clastic diamictons are assigned genetic attributes best, however, only where substantiating data are available. Sedimentologic studies have proved equivocal in other similar studies, but landforms are more reliable. Our reexamination of extensive diamictons in the La Sal area shows only nine small glaciated areas, in contrast to a previously mapped extent nearly 20 times larger. Evidence against previously supposed widespread glaciation is the presence of "V"-shaped valleys, lack of cirques, lack of clasts with striations or polish, common natural fracturing of igneous rock which mimics glacial faceting and soling, and supposed ice-wastage areas disproportionately large compared with possible ice-accumulation zones. The South Mountain Group has not been glaciated, and in the Middle Mountain Group only Dark Canyon, Gold Basin, and Horse Creek were glaciated.The North Mountain Group glacial valleys include Bachelor Basin, Bear Creek, Beaver Creek, Deep Creek, Geyser Creek, and Miners Basin; the headwaters of Mill Creek probably also were glaciated, but the evidence is unclear. Pre-Bull Lake glaciations are not identified, but Bull Lake (?), Pinedale (?), and post-Pinedale (?) moraines are plentiful.

Evidence for massive landslip activity is widespread and includes: (1) shale intermixed in diamictons; (2) massive ground cracks; (3) large landslip levees; (4) convex-up cross valley profiles; (5) lineated flow topography; and (6) rock-block-slide topography. Numerous sites have plentiful igneous-rock rubble that accumulated through rock fall and snow avalanches onto unstable substrate, loading it to failure. Many rock-glacier-like deposits are thick, ice-cemented, igneous-rock rubble overlying unstable sedimentary substrate in which landslip basal shear predominates over ice as a motive mechanism; a few other exclusively ice-motivated rock glaciers also occur.

Twenty major landslip groups are now identified in the La Sal Mountains. These include Dorry Canyon, Hell Canyon, La Sal Pass, Southwest Peale - La Sal Pass, Southeast Peale, Dark Canyon, Hop Creek, Geyser Pass, Horse Creek - Geyser Pass, Lake Oowah, Boren Mesa, Warner Lake, Bald Mesa, Harpole Mesa, Fisher Ridge, Willow Basin, and Beaver Creek. Other areas of landslips also occur but are less defined.

Snow avalanches in the La Sal Mountains are plentiful and a major process of erosion and deposition, as well as providing local concentrations of snow load and subsequent meltwater that accelerate slope failure in unstable substrate.

Extensive mass movement in the La Sal Mountains poses little threat to present human land use. On the other hand, major logging, mining, road building, or water diversion projects would be likely to adversely affect unstable bedrock. Snow avalanches do pose a clear hazard to the ever increasing wintertime use of high areas but can be avoided through standard mountain safety techniques.

INTRODUCTION

Distinguishing between colluvium and till, a

[1]Professor, Dept. of Geography and Geology, University of Nebraska

[2] Omaha, NE

common problem throughout the Rocky Mountains, is especially troublesome in the La Sal Mountains because of the thorough spatial and temporal intermixing of the two deposit types with each other as well as with coarse fluvial material. Diamicton deposits, or unsorted, largely unstratified mixtures of fine and coarse terrigenous clastics without genetic designation, are easily mapped as such but preference is for differentiation into process types. The superficial similarity in diamicton sediment and landform produced by mass movement and glaciers in the La Sal Mountains has made mapping of Quaternary deposits there difficult for many previous workers (Carter and Gualtieri, 1958; Weir and Puffet, 1960; Richmond, 1962), but our recognition of this diamicton problem in the La Sals (Shroder and Sewell, 1980, 1981b) and methods to deal with it are producing selected results.

The La Sal Mountains of east-central Utah comprise three different massifs and numerous separate peaks above 3660 m in altitude (fig. 1). North Mountain Group is largest, having about 38 km^2 of mountain above 3350 m. Middle Mountain Group has about half as much area above that, and South Mountain Group has only about 6.4 km^2 above 3350 m. Each massif is composed of laccoliths intruded from central stocks of Tertiary age that uplifted late Paleozoic and Mesozoic sedimentary rock (Hunt, 1958). The resulting strongly jointed igneous rock and steeply dipping unstable sedimentary units at high altitudes have combined to produce a wide variety of mass-movement phenomena. Glaciation is now known to have occurred in only nine canyons during the Pleistocene, and the resulting deposits are stratigraphically above, coeval, and below mass-movement features. Heretofore, glacial origin has been stressed (Richmond, 1962), but this agrees neither with newly available geomorphologic data, nor with accepted notions of mass-balance-area ratios between likely accumulation and wastage zones. It is not, however, the purpose of this paper to delimit the total extent of glaciation in the La Sal Mountains. Rather we intend to show only general character and extent of prior glaciation, while emphasizing extent of major slope failures of late Pleistocene and Holocene age. In addition, rock glaciers are stressed because of diverse motive mechanisms caused by combinations of plentiful rock rubble, discontinuous permafrost, and unstable, landslip-prone substrate. Variations in rock-glacier morphologies and mechanics are common therefore, as well as in the High Plateaus (Flint and Denny, 1958; Shroder, 1972, 1978). Finally, the influence of snow avalanches on the origin of some landforms and collateral effects on rock glaciers and landslips is obvious in the La Sals and was studied.

Recent attention is given to the La Sal Mountains because of their varied potential for increased use. This paper suggests problems that may result from new timbering, road building, oil-well drilling, water diversion or dam building, uranium-vanadium mining, gold mining, nuclear waste disposal, skiing, and snowmobiling. Adequate planning and attention to geologic and geomorphologic detail should minimize serious difficulties.

CLIMATE

Considerable climatic range occurs in the La Sal Mountains because of local relief of about 2660 m from the nearby Colorado River at 1220 m altitude to Mount Peale at 3880 m. At lower altitudes mean annual precipitation is about 250 mm and mean annual temperature is about 12° C. Mean annual temperaure in the mountains is estimated to be about 7° C lower (Richmond, 1962, p. 11) and total annual precipitation is above 760 mm (Anonymous, 1960). Limited precipitation data from 2865 m altitude at Warner Lake show winter precipitation about double that of summer (Martin and Corbin, 1930, p. 16-21; Whaley and Lytton, 1979, p. 230). A winter-spring precipitation shadow occurs toward the southeast over the mountains (Alter, 1930, p. 3) because of the effect of the highlands on the dominant northwesterly storm tracks during that season.

Snowfall data (fig. 2) were collected intermittently from 1893 to 1931 (Martin and Corbin, 1930, p. 17-21) in the town of Moab which is at an altitude of about 1220 m. An annual average of 414 mm was recorded with an extreme fall in December and January (1915-16) of 1880 mm. Subsequently collection of such data was moved to a mountain-site snow course (La Sal Mtn. Lower) at 2682 m altitude in Geyser Pass (Whaley and Lytton, 1979, P. 228-230). In 1956 another nearby course (La Sal Mtn. Upper) was established at 2865 m. These two courses give mean values for April 1 of all years of about 245 mm and 432 mm depth, respectively. These relatively low total values do not reflect the high variation ranging from zero to more than double the average from year to year (fig. 2).

In the future these snow courses will transmit data through a satellite-based automatic snow-telemetry system (SNOTEL) on a demand or daily basis (Barton, 1977). This will revolutionize snow-depth data collection for high mountains and, although designed for water-supply forecasting, should enable in-

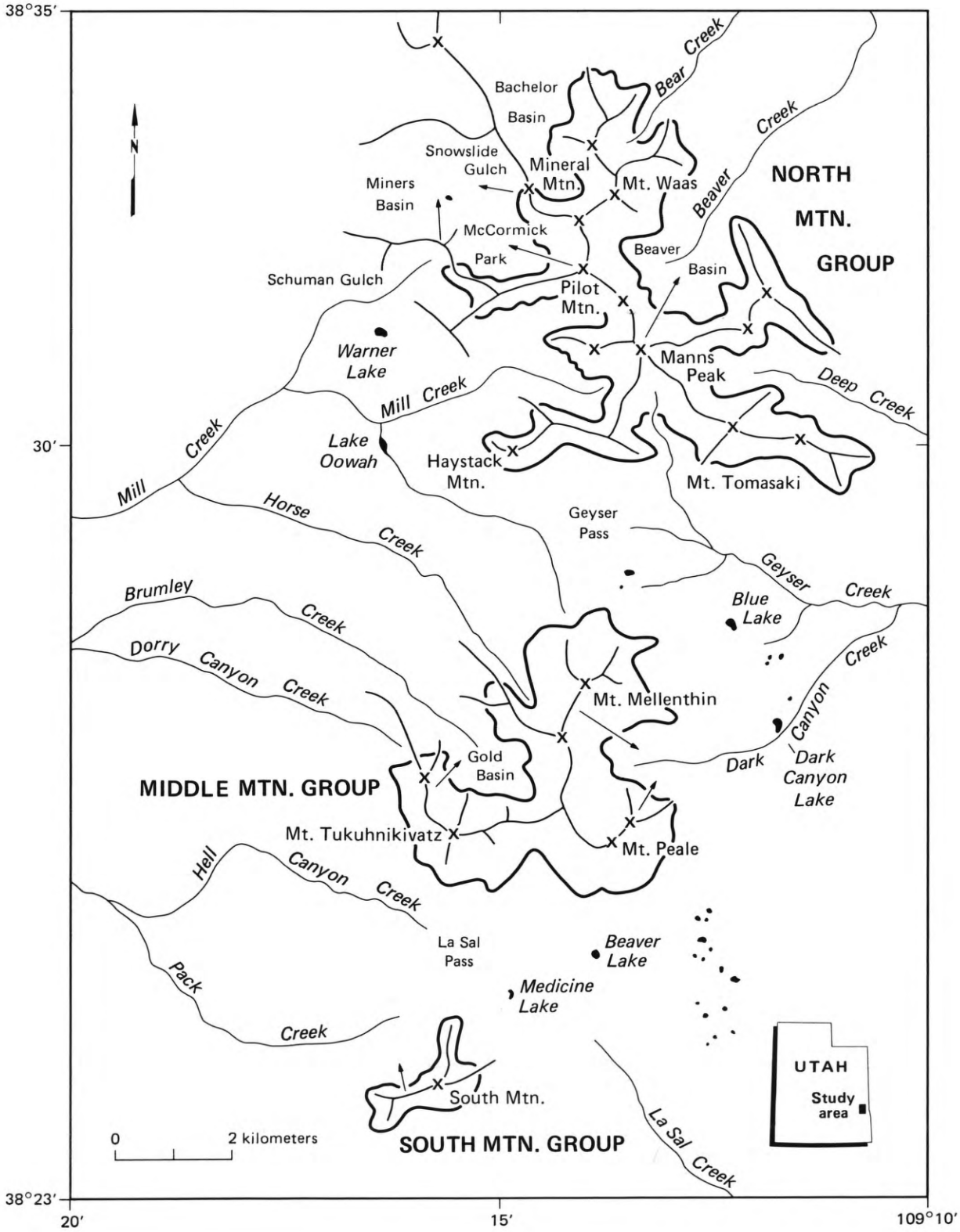


FIGURE 1. Index map of locations discussed in text. Lakes and ponds are in black; arrows indicate snow-avalanche paths discussed in text. Heavy solid lines between peaks are ridge crests; curvilinear lines encircling North, Middle, and South Mountain groups are 3350 m contour.

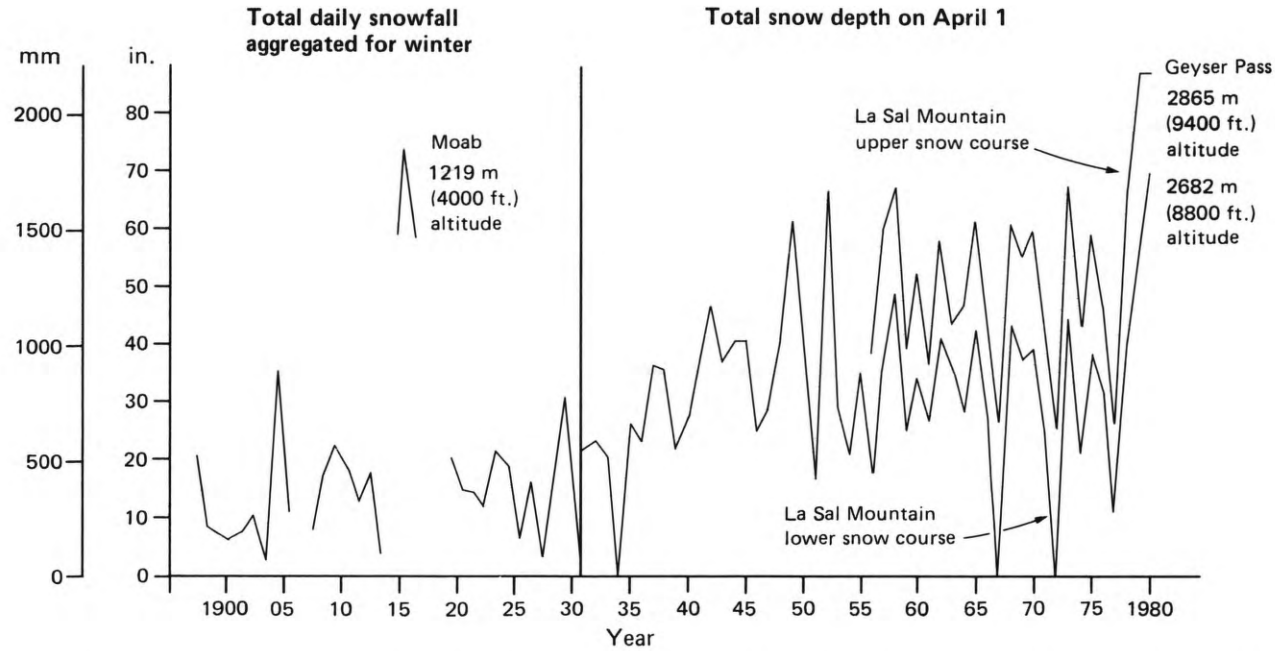


FIGURE 2. Snowfall depth data plotted for La Sal Mountain area. Data in Moab are discontinuous and were collected about 25 km WNW from Geyser Pass site.

creased understanding of water-induced mass-movement and snow avalanche hazards in the La Sal Mountains.

GENERAL GEOLOGY

Rocks exposed in the area range in age from Pennsylvanian to Tertiary, and there is an extensive varied sediment cover of Quaternary age (table 1) (Carter and Gualtieri, 1958; Hunt, 1958; Weir and Puffet, 1960; Richmond, 1962; Williams, 1964). Each of the three massifs of the La Sal Mountains consists of a stock surrounded by a cluster of laccoliths that radiate outward. In the North and South Mountain Groups the intrusions spread mostly in the unexposed, very incompetent Paradox Member of the Hermosa Formation of late Pennsylvanian age. The intrusion in the Middle Mountain Group was into the Morrison Formation of Late Jurassic age, and Mancos Shale of Late Cretaceous age. The physical injection of the stocks produced doming of the overlying sedimentary units (Hunt, 1958). Because the Morrison Formation and the Mancos Shale are two of the most landslip-prone units in Utah (Shroder, 1971), the intrusion and doming produced an ideal situation for extensive landslip failure.

In Pliocene (?) time the Castle Valley conglomerate resulted from deposition of gravel that was eroded from the intrusions and overlying sedimentary units. This is the oldest recognized diamicton that resulted from direct erosion of the mountain massif, and it appears to be partly fluvial because much is well stratified. Hunt (1958, p. 314) noted a paucity of both Morrison and Mancos Shale fragments in this diamicton, from which he deduced that those formations had been removed from the top of North Mountain dome prior to Quaternary time. A new road cut, however, has exposed the Castle Valley conglomerate and the overlying Harpole Mesa Formation of supposed early and middle Pleistocene age. In this new exposure occur two large blocks (10 m thick) of Morrison Formation shale, mudstone, and fine-grained sandstone that strike at nearly right angles to each other and dip nearly vertical, yet are only a few tens of meters apart (Shroder and Sewell, 1980). These blocks were almost certainly emplaced through mass movement and thus attest to the efficacy of this process in the origin of at least part of the older diamictons of the area.

A comprehensive and complex soil chronostratigraphy was derived for the La Sal Mountains and was combined with relative topographic position, surface expression, internal character, and relation to erosion surfaces and disconformities in order to subdivide the Quaternary (Richmond, 1962). Four formations with numerous members were recognized, each with multiple facies in various combinations of

TABLE 1. *Stratigraphy of the region of the La Sal Mountains*

SYSTEM	SERIES	FORMATION	MEMBER	THICK-NESS	DESCRIPTION
Quaternary	Holocene	Gold Basin*	Upper Lower		Multiple facies of clay, silt, sand, gravel, and diamictons of glacial, periglacial, mass wasting, fluvial, and eolian origin.
		Beaver Basin*	Upper Lower	Highly variable	
	Pleis-tocene	Placer Creel*	Upper Lower		
		Harpole Mesa*	Upper Middle Lower		
Tertiary		Conglomerate in Castle Valley		30+m	Competent conglomerate of variable-sized, local, igenous and sedimentary clastics.
		Intrusives of La Sal Mtns.			Competent but strongly jointed igneous rocks.
Cretaceous		Mancos Shale		240+	Incompetent marine shale with some modular limestone in concretions and some sandstone.
		Dakota Sandstone		15-35m	Sandstone, shale, siltstone and coal. Only sandstone is competent.
		Burro Canyon Formation		38m	Mostly competent sandstone and conglomerate.
Jurassic		Morrison Formation	Brushy Basin Shale	100-120m	Highly incompetent, variegated shale; some thin and lenticular sandstone beds.
			Salt Wash Sandstone	90+m	Competent conglomerate sandstone; some interbedded thin shale.
		Summerville Formation		15m	Incompetent thin-bedded shale and sandstone; some limestone.
		Entrada Sandstone			Competent massive and thick-bedded sandstone.
		Carmel Formation		7m	Incompetent sandstone and mudstone.
Jurassic and/ or Triassic(?)		Navajo Sandstone		90+m	Competent massive and thick-bedded, fine-grained sandstone.
		Kayenta Formation		75+	Incompetent sandstone, shale, mudstone and limestone.
Triassic		Wingate Sandstone		80-105+	Competent thick-bedded sandstone.
		Chinle Formation		100+m	Incompetent mudstones with some sandstone and conglomerate.
		Moekopi Formation		150+m	Incompetent shale and some sandstone.
Permian		Cutler Formation		300+m	Competent shale, sandstone, and conglomerate.
Permian (?) Pennsylvanian		Rico Formation		60?m	Competent sandstone and conglomerate; some limestone.
Pennsylvanian		Hermosa Formation	Unnamed Member	300+m	Competent limestone, shale and sandstone
			Paradox Member	3000+m	Highly incompetent evaporites.
Precambrian					Unexposed crystalline rocks.

*Recommend disuse of Quaternary formation names.

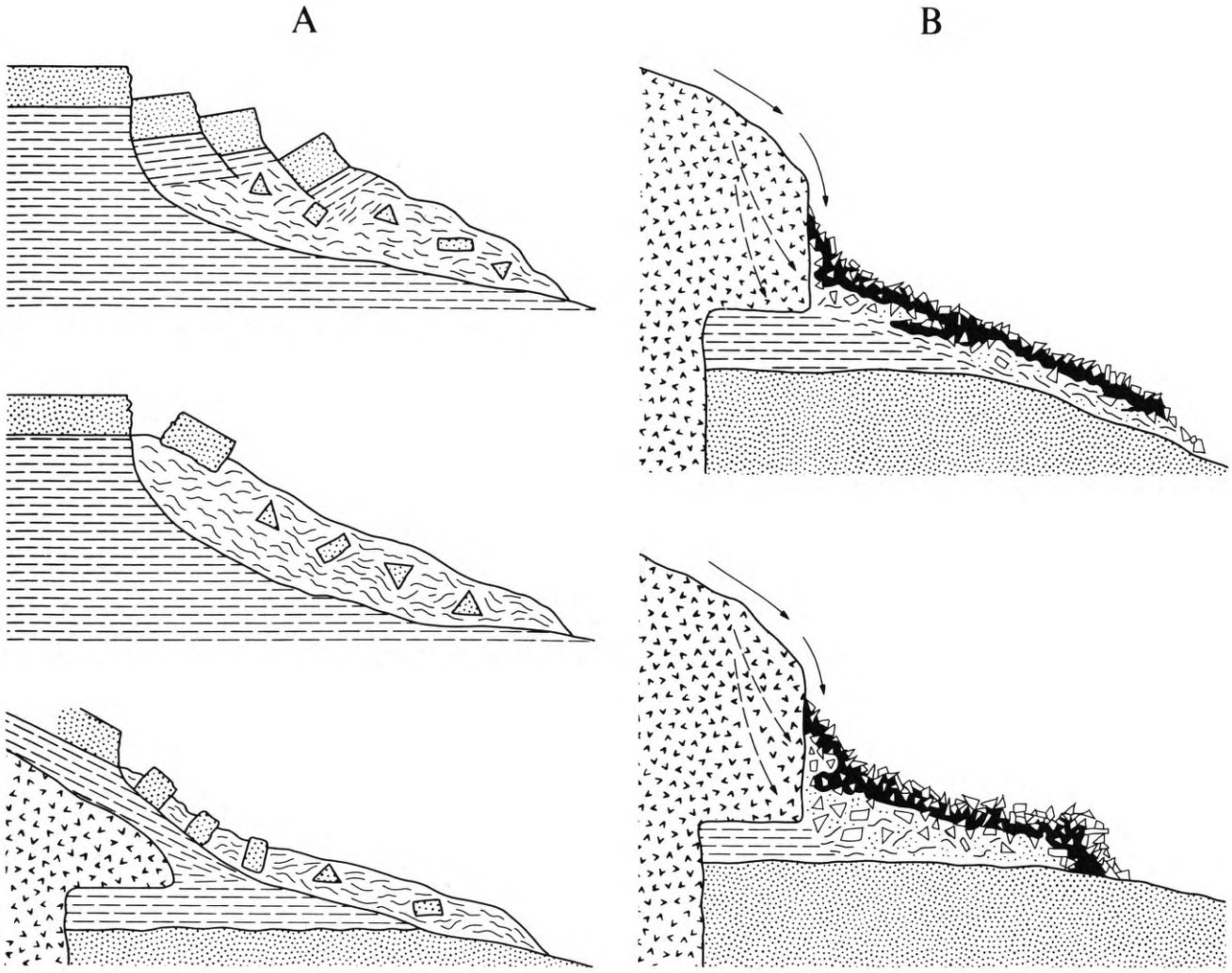


FIGURE 3. Large mass-movement types typical in the La Sal Mountains. A — three varieties of complex block slides and slow debris flows in stable and unstable bedrock that is flat lying or uparched around igneous intrusions. B — two varieties of talus accumulation, thin block streams, and thicker rock glaciers with ice cement and diverse configurations of stable and unstable bedrock. Dashed arrows show some possible paths of water through mountain slopes and beneath ice-cemented rock fragments. Solid arrows show paths of snow avalanches and rock fragments. All cross sections are based on actual examples, but some subsurface configurations are conjectural.

till, alluvium, colluvium, landslip and eolian material. The basal Harpole Mesa Formation was associated with three Spring Draw soils and correlated with Buffalo glaciations of the Rocky Mountains, and with Nebraskan through Sangamon stage deposits of midcontinent. The Placer Creek Formation was associated with Porcupine Ranch and Lackey Creek soils and correlated with Bull Lake stage. The overlying Beaver Basin Formation was associated with the Castle Creek soil and correlated with Pinedale stage. The youngest unit, the Gold Basin Formation, had the Spanish Valley soil on the lower member and was correlated with various post-Pinedale phenomena elsewhere.

Our investigation of this stratigraphy and the associated geomorphology has revealed numerous problems:

1. Many of the supposed tills or moraines come from areas without cirques, or without even large valleys or extensive uplands to support glaciation (Richmond, 1962, plate 6).

2. Some supposed till occurs at unlikely altitudes as low as 2260 m (Sec. 22, T. 28 S., R. 25 E., Richmond, 1962, plate 1).

3. Large lateral extent of the supposed tills is overwhelmingly disproportionate to any possible source areas (Richmond, 1962, plate 6). In some cases the

ice-wastage areas are two to four times as large as the possible accumulation areas.

4. Each of the four main supposed glaciations is most improbably less extensive than the last (Richmond, 1962, plate 6).

5. The four major formations described, each with two or three supposed separate till units, are not consistent with what is presently understood about subdividing the Bull Lake and Pinedale glaciations in either the type locality of the Wind River Mountains or elsewhere in the Rocky Mountains.

6. Only nine canyons show clear and unequivocal topographic and stratigraphic evidence of glaciation, instead of the nearly universal glaciations previously hypothesized (Richmond, 1962).

7. Of the two largest areas of supposed older tills on Bald Mesa to the northeast and La Sal Creek on the southeast, the La Sal Creek locality has an unlikely low-altitude southern exposure and no cirques (Richmond, 1962, plate 6).

8. Both Bald Mesa and La Sal Creek areas are the most extensive locations near the mountains of Morrison Formation and Mancos Shale outcrops, which suggests a probable landslip origin for the supposed tills in those areas.

9. Designations of "older moraine deposits" (Weir and Puffet, 1960) are also concentrated along outcrops of Morrison Formation and Mancos Shale.

10. The northern half of Bald Mesa contains no igneous rock fragments from supposed tills as previously described (Richmond, 1962, p.33). Instead the distribution of such clasts only on the south edge of Bald Mesa suggests remnants of an early alluvial fan deposit with its apex in Geyser Pass.

11. The Castle Creek soil described in the upper section of Pack Creek has an old gold-mining sluice in it and is therefore much younger than designated by Richmond (1962, fig. 49). This is reinforced by tree-ring dates and geomorphologic evidence that suggests soils buried by relatively recent flash-flood deposits.

12. New deep excavations by machine into lower altitude alluvial-fan and pediment gravels show disproportionate amounts of soil formation relative to their previously assigned age designations (N. Biggar, oral communication, 1980).

13. Analysis of clay mineral content from Gold Basin moraines fails to support the previous age designations (J.R. Giardino, oral communication, 1981).

14. Some of the supposed paleosols (Richmond, 1962) appear to be the result of other than pedogenic processes, or are so diversely interfingering, weakly developed, truncated, or morphologically indistinquishable from one another that they are not laterally traceable and of little use as stratigraphic markers.

The above mentioned prior designation of lithostratigraphic and soil-stratigraphic units in the La Sal Mountains seems to be considerably invalidated by the problems listed above. We recommend only informal use of prior terminology as well as redefinition and revision of the Quaternary units of the La Sal Mountains. Additional specific problems that have emerged from new identifications of landslips at each of the sites investigated in detail for this paper are discussed below.

DIAMICTON DIFFERENTIATION IN LA SAL MOUNTAINS

Glaciers, some periglacial processes, general mass wasting, and landslips produce remarkably similar deposits. In areas such as the La Sal Mountains where all these processes long have been active, complex interrelationships should be considered the norm. Unsorted, unstratified deposits should not be mapped as the result of any process unless rigorous selection procedures are utilized. In any case, we believe it far safer to assume widespread mass movement in areas of steeply dipping unstable bedrock at high altitudes than to assume extensive glaciation. The situation in the La Sals is further complicated by the fact that the various geomorphic processes certainly were controlled differently by climatic variation. Glaciers there would have been sensitive to both precipitation and temperature fluctuations, as would many periglacial processes. Much of the periglacial activity would likely have been shallow. The large bedrock failures, however, were probably more sensitive to precipitation variation only, or to ground-water changes as a permafrost increased or decreased at onset or end of glaciations. In addition, large landslips with rupture surfaces deep in unstable substrate are likely to be intermittently susceptible to minor precipitation fluctuations (Shroder, 1978; Shroder and Sewell, 1981a). All processes, however, produce diamictons of varying thickness (fig. 4).

The result is that differentiating between complexly interrelated glacial and landslip phenomena in the La Sal Mountains is difficult. Use of existing techniques for genetic discrimination between some deposits appears impossible in many places as age and reworking of the deposits increases. Three summers of fieldwork in the La Sals in 1977, 1979, and 1980, and brief visits in the spring of 1981 and 1983, make us confident that we can adequately differenti-

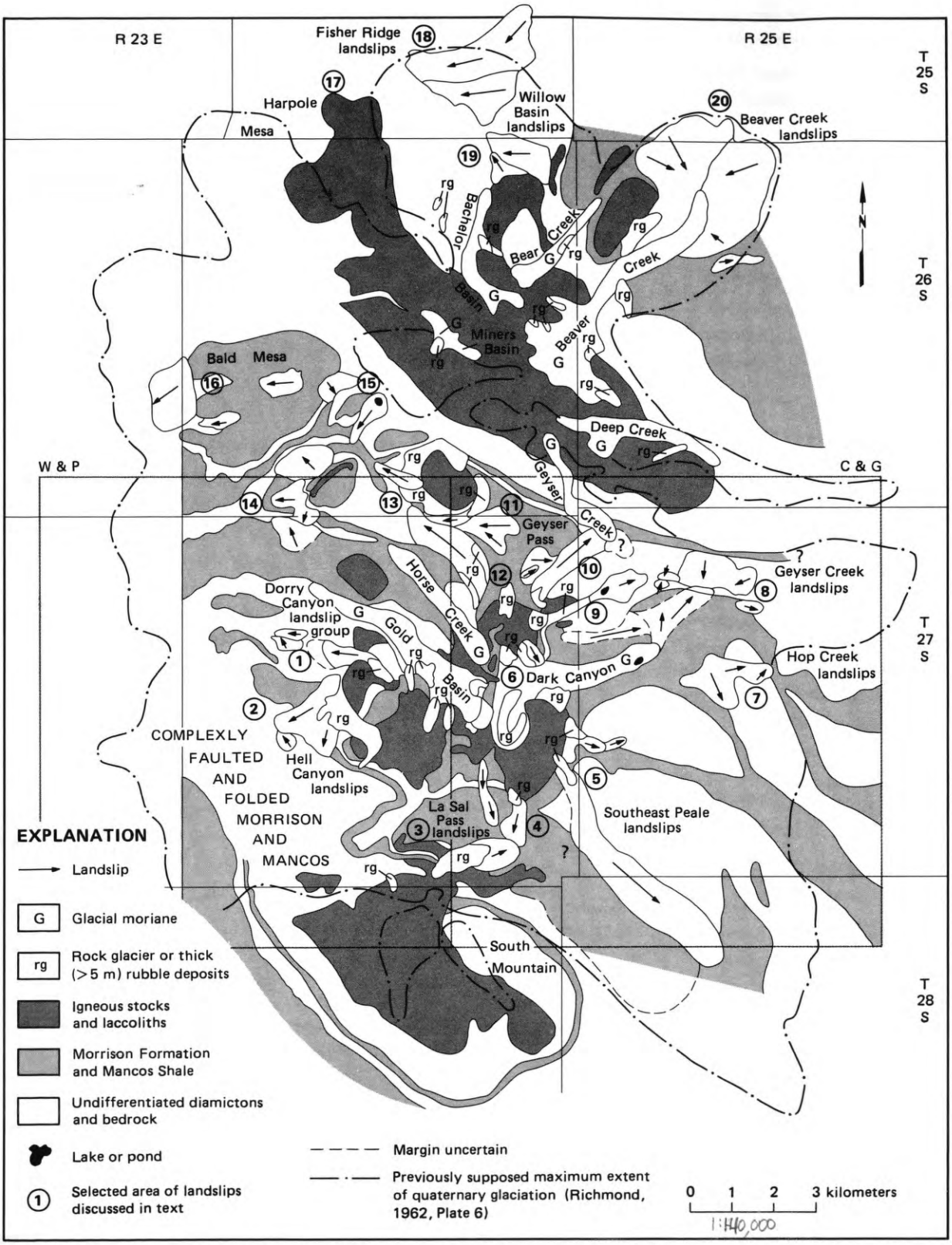


FIGURE 4. General map of diamicton deposits produced by mass movement and glaciation in the La Sal Mountains. Basic bedrock geology is after Hunt (1958). The rectangles designated "C & G" and "W & P" are bedrock mapped by Carter and Gualtieri (1958) and by Weir and Puffet (1960).

ate only some of those deposits of late Pleistocene and Holocene age. This is because we regard the topographic evidence as paramount; stratigraphic evidence alone in this context is essential but could be ambiguous. Some previous workers, however, have made attempts to differentiate between diamictons without appreciable landform evidence.

Landslip and glacial landforms in the La Sals are more easily distinguished on air photos than by inspection of the stratigraphy. In general, the quite different topographic character of large-scale mass movement, especially at the head or source areas, provides adequate criteria for differentiation. The bedrock context of the head areas is also important. The greatest problems of discrimination are between the lower toe regions of landslips and the end or recessional moraines. Clay-mineral determination might facilitate differentiation (J. R. Giardino, oral communication, 1981), but our reconnaissance suggests problems. Analytical results are likely to be enigmatic, considering the possibilities of variable types and quantities of clays produced by weathering in old moraine and landslips, coupled with those derived from clay-rich, terrestrial Morrison Formation mudstones and the marine Mancos Shale. Any given deposit may have too wide a variety of Mesozoic and Quaternary clays for use in more than general inventory work.

Fabric analysis might offer greater potential for discrimination between diamictons, but such data collection by us has been only cursory because of the need for precise delineation of a control group of known moraine and mass-movement types. Establishment of a standardized process-related fabric is necessary because of two problems: (1) many clasts in the La Sals are equant or platy with too few long axes; Yeend (1969, p. 39) recommended a long-to-short axis ratio of 2.5 to 1, or greater; (2) processes controlling till-fabric orientation and dip are still unclear (Andrews, 1971, p. 5), and mass-wasting fabrics are equally enigmatic because of the great diversity in process. The solution to the first problem is a sample large enough to ensure sufficient long axes, but the second problem is more difficult and may not be easily solved.

Fabric studies have been performed on a wide variety of glacial and mass-wasting phenomena up to now, but the results are somewhat ambiguous. Glacial fabrics generally show a preferred mode parallel to ice-flow direction and a strong tendency for up-glacier dips. These orientations are thought to result from lineation of clasts within the well-known imbricated basal and upward-curving shear surfaces in basal till and overlying ice (Yeend, 1969, p. 40-41). A secondary long-axis maximum occurs at right angles to the fabric-determined, ice-flow direction in some places but not in others (Holmes, 1941; Mills, 1977). Mass-wasting processes, on the other hand, involve such a wide variety of different materials and types of movement (Varnes, 1978) that fabric studies show only great diversity. Fabric patterns in talus are commonly isotropic, although a sub-horizontal component is indicated by some methods of analysis (Caine, 1967). Slope deposits resulting from creep or solifluction have preferred orientation of stones parallel to slope, and clasts in the front of solifluction lobes are oriented transverse to flow and parallel to the front (Lundquist, 1949; Watson, 1977). Preferred orientation of stones in frost-moved rubbles can be characteristically imbricated and dip upslope in central parts of lobes, although along the sides they can be oriented parallel to slope and dip inward beneath the lobe (Wahrhaftig, 1949, p. 223; Dahl, 1966; Potter and Moss, 1968). Rock glaciers show fabric variations such as near-vertical platy fragments oriented parallel to the length of furrows, irregular patterns of clast orientation interspersed with highly regular aligned clasts, and fabric parallel with slope along the sides and dipping steeply upslope in the lower furrow and ridge regimes (Wahrhaftig and Cox, 1959; Giardino, 1979; Giardino et al., 1979). Shale fragments in a mudflow on glacial till have been observed to dip preferentially into the slope (Harrison, 1957), and some earthflows and mudflows have a heavy concentration of clasts parallel to the slope and plunging downhill in the direction of movement (Yeend, 1969, P. 39-40). Richmond (1952) originally noted that rock glaciers in the La Sals had no fabric, but that block streams did. Later (1962) he did not report this observation in his final synthesis; in fact we have observed preferrential orientations at various places in the La Sals in both kinds of deposits.

From the above studies it is clear that anisotropic fabrics occur in many types of diamictons, but no ready means of differentiation between process types has emerged. This is an expectable result because shear surfaces responsible for fabric, both beneath and within ice as well as in many types of mass-movement and periglacial slope processes, may be similar to slow-flow phenomena. There has been limited success in fabric-analysis differentiation only in a few local contexts where glaciers moved down large valleys and mass-wasting came in at angles from the side slopes (Watson, 1977; oral communication, 1977). The superimposed or closely juxtaposed deposits then may be differentiated on the basis of

angular relations within and between fabrics considered against slope and flow directions (Yeend, 1969, p. 39-41). Many situations of ambiguity could exist, however. In addition, fabric studies on large, complex landslips have not been attempted yet in part because of problems with fabric inherited from other processes, the deposits of which can be incorporated into the landslip without significant reworking. Thus any possible fabric analysis in the La Sal Mountains must be carefully considered for potential of success weighed against the labor-intensive nature of the research. Fabric analysis is probably far more elucidating of process mechanics than in diamicton differentiation. Finally, scanning electron microscopy also has been attempted for analysis of grain-surface textures in diamicton differentiation, but variable process mechanics and reworked deposits have contributed to uncertainty; landform evidence is preferred (Madole, 1981; oral communication, 1981).

Our field analysis of the general problem of differentiation between glacial and landslip phenomena in the La Sal Mountains has been primarily devoted to recognition of the two major thick (5 m) deposit types on the basis of landforms, in order to pursue separate tracks of chronology building using all presently available methods. Thus radiocarbon dating, tree-ring analysis, clast-weathering ratios, soils, and the many other multiparameter criteria were all investigated and found feasible as a basis for future work.

Reconnaissance to date suggests that only nine canyons were glaciated during the Pleistocene; six in North Mountain Group, three in Middle Mountain Group, and none on South Mountain. A brief synopsis of our results on glaciers is presented herein; the landslip material is emphasized, however.

GLACIATION RECONNAISSANCE

We define glaciation in the La Sal Mountains more rigorously than previous workers. We use geomorphologic data as our first criteria, and stratigraphic data secondarily. The geomorphologic criteria constitute cirques, "U"-shaped valleys and polished or striated bedrock in drainage-basin headwaters, and lateral and looped morainal topography in the lower reaches. Soled and faceted cobbles are not regarded as diagnostic by us (cf. Richmond, 1962, p. 27, 30-31) because of similar such surfaces on freshly broken and water-eroded clasts in the La Sals. Valleys that head in unstable sedimentary rock, that are "V"-shaped, or that have obvious extensive landslip deposits in their lower portions are not considered to be of glacial origin. A few valleys have rock glaciers at their heads and most have internal ice, but these are regarded as essentially mass-movement phenomena, not glacial. These standards were used to designate the following valleys as the only clearly glaciated areas in the La Sal Mountains.

MIDDLE MOUNTAIN GLACIATION VALLEYS

The southernmost glaciations in the La Sal Mountains originated in Dark Canyon, Gold Basin, and Horse Creek (fig. 4). Each of the main cirques has a northern exposure and floor altitudes of 3350-3500 m. The Dark Canyon cirque floor is the highest in the La Sals and occurs directly below Mt. Peale (3877 m), which is the highest peak there.

Dark Canyon glaciations Erosional evidence for glaciation in the upper part of this canyon is unequivocal and several closed moraine loops occur down valley. The main cirque directly west of Mt. Peale is overdeepened, has a small seasonal tarn, polished boulders, and prominent smooth stoss and plucked lee topography. Two clear recessional-moraine loops occur in the upper part of the valley (fig. 5), and we tentatively correlate them with late Pinedale. The lower moraines around Dark Canyon Lake (fig. 4) are not easily subdivided because they are an irregular area of hummocky topography overlying the unstable Brushy Basin Member of the Morrison Formation and are thus transitional into extensive landslip deposits. Richmond (1962) mapped and described virtually his entire sequence in this valley, except for the youngest Gold Basin, but we are unable to replicate his chronology.

Gold Basin glaciations Gold Basin has one of the best preserved series of lateral and recessional moraines in the La Sal Mountains, as well as two major active rock glaciers with observed internal ice (fig. 4). Richmond (1962, plate 1) subdivided the rock glaciers into upper and lower facies of the Gold Basin Formation, but these assignments are untenable. Viewed from a distance or on air photos, a color variation appears as the result of age-linked lichen-growth or weathering phenomena and thus to variations in rock-glacier age. Instead, our close ground inspection reveals that strong lithologic variation in the surrounding cliffs causes great variation in clast color and size. While these variations do occur they are not age related and the existing chronology is incorrect. Elsewhere various snow avalanche-deposition features were interpreted as Gold Basin (post-Pinedale) till, as in Gold Basin or on the north slope of Mt. Mellenthin (Richmond, 1962, plate 1), but although the landform designation is wrong the youthful chronology is reasonably

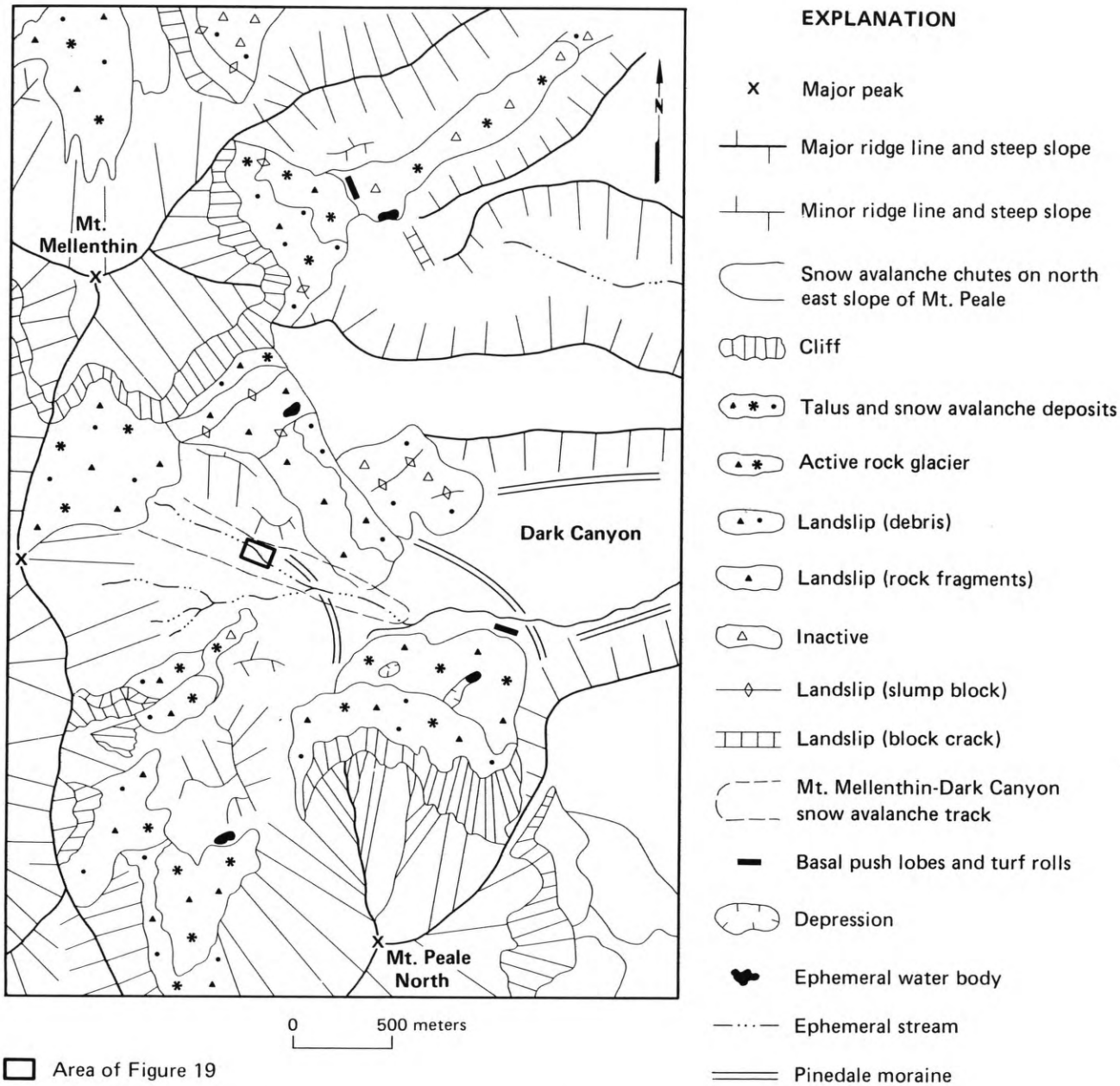


FIGURE 5. Sketch map of Dark Canyon area between Mt. Mellenthin and Mt. Peale drafted from aerial photograph. See Shroder and Putnam (1972) for further discussion of mass-movement symbology used here.

correct. In upper Gold Basin directly below the main cirques occurs a weakly developed broad moraine loop over a low bedrock step. This may be youngest Pinedale. A few hundred meters down canyon a prominent lateral moraine begins on the east side, and runs for about 2 km to where it disappears. In the vicinity of Brumley Ridge a higher and apparently older lateral moraine has closed a side valley to create a swamp and the moraine passes into a closed recessional loop further down (fig. 6). Below that point, at the upper entrance to the narrow canyon cut by Brumley Creek into Brumley Ridge, occurs a dissected moraine loop. There is thus morphologic evidence for four phases of youngest glaciation that we assign to Pinedale(?). Below the canyon cut through Brunley Ridge, where the valley opens out again, two additional moraine loops occur. The steeper east wall of the canyon has only one lateral moraine there, but the more gentle western slope has two laterals side by side that pass down valley into the main moraine

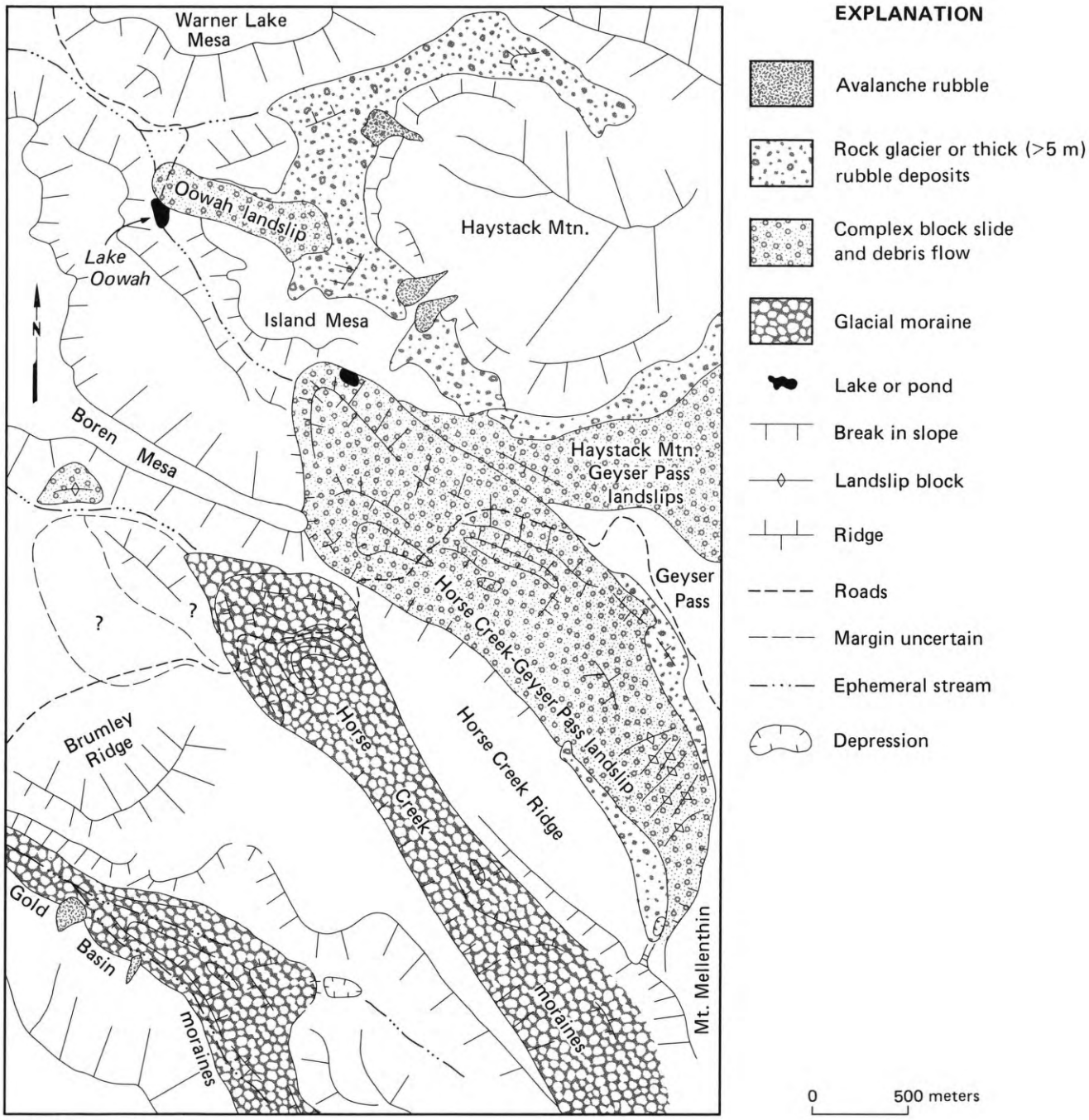


FIGURE 6. Sketch map of area west of Geyser Pass showing major moraines and landslips of the area. Drafted from aerial photographs. Compare with figure 4 for location.

loops. Soil and weathering-rind analyses on these laterals do not show significant age variation between moraines, although the two together are clearly older than the moraines up canyon. We tentatively assign these to a Bull Lake(?) age, which agrees with Richmond's assessment. We were unable to replicate the remainder of Richmond's chronology in Gold Basin, and Giardino (oral communication, 1981) concluded from his analysis of clay minerals in the deposits that a different chronology would be required to explain them.

Horse Creek glaciations The Horse Creek cirque

directly west of Mt. Mellenthin (fig. 4) is littered with considerable snow-avalanche rock debris but has no discernable moraines. About 1 kilometer down valley from the cirque head a lateral moraine occurs on the east about where the first recessional moraine loop is located (fig. 6). Near where the Geyser Pass road crosses the valley three other prominent recessional moraines occur. These constitute the younger glaciation from this valley. Two other subdued ridges of debris, indicated by queries (?) in figure 6, may be older moraine, but we are uncertain.

NORTH MOUNTAIN GLACIAL VALLEYS

The northern glaciations in the La Sal Mountains occurred in Geyser Creek, Miners Basin, Bachelor Basin, Bear Creek, Beaver Creek and Deep Creek (fig. 4). The average cirque floor altitude is about 3370 m with a small range of 3300-3400 m. The Geyser Creek glaciation was small because of the warmer southern exposure, Miners Basin was small because of the small and lower cirque basin, and Beaver Creek was the most expansive, probably because it is located in the lee of the tallest peaks in the north part of the range (Mt. Waas, 3758 m). Either or both of the heads of Dry Fork and Wet Fork of Mill Creek were possibly glaciated, but no clear cirque or morainal topography shows on the air photos. On the other hand, snow-avalanche and alluvial-fan deposition is well developed there and largely covers the valleys' sides. Future detailed fieldwork in these somewhat inaccessible valleys may solve the problem of whether or not they were glaciated.

Geyser Creek glaciations This area of former glaciation is unusual in that it has a poorly developed irregular cirque without steep head or side walls. Nevertheless, the abraided bedrock and moraine configuration show clearly that the area indeed has been glaciated. The valley begins in igneous rock of the North Mountain Group but passes through a narrow gap in upturned steeply dipping late Paleozoic and Mesozoic rock before exiting into Scorups Pasture area near Geyser Pass.

One possible moraine loop occurs in the cirque, but three obvious recessional moraines are located down valley from the cirque; one occurs directly on the Geyser Pass side of the narrow gap, and the other two are part of the headwater area of Geyser Creek. The lowermost of these moraines is in contact with the East Geyser Pass landslips (table 2) and, because of similar appearance and lack of disturbance by the failure, may closely relate in time to it. On the basis of similar morphology and soil development we equate these moraines with the Pinedale(?) tills in Dark Canyon and Gold Basin. Bull Lake tills may occur further down valley, but they have been thoroughly eroded and mixed by later landslips and fluvial erosion.

Miners Basin glaciations Miners Basin, formerly thought to have been extensively glaciated and presumably the supposed source of the type Harpole Mesa Formaton tills down valley (Richmond, 1962), is difficult to assess because of a number of complicating factors. A thin band of undifferentiated Pennsylvannian, Permian, and Triassic sedimentary rock passes over 2 km up the axis of the lower valley and produces valley side benches that mimic lateral moraines where covered with colluvium. This steep narrow lower valley has several bedrock buttes in it that would have been eroded by any glaciers capable of emplacing the supposed lower till of the Placer Creek Formation at the mouth of the canyon as Richmond (1962, p. 45, fig. 13) suggested. On the other hand, his upper till of the same formation in the mining camp area is obviously a well-developed closed moraine loop. Other possible moraines in the upper valley are not clear.

Bachelor Basin glaciations This area has two clear cirques at its head, each with a small moraine at its outer margin. The higher cirque below Mt. Waas contains a prominent snow-avalanche fan tongue. Numerous other snow-avalanche depositional features occur further down the valley, including one prominent rock glacier. No obvious moraines occur anywhere, except possibly two dissected breaks in slope at the valley mouth, which Richmond (1962) identified as upper and lower till of the Placer Creek Formation.

Bear Creek glaciations This area has no well-developed cirque or "U"-shaped valley, but this is probably because of the extensive snow-avalanche deposition, streams of rubble, and rock glaciers that have infilled so much of the valley. One possible moraine occurs in the upper basin, but Richmond (1962) did not identify it. A probable moraine loop mapped as upper Placer Creek till by Richmond (1962, plate 1) occurs somewhat below the mouth of the valley. Other older till may occur down valley from this, but the presence of the unstable Morrison Formation and Mancos Shale there with downslope dips of 25° makes doubtful any diamicton identified as glacial.

Beaver Creek glaciations The glacial activity in Beaver Basin was the most extensive and long lived of any we can identify in the La Sal Mountains. Although we cannot replicate all of Richmond's (1962)

TABLE 2. *Description of basic character of 20 major landslip groups in higher parts of La Sal Mountains*

LANDSLIP GROUP NAME & NO.	FIG. NO.	OVERLYING OR SOURCE ROCK	UNDERLYING ROCK	STRUCTURE	CHARACTERISTICS OF HEAD AREAS	CHARACTERISTICS OF FOOT & TOE AREAS	SPECIAL FEATURES & CAUSES
Dorrey Canyon 1	4 7 8	Igneous rock rubble, sandstone blocks & rubble	Friable mudstones Brushy Basin Member Morrison Formation	Intrusion updomed sed. rock to 15°W. dip downslope	Snow-avalanche & talus-fed rock glacier at head; protalus, longitudinal & transv. ridges & furrows	Two lobes of boulder-laden debris; undrained depressions; several large landslip blocks below main toe	Rubble loaded on unstable mudstone at head; downslope dip.
Hell Canyon 2	4 7 9	Igneous rock rubble, sandstone blocks & rubble	Friable mudstones Brushy Basin Member Morrison Formation	Intrusion updomed sed. rock to 17°SW. dip downslope	Head is smooth sandstone dipslope; slide & slump blocks; talus, protalus; linear flow features; hummocky	Landslip from both sides restricts Hell Canyon & causes upstream alluviation & downstream knickpoint	Rubble loaded on unstable mudstone at head; downslope dip; strong erosion in Hell Canyon
La Sal Pass 3	4 10	Igneous rock rubble, sandstone blocks & rubble	Mancos Shale Brushy Basin Member Morrison Formation	Intrusion updomed sed. rock to 40-9°SE. dip downslope	Snow-avalanche & talus-fed at heads; linear flow features; Lndslp I moved bldr group w/o mixing	Medicine & Beaver Lakes impounded by flow; transverse ridges & furrows	Rubble loaded on unstable mudstone at head; downslope dip
Southwest Peale - La Sal Pass 4	4 11	La Sal Pass landslips provided debris	Mancos Shale	Dip about 10° SE downslope	Hummocky topography & unclear margins for over 4 km	Toe area not clear but hummocky topography throughout	Other landslips loading head; unstable dipping Mancos Shale
Southeast Peale 5	4 11 12	Igneous rock rubble, sandstone blocks & rubble of Dakota Sandstone & Burro Canyon Formations	Mainly Mancos Shale some Brushy Basin Member Morrison Formation	Steep dips close to Peale, but 9°SE. downslope several km away	Complex head w/many generations movement; youngest 430 ± 65 B.P. talus, rock glaciers; rubble lobes moved intact on debris flow; large slump block	Hummocky topography & many ponds, linear shear zones on right flank where flow turns from SSE to SE	Largest & longest landslip in area; 7 km long, up to 1 km wide; rubble loading head & mudstones dipping downslope
Dark Canyon 6	4 5 13	Igneous rock rubble & sed. rock	Mainly Mancos Shale & Morrison Formation	Numerous faults & sed. rock updomed & fractured by intrusions	Extensive ice-cemented rock glaciers, talus cones & slump blocks below Mts. Peale & Mellenthin	Some rock glaciers have steep front at angle of repose; other similar features do not	Thick accumulations of ice-cemented rubble, in part above unstable substrate
Hop Creek 7	4 15	Dakota Ss. & Burrow Canyon Formation	Brushy Basin Member Morrison Formation	12-28°E. dip downslope	Head is convex in direction of failure; 600-m-long slump block; ponds	Movement to both NE & SE; common ponds & hummocky topography	Several episodes of movement; unstable substrate
Geyser Creek 8	4 15	Dark Canyon moraines	Mainly Morrison Formation	> 45° downslope along North Mountain pluton	Heads along North Mtn. & is cuspate; slump blocks common	Lower flows forced Geyser Creek south to undercut bank; extensive upstream alluviation	Cause mainly steep dips in unstable rock & Geyser Creek downcutting
Blue Lake 9	4 5 15	Igneous rock rubble from Mellenthin pluton	Mainly Mancos Shale & Morrison Formation	Mellenthin pluton both updomes & overlies sed. rock	Head has two large slump blocks on both flanks of main rock-fragment mass; upper lobe active	Lower front inactive but still at angle of repose; main debris flow mass below with Blue Lake inactive	Caused by ice-cemented rock rubble over unstable substrate

Continued

TABLE 2. *Description of basic character of 20 major landslip groups in higher parts of La Sal Mountains (Continued)*

LANDSLIP GROUP NAME & NO.	FIG. NO.	OVERLYING OR SOURCE ROCK	UNDERLYING ROCK	STRUCTURE	CHARACTERISTICS OF HEAD AREAS	CHARACTERISTICS OF FOOT & TOE AREAS	SPECIAL FEATURES & CAUSES
East Geyser Pass 10	4 15	Sandstone blocks & igneous rock rubble from Mellenthin	Brushy Basin Member Morrison Formation & Mancos Shale	18-25°N. downslope dip	Several movement episodes; smaller NW failure cut by 500-m-long slump block & flow to NE	Failure produced linear flow terrain caused by shear zones. Overlain in part by Geyser Creek moraine	Caused by steep dips close to Mt. Mellenthin & friable shale & mudstone
Haystack Mtn. - Geyser Pass 11	4 6 15	Igneous rock rubble from Haystack Mtn. pluton	Mancos Shale	Near 90° dips around North Mtn. pluton	Rubble lobes & sheets near Haystack Mtn; slump blocks in Mancos Shale	Several episodes of movement; hummocky topography, debris flow over rubble	Igneous rock rubble load & downslope dips
Horse Creek - Geyser Pass 12	4 6 15	Igneous rock rubble from Mt. Mellenthin; block sandstone from Burro Canyon Formation & Dakota Sandstone	Brushy Basin Member Morrison Formation	15°N. dip downslope	>10 large (>300 m) rock-block-slides of sandstone with lateral shear zones of igneous rock rubble on flanks	Flow NW to Boren Mesa & then down 16°N dip slope into Oowah Creek to form pond	Possible melt-water from Horse Creek glacier
Lake Oowah 13	4 6	Igneous rock rubble from Haystack Mtn.	Mancos Shale	Steep dips close to Haystack Mtn. pluton	Snow-avalanche & talus-fed rock glacier & rubble sheets around Haystack Mtn.	Flow down into valley to impound Lake Oowah	Thick igneous rock rubble load on Mancos Shale
Boren Mesa 14	4	Mancos Shale, Mill Creek sill, Dakota Ss., Burro Canyon Formation	Brushy Basin Member Morrison Formation	5°W-NW downslope dip	Slump blocks around nose of Boren Mesa at intersection of Mill & Horse Creeks	Several landslip benches caused by resistant rock layers	Deep dissection of two creeks; unstable rock; disruption by sill
Warner Lake 15	4	Mancos Shale, Dakota Ss. & Burro Canyon Formation	Brushy Basin Member Morrison Formation	Varied dips in upturned edges at North Mtn. pluton	Large slump block on backslope of which occurs Warner Lake	Lower zone of flow to Mill Creek valley; subsidiary failures on Wilcox Flat near Schuman Gulch	Uptilting and disruption near North Mtn. pluton; water load from canal future hazard?
Bald Mesa 16	4	Colluvium & alluvium over Dakota Ss. & Burro Canyon Formation	Mancos Shale on mesa top; Morrison on Jimmy Keen Flat below mesa scarp	4°W dip downslope	Many landslip blocks & hummocky topography; landslip amphitheatres; cambering, topple failure & slump blocks	Several episodes of movement; scarp retreat through landslips	Landslip drag folds, not ice loading as previously thought (Richmond, 1962)
Harpole Mesa 17	4 16	Colluvium & alluvium	Morrison Formation	Bedrock not exposed but steep dips next to North Mtn. pluton	Most landslips covered except for several large blocks of vertical Morrison	Blocks strike at right angles & appear not related to faults or intrusion	Caused by uplift & disruption near pluton
Fisher Ridge 18	4	Dakota Sandstone & Burro	Morrison Formation	15°SW dip downslope; Castle	Several episodes of movement; 14-16 land-	Linear flow features & hummocky topography;	Caused by dip downslope & fault; main scarp on up-

Continued

TABLE 2. *Description of basic character of 20 major landslip groups in higher parts of La Sal Mountains (Continued)*

LANDSLIP GROUP NAME & NO.	FIG. NO.	OVERLYING OR SOURCE ROCK	UNDERLYING ROCK	STRUCTURE	CHARACTERISTICS OF HEAD AREAS	CHARACTERISTICS OF FOOT & TOE AREAS	SPECIAL FEATURES & CAUSES
		Canyon Formation		Valley fault crosses slope	slip blocks below steep main scarp	toe interdigitates with alluvium of Castle Valley	thrown fault block
Willow Basin 19	4	Igneous rock rubble from North Mtn. pluton & Willow Basin sill; Dakota Ss. & Burro Canyon Formation	Morrison Formation	Downslope dip	Arcuate main scarp curves nearly 90°; Hunt (1958) mapped landslip scarp as fault	Downslope flowed part of debris & rock fragments	Cause probably downslope dip, incompetent rocks, rubble load & pluton disruption
Beaver Creek 20	4	Mancos Shale, Dakota Sandstone, Burro Canyon Formation & rubble from Beaver Creek laccolith	Morrison Formation Mancos Shale	Gentle dips; downcutting by Beaver Creek	Several episodes of movement from scarp surrounding Beaver Creek; rock glacier from Beaver Creek laccolith on Mancos Shale	Landslip blocks, ponds, swamps, hummocky topography	Entrenched Beaver Creek, unstable rock, & probable glacial meltwater

previous effort there, his basic moraine configurations in the lower valley seem reasonable. This lower till of the Placer Creek Formation occurs as two ridges without discernible age variation between them (Richmond, 1962, fig. 31), similar to our observation of the two lowest, positively identified double lateral moraines from Gold Basin. This double moraine in Bull Lake time, but without observable variation in age criteria between the two, has been observed elsewhere (Roy and Hall, 1980). The double moraine passes up Beaver Creek valley as a prominent lateral moraine on both sides that Richmond (1962, plate 6) also mapped as late Placer Creek (Bull Lake). Moraines of Beaver Basin time (Pinedale?) also occur in the greater cirque basin but are not well developed or easy to delinate spatially or temporally. A post-Pinedale moraine (Gold Basin Formation; Richmond, 1962, plate 6 and fig. 47) occurs high in the northernmost cirque of Beaver Basin and is one of the youngest moraines that we can identify.

Deep Creek glaciations Glaciers in the headwater of Deep Creek area carved several cirques and deposited minor moraines. Down valley on the north side occur three lateral moraines: one old one was defined on the basis of isolated diorite boulders overlying sedimentary bedrock, and some tonal variation on air photos; and the two lower younger moraines are morphologically distinct. The middle moraine has a large (>5 m) fractured diorite erratic on its upper surface.

The older lateral moraine passes out of the canyon high on the north side onto an eroded ridge with diorite clasts and terminates. Richmond (1962, plate 1) mapped this as upper Harpole Mesa till. The other two younger moraines terminate in the bottom of the deep valley; both were considered lower Placer Creek till by Richmond (1962, plate 1), but they appear younger to us. A large, marginally active rock glacier originating on the north slope of Mt. Tomasaki overlies the front of the middle moraine.

There are probably many as yet undelineated low moraines hidden beneath forest cover in the La Sal Mountains. Our general ground reconnaissance of all the glacial valleys involved close-spaced traverses in several canyons in order to discover several moraine loops only a few meters high that do not show well under forest cover on the aerial photographs. We feel that intensive ground survey is the only way to map these, but this will require months of vigorous effort in difficult, overgrown terrain in order to define the features. Further studies could also integrate the mass-movement with the glacial landforms and

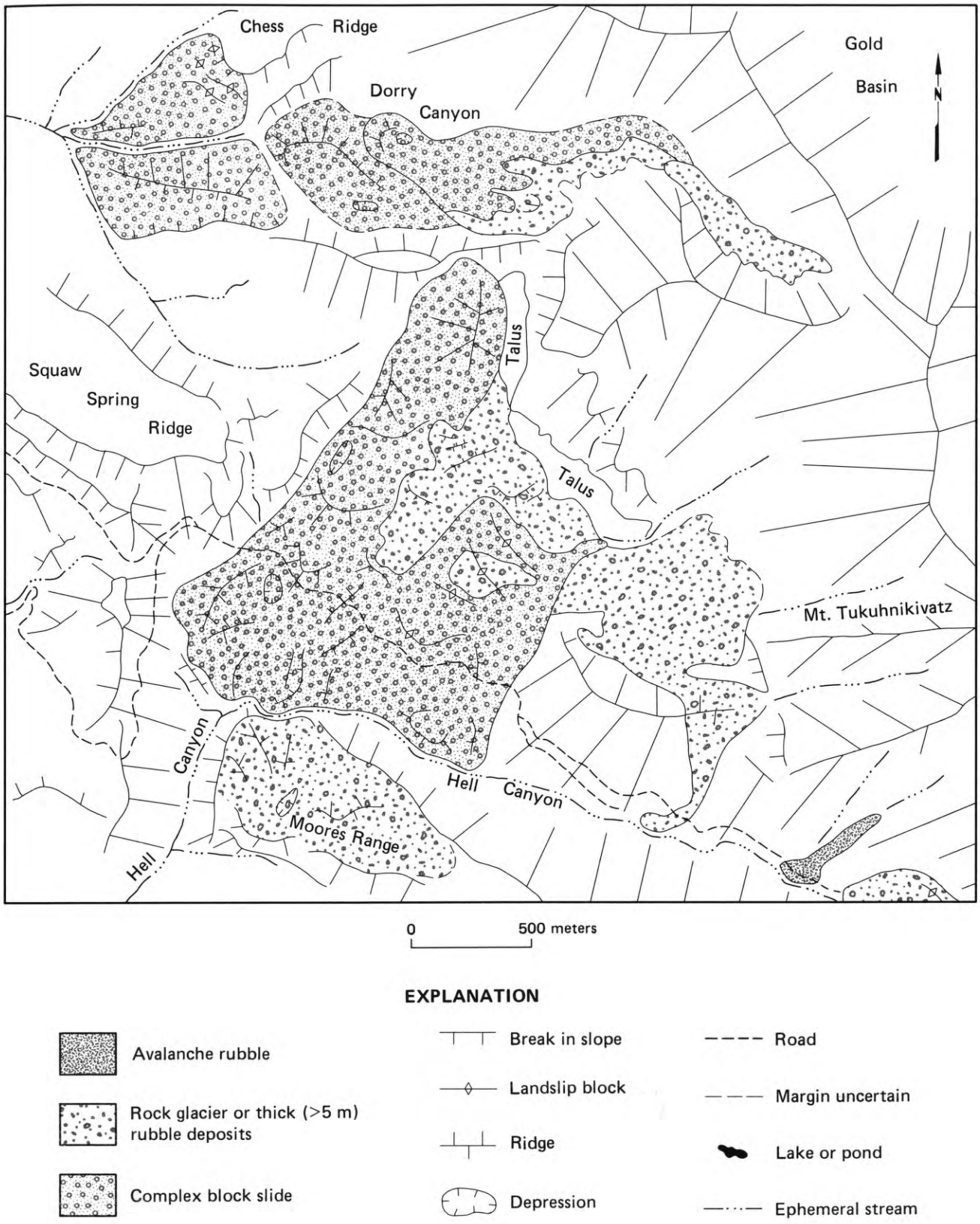


FIGURE 7. Sketch map of Dorry and Hell Canyon areas, drafted from aerial photograph.

define a new comprehensive chronology. At present only a general description of moraine location, configuration and chronology is possible.

MASS MOVEMENT IN THE LA SAL MOUNTAINS

Mass movement as used here includes landslips and snow avalanches. Of these two, landslips have moved by far the greater mass of material, and a conservative estimate of volume of obviously landslip-transported material is 4×10^6 m^3. Snow avalanches are much less obvious as transporters of rock debris, but significant quantities are deposited in more localized sites. In addition, snow avalanches melt to produce considerable water concentrations in small areas, which further aid in slope movement later on.

In addition to their geomorphic role, both landslips and snow avalanches constitute natural hazards that have been little studied in the La Sal Mountains. With the exception of an occasional rare rockfall, most of the landslips present little danger to life. On the other hand, with increasing development in the area, the road building and water management associated with the newly proposed oil-well drilling in Gold Basin, gold placer operations in upper Castle Valley, or other similar activities could be seriously affected. Snow avalanches present danger to both life and development in many areas, however, especially in light of the abnormally high snowfalls of the last few years.

LANDSLIPS

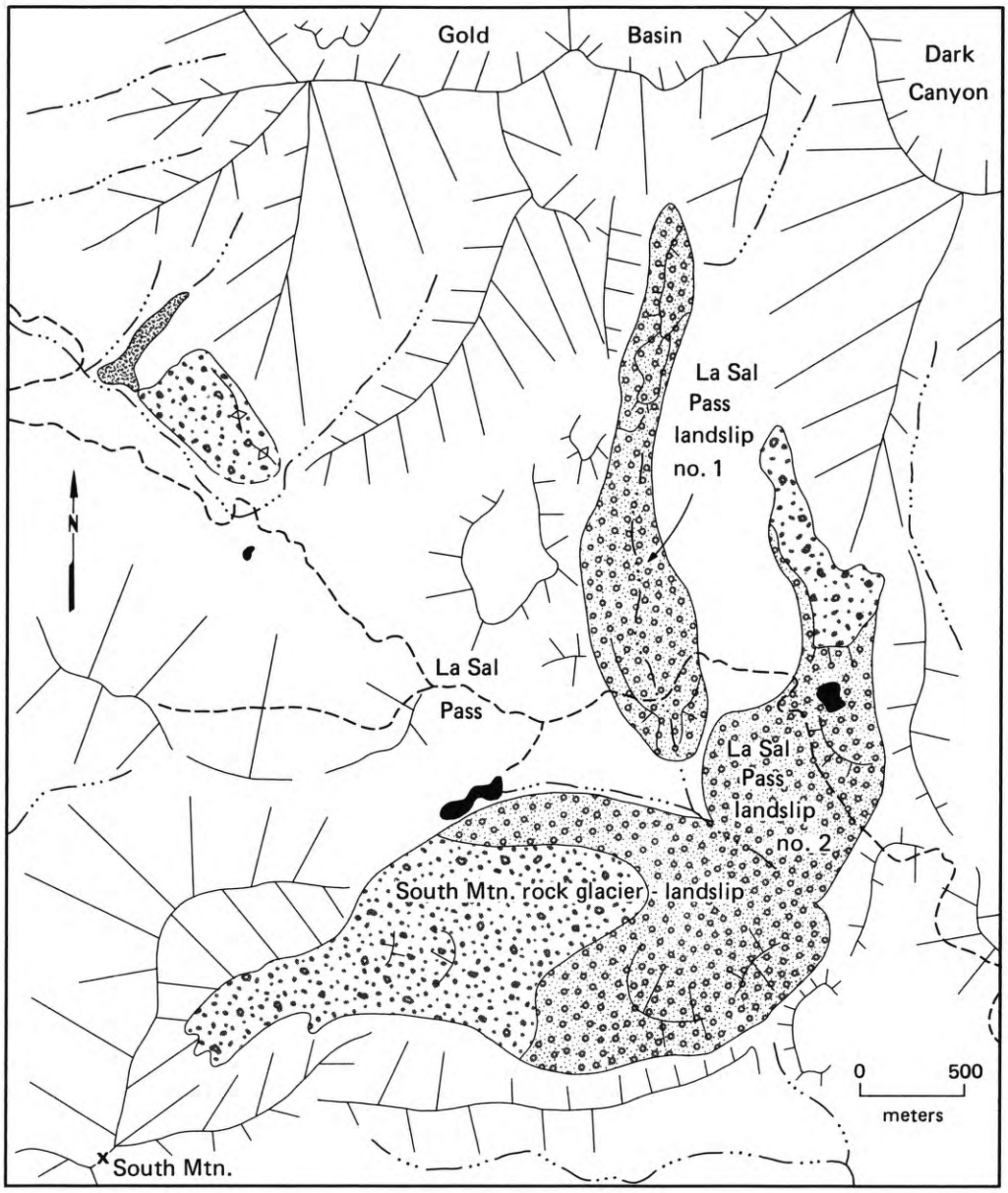
We identify a minimum of 20 major landslip groups in the higher parts of the La Sal Mountains (table 2; fig. 4). Numbers following the group names refer to location on figure 4. Actually there are far more landslips than this, but we have excluded many of smaller size or lower altitude. All of those discussed herein involve mainly bedrock failure, are several tens of meters thick, and have various combinations of unstable mudstones, sandstone blocks, and igneous rubble. Most of the landslip groups involve several different directions and times of movement. None of the failures selected here are shorter than 500 m and the largest is over 7 km long. With a few minor exceptions all would be classified as complex because they involve various types of slides

FIGURE 8. Complex landslip filling head of "V"-shaped Dorry Canyon. The convex-upward toe (outlined by conifers on the lower slopes in the photograph center) is a typical topographic expression of large, slow-debris-flow lobes derived from unstable bedrock. Areas of hummocky topography in both right and left foreground are other landslips mapped in figure 7.

FIGURE 9. Dip slope in foreground developed on sandstone unit in Brushy Basin Member of Morrison Formation. The mountain in the background is a laccolith. The ridge on the right foreground is the head of Hell Canyon landslip which moved to the right down the dip slope.

and flows (Varnes, 1978). Two main kinds of complex landslip occur (fig. 3). The first involves large blocks of sedimentary rock that topple forward, slump with backward rotation, or slide downward and outward in a linear nonrotational or translational fashion. The lower part of these landslips is almost everywhere a slow debris flow developed in plastic mudstones. The second common landslip type occurs where igneous-rock fragments from talus, block streams, and rock glaciers accumulate on unstable mudstone, load it to failure, and cause massive slow debris flows. These are unusual landslips because most have ground-ice permafrost in the rock-fragment-matrix interstices that contributes to greater water saturation of the underlying mudstones. Movement of many of these is probably also related to increased hydrostatic head beneath impermeable frozen rubble (Shroder, 1978).

This leads directly to the problem of what constitutes a mappable rock glacier. Previously there has been a strong trend to map as rock glaciers all thick rock-rubble deposits with internal ice, longitudinal and transverse furrows and ridges, and a steep front at the angle of repose, with the idea that the ice is somehow responsible for the movement and that basal shear is not supposed to be a significant factor (Wahrhaftig and Cox, 1959; White, 1976). This idea seems to hold well in areas of stable competent bedrock such as granite, as in the Colorado Front Range, quarzite as in the Uinta Mountains, or diorites as in the southwest part of Gold Basin in the La Sal Mountains. In many places there are sediments or sedimentary rocks upon which considerable rock fragments may accumulate to force subsequent failure of a landslip type where basal slip may predominate. In those deposits with sufficient basal shear, the steep front at the angle of repose may not occur because of elimination of the requisite basal-shear resistance necessary to pile large boulders to the angle of repose. Nevertheless, the remainder of the typical morphology may still develop, expecially the internal ice and the transverse furrows and ridges. Appearances of all possible varieties of this phenomena in the La Sal Mountains suggest a continuum of process between certain rock glaciers and landslips. This is an unpopular idea because of the



EXPLANATION

- Avalanche rubble
- Rock glacier or thick (>5 m) rubble deposits
- Complex block slide and debris flow
- Lake or pond
- Break in slope
- Landslip block
- Ridge
- Roads
- Margin uncertain
- Ephemeral stream

FIGURE 10. Sketch map drafted from aerial photographs of landslips in La Sal Pass area directly southwest of Mt. Peale.

"tyranny of the pigeonhole" and was dealt with in this project by mapping rock glaciers and thick (> 5 m) rubble deposits as a single entity (fig. 4).

Both of these two main kinds of landslips in the La Sals move through long-term slow flow, and where measurements of similar active phenomena have been made (Shroder, 1972; Shroder and Putnam, 1972), a meter or two a year is characteristic. These features also seem to be intermittently active, retrogressive, and turbulent in the sense of different parts moving more or faster in scattered places at various times (Shroder, 1978). A few of the examples in the La Sal Mountains are obviously active at the present time, but only the Dark Canyon landslip offers prospects for a long-term chronology of movement to be developed from the several hundred trees (*Picea engelmannii* and *Pinus flexilis*) affected by long-term failure. Analysis of extensive cores and cross sections taken suggests a relationship to precipitation or to snow-avalanche meltwater infiltration, as in other similar cases elsewhere (Shroder, 1978).

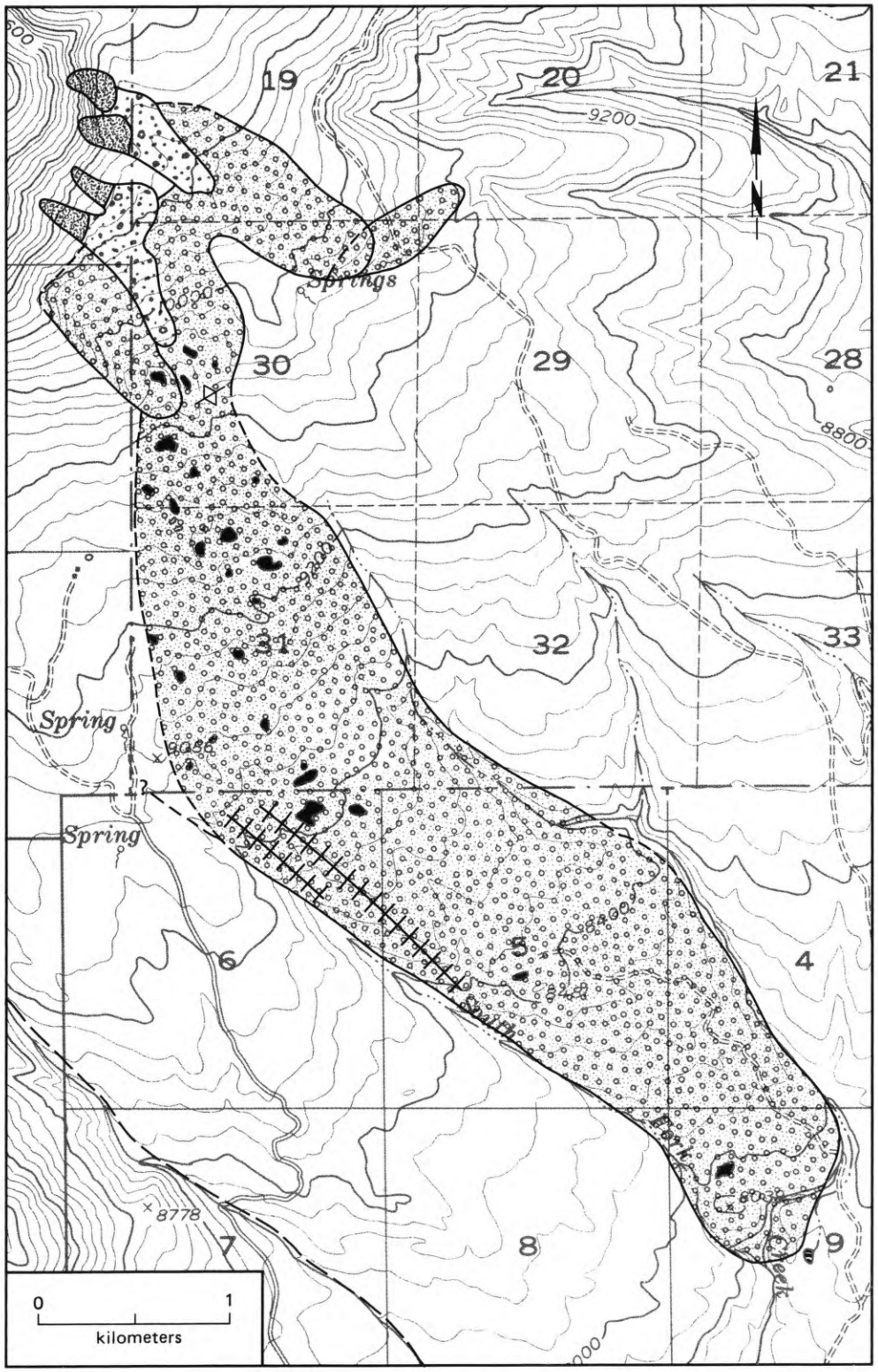
Twenty major landslip groups have been selected as the most representative of the phenomena in the La Sal Mountains (table 2); detailed descriptions below of four examples best exemplify all the varieties and causes of landslips in the La Sal Mountains. The landslip groups of Southeast Peale, Dark Canyon, Blue Lake, and Horse Creek - Geyser Pass around Mt. Peale and Mt. Mellenthin show a full range of typical features and some of the problems of previous workers who did not recognize certain deposits as landslip in origin.

Southeast Peale landslips (5) This failure is about 7 km long and over 1 km wide in places (fig. 11). The many ponds on its surface probably were responsible partly for the moraine designation of this area by Richmond (1962) and Carter and Gualtieri (1958), in spite of a complete absence of cirques and the presence of unstable substrate. The Brushy Basin Member of the Morrison Formation outcrops extensively at the head, except where it is covered with diorite rubble from the Mt. Peale intrusion (Richmond, 1962, fig. 36). The head area is complex and has had multiple generations of movement. For example, the northernmost lobe had at least two generations of movement; the front of the youngest bifurcated with one lobe that flowed south down to the main large landslip and another lobe that carried a mass of diorite southeast from the mountain to be deposited as an isolated rubble sheet overlying an earlier debris-flow lobe. Wood buried 1 m in sediment in the most northwestern pond of the main landslip (fig. 11) was dated at only 430 ± 65 C^{14} B.P., yet much of the failure has the appearance of greater antiquity. A large slump block occurs in the head area and is responsible for several ponds there. The block is Burro Canyon overlying Morrison Formation. Directly below, in the most extensive area of beaver ponds, the Morrison, Burro Canyon and Dakota Formations failed together and flowed south-southeast for about 3.5 km. In the area marked by several shear zones on the right (southwest) flank, the mass turned to the southeast and slowed in the Mancos Shale. From the flexure point for the final 3.5 km, the landslip moved down a 9° dipslope and along a cliff outcrop of Burro Canyon and Dakota outcrops on the left (northeast) flank.

Dark Canyon landslips (6) Dark Canyon heads in a large cirque with Mt. Peale to the south and Mt. Mellenthin to the northwest (fig. 4). The two peaks are laccoliths and the canyon itself is eroded in softer sedimentary rock. Morrison Formation and Mancos Shale provide considerable instability, along with several faults associated with the intrusion (Carter and Gualtieri, 1958).

Numerous rock glaciers occur in the area, being derived from the plentiful diorite rubble that is transported from the peaks to the valleys by rock falls and snow avalanches. Internal perennial ice in the interstices between clasts is characteristic and is related to numerous springs associated with the features. The ice presumably is formed by cold air drainage through the rubble, as well as by snow-avalanche deposition. In places where avalanche snow and rock rubble have accumulated extensively on Mancos Shale and Morrison Formation, massive slump-block failures have resulted. Downslope are large flow features, some with partially overlying thick, igneous-rock rubble as the main Dark Canyon landslip.

This Dark Canyon landslip originates on the southeast slopes of Mt. Mellenthin directly at the contact between the laccolith and the Mancos Shale. Downslope the Brushy Basin Member of the Morrison Formation is involved and variable dips are common. The top of the failure is characterized by a series of talus cones. Four major ones average 100 meters in length with 35° slopes characteristic. Water at 5° C in midsummer enters the very top of the slope at the bedrock contact and flows beneath the rock fragments deep into the talus. The igneous bedrock contact beneath the talus slopes 70-80°. The talus cones rest on one large and several smaller slump blocks of Mancos Shale (fig. 13). Some of the rock rubble spills over the crest of the slump block and passes 210 m down to the lowest major slump block. This block, also of Mancos Shale, has a small pond on the



EXPLANATION

Talus and avalanche rubble

Rock glacier or thick (>5 m) rubble deposits

Complex block slide and debris flow

Landslip shear zone

Lake or pond

Southwest margin of landslip-disturbed terrain of southwest Peale-La Sal Pass landslips (4) on Mancos Shale

Uncertain contacts between main southeast Peale landslip-disturbed terrain of southwest Peale-La Sal Pass landslips (4)

Landslip block

FIGURE 11. Map of Southeast Peale landslips. Much of the area adjacent on the southwest (right) flank along both sides of the road is the unclearly defined area of the Southwest Peale - La Sal Pass landslips.

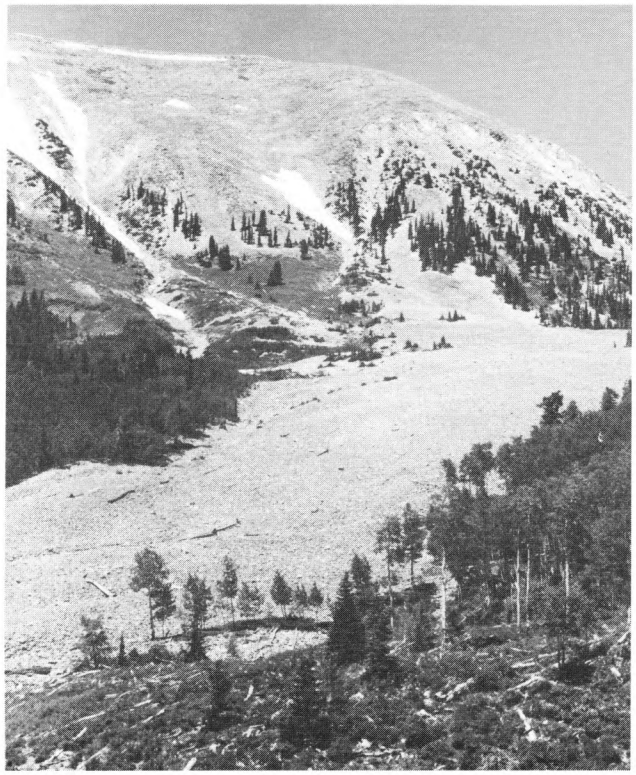


FIGURE 12. Rubble deposit below Mt. Peale at the head of the large Southeast Peale landslip. This rubble deposit was produced by extensive snow avalanches and small rock falls onto unstable Morrison Formation.

left (northeast) flank, and a prominent ridge or landslip levee on the right (southwest). Almost no igneous-rock rubble spills over the crest of the lower slump block, yet the entire right (southwest) lower lobe of this failure is a thick (.5 m) rubble lobe; it is therefore probable that this rubble originated from talus accumulated before the slump blocks occurred and was subsequently transported further downslope by the failure.

The rubble lobe is over 800 m long, several hundred meters wide, and has numerous transverse furrows and ridges with amplitudes of a few meters and wavelengths of 5-10 m (fig. 14). Vertical stones in the furrows and oriented fabric produced by differential flow and shear zones are characteristic and similar to such features in rock glaciers elsewhere (Giardino et al., 1978; 1979). Several springs issue from the rubble with midsummer water temperatures of 0.3-1.0°C, which indicates the presence of internal ice. The rubble lobe is actively moving in different

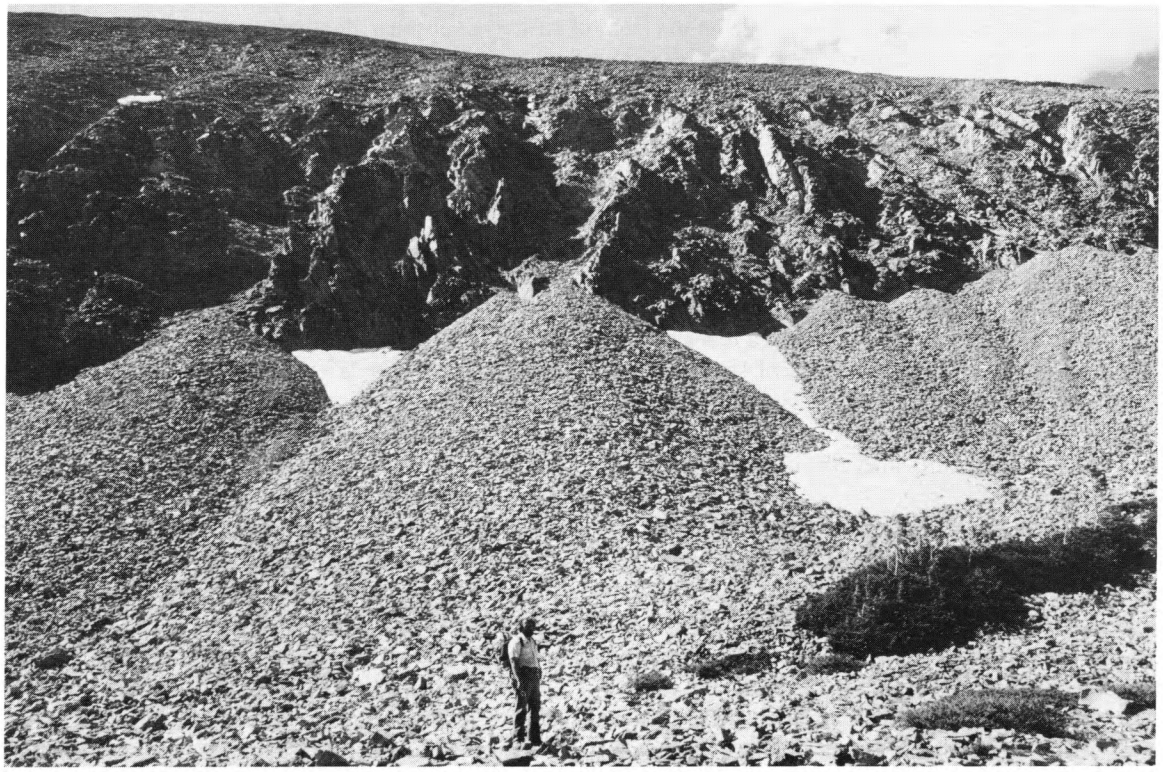


FIGURE 13. Largest talus cone at head of Dark Canyon landslip below Mt. Mellenthin laccolith. The person is standing on the backslope of a slump block of Mancos Shale.

FIGURE 14. Transverse furrows and ridges on lower rubble lobe of Dark Canyon landslip. Water at near 0° C emerges from rubble in middle of foreground vegetation, flows over shale fragments, and back into diorite rubble a few meters below. This occurs at the front of an active subsidiary lobe within the main mass of the landslip. A large rock glacier occurs in the background.

parts, most particularly in its steeper upper half where slopes averaging 20° are characteristic. Shale fragments occur in local zones on the surface. Many clasts are freshly overturned and one large (3.5-m diameter) boulder has been freshly exposed around its base by the dropping away of rubble for 1 m depth around it. The rubble lobe has advanced a few tens of meters beyond the finer-grained parts of the landslip to the northeast. This may be because the rubble provides greater loading on the shale, as well as serving as a locus for development and retention of permafrost and meltwater.

In a few places on the upper right flanks of the rubble lobe, trees are split and tilted by the movement. On the left flank and in the fine-grained and debris part of the main landslip on the northeast, several hundred trees have been affected by the long-term movement. In several places the trees are aligned in rows parallel with the slope, which is a reflection of the numerous linear-shear zones in this area. Such features are characteristic of similar slope failures elsewhere (Shroder, 1978).

Blue Lake landslips (9) Lithology and structure at the head of the Blue Lake slope failure are similar to those of the Dark Canyon landslip; both involve the Mt. Mellenthin laccolith overlying Mancos Shale (Figs. 4, 5 and 15). The Blue Lake feature, however, is almost entirely covered with coarse rock rubble and was mapped as a rock glacier by both Carter and Gualtieri (1958) and Richmond (1962, plate 1, fig. 32). In fact, this deposit is an ideal example of a transition between a "classic" ice-cemented rock glacier and a complex landslip.

The main scarp on Mt. Mellenthin exposes in a few places the baked hornfels of Mancos Shale below the overlying diorite laccolith. Two slump blocks of the shale occur on both the right and left upper flanks (fig. 4). Two main lobes of rubble occur; the upper active one, about 500 m long, and the remaining lower inactive part. Richmond (1962) mapped the upper lobe as lower Gold Basin Formation and divided the lower lobe into upper and lower Beaver Basin Formation, apparently because it is crossed by a band of trees rooted in some surficial fines. In fact, there appears to be no age difference between his two lower parts, and the upper lobe is quite active.

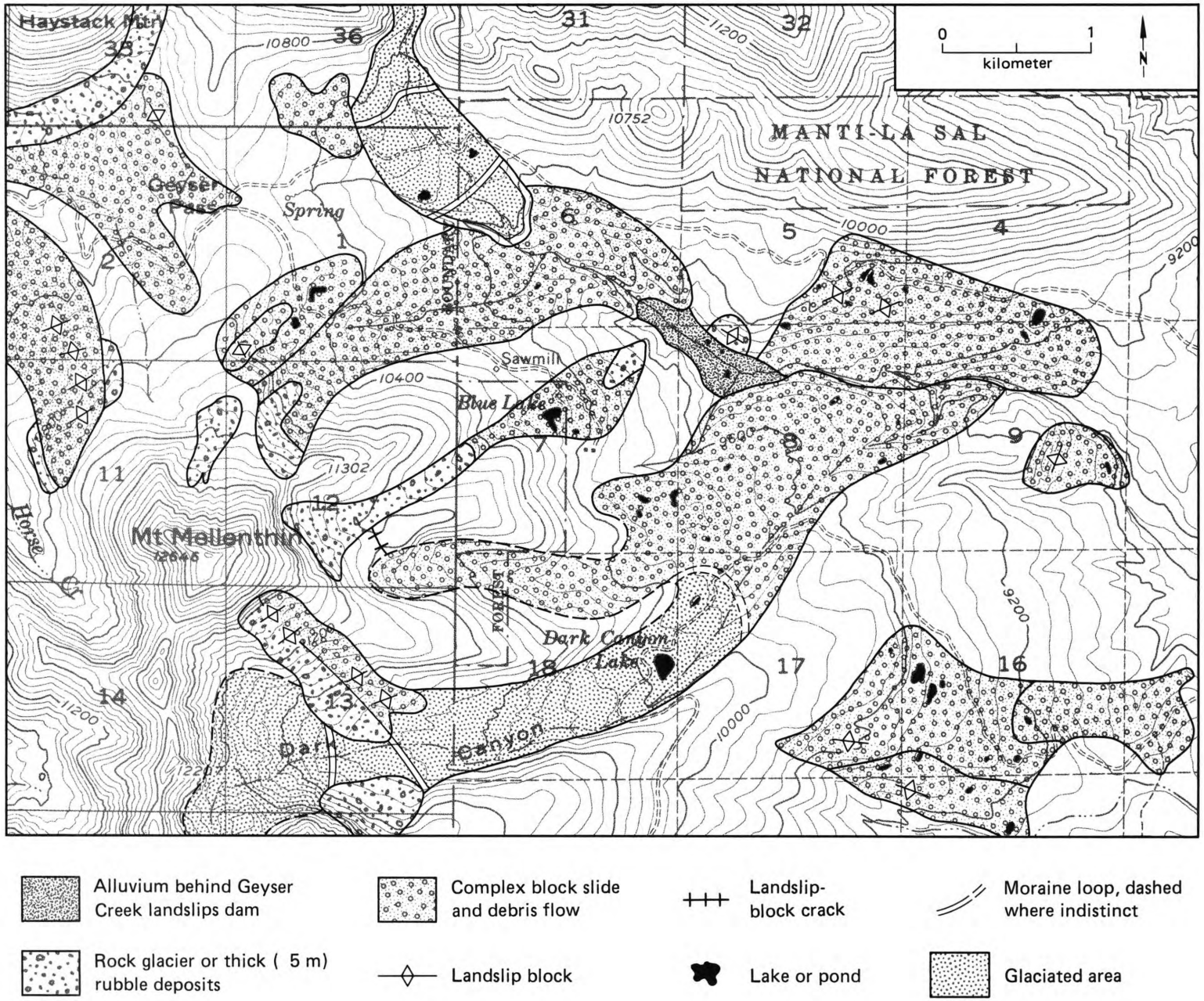


FIGURE 15. Map of landslips of Dark Canyon (6); Hop Creek (7); Geyser Creek (8); Blue Lake (9); East Geyser Pass (10); and the upper parts of landslips of Haystack Mountain - Geyser Pass (11); and Horse Creek - Geyser Pass (12). Compare with figure 4 for locations and with figures 5 and 6 for overlap into peripheral areas.

The upper active lobe is about 500 m wide, has several subsidiary steep slopes at the angle of repose (32-38°) that are up the lower inactive lobe. Between the two occurs a narrow 6-m-deep furrow with a 32° slope on the downhill-facing upper active lobe and a 40° uphill-facing scarp on the lower part. This basal push lobe is about 20 m wide across its top, about 100 m long, 4-7 m high and has shale fragments bulging out from inside in a few places. Nearby, on the right flank, occurs an ephemeral water body that seems to connect with a landslip-block crack developed in the Morrison Formation bedrock of the adjoining ridge. The crack produces a narrow valley much like those developed in the High Plateaus due to unstable substrate (Shroder, 1978).

The lower inactive lobe of this feature measures about 1 km long and about 150 m wide midway down. The front has a sharp angle between the top and frontal slope at the angle of repose and is over 35 m thick. As is also typical of rock glaciers, the upper surface at the front consists of open-matrix boulders, but a few meters below the surface, and exposed in the new road cut directly across the frontal slope, of unsorted coarse and fine clastics.

Below the rock-glacier-like front of the main rubble mass occurs about 1.5 km of hummocky and linear-flow topography in which Blue Lake and several other small ponds occur (fig. 15). Close inspection reveals this mass to be of landslip origin in the extensive Brushy Basin Member of the Morrison Formation. The extreme lower toe of this unit has scattered zones of unvegetated rock fragments that probably represent the original talus accumulations near Mt. Mellenthin that were carried several kilometers downslope by the failure.

Horse Creek-Geyser Pass landslips (12): This failure originated in the Brushy Basin Member of the Morrison and carried with it considerable quantities of overlying Burro Canyon Formation and Dakota Sandstone. Rock of these units dips 15°N in the main scarp, and the landslip moved generally northwest down a component of the dipslope into Lake Oowah Creek valley (fig. 6). A pronounced landslip-block crack or small tension valley opened at the extreme head of the landslip and breaks Horse Creek Ridge from Mt. Mellenthin. Sedimentary rock of the main scarp is only a few hundred meters from the Mt. Mellenthin intrusion, and plentiful igneous rubble from the pluton occurs above the head and both flanks where it was mapped as rock glaciers by Carter and Gualtieri (1958). In fact, these isolated rubble deposits were emplaced in the linear-shear zones oriented along both flanks of the landslip where disruption of the main failure was not as great as in the center between the two deposits. In the central zone occurs a chaotic rock-block-slide terrain of igneous rubble and more than ten large (300-m-long) sandstone blocks that are oriented in a series of closely spaced, near-parallel ridges across the upper slope.

The landslip flowed northwest from the zone of these landslip blocks but, further downslope on the left flank close to the Horse Creek terminal moraines (fig.6), a subsidiary rupture developed that separated Horse Creek Ridge from Boren Mesa and allowed the failure to move directly down a 16° dip to the north. There may have been a meltwater connection from the glacier front into the Brushy Basin Member that contributed to the failure. The resulting northward movement of the mass piled up much debris in hummocky topography concentrated on the lower right flank and toe. This movement partially blocked Lake Oowah Creek valley and produced a small lake in the toe of the landslip (fig. 6).

SNOW AVALANCHES

Analysis of snow avalanches in the La Sal Mountains provides information relevant not only to the people involved in increasing winter usage of the area, but also to foresters and landform specialists. The emergence of helicopter-lift skiing there has produced a new winter revenue for the many river-rafting companies based in Moab or nearby. Local residents also snowmobile in the mountain basins and, although no lives yet have been lost, some concerns have been voiced towards gathering information to avoid future snow-avalanche hazards. The U.S. Forest Service commonly has responsibility for this research elsewhere but has not yet undertaken the task in the La Sals because of relatively limited use of the mountains. Such studies are increasingly common in the Rocky Mountains (Ives et al., 1976), and more needs to be done in the mountains of Utah. One example of lack of information concerns several people who regularly use the La Sals. They wondered about the large masses of snow on which we were working in midsummer. They were surprised to discover that massive snow avalanches occur in an area through which they were accustomed to snowmobile.

Schaerer (1972) noted that the main problem with snow-avalanche-hazard research was not only to determine magnitude and frequency of events, but also to locate the main sites of avalanches. Fohn (1978) showed the difficulty in estimating avalanche frequency and attendant risk in forest sites over long time periods. Our observations of a few large destructive avalanches in forests of the La Sal Mountains in

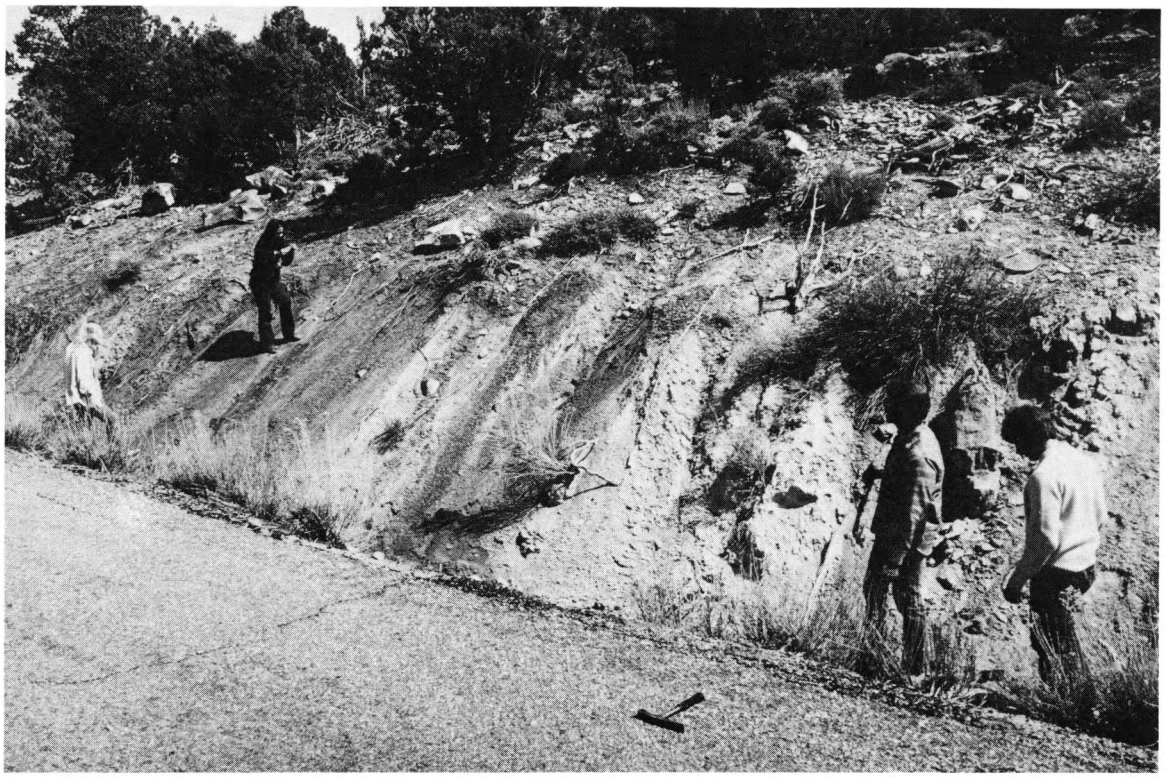


FIGURE 16. Vertically dipping landslip rock-block slide of Morrison Formation at Harpole Mesa. About 100 m upslope is another similar block striking orthogonally to this one.

the high-snow years of 1978-79 and 1979-80 led to this general analysis of magnitude and frequency for such events. At the same time, several years of work on dendrochronology, mass movement, and glaciation of the area (Shroder and Giardino, 1978; Shroder et al., 1980; Shroder and Sewell, 1980) led us to recognize the importance of snow avalanches in the genesis of some landforms (Shroder and Sewell, 1981a). This project therefore was designed to provide only a reconnaissance of a few avalanche zones and associated phenomena; direct observation during winter and dendrochronology could provide significant additional data in the future (Potter, 1969; Shroder et al., 1976; Burroughs and Burroughs, 1976; Shroder, 1980). In this study we concentrated mainly on summarizing the limited information relevant to past and potential effects of snow avalanches on human activity in the area, as well as the geomorphic effects and other phenomena associated with a few extreme snow-avalanche events.

OBSERVED SNOW AVALANCHES

Observations of snow avalanches in the La Sal Mountains have been sparse. The only intermittently inhabited part is in Miners Basin, where some direct observations have been made recently (B. Sherman, written commun., 1980). In spring of 1970 several people on snowmobiles noted two large events, one that flowed north past the base of Miners Basin pond, and another that came west down Snowslide Gulch (fig. 1) to destroy a cook shack built about 1900. In 1980 the first direct records of avalanches were taken largely according to U.S. Forest Service rules (R. Murphy, written commun., 1980). A variety of thick compacted snow-slab and loose-snow, point-release avalanche types were noted; about five of the total 11 observed came down one of several chutes in the Snowslide Gulch area of Mineral Mountain (fig. 1). Although several avalanches were observed elsewhere, the presence of the camp at the base of Snowslide Gulch favors a greater frequency of event observations in that area. In any case, the largest occurred on March 28, 1980, and was loose. General avalanche activity was frequent throughout that month and most took place during storms having strong wind transport so that cornices developed to massive proportions. It is likely that cornice failure is one of the most important producers of extreme avalanches in the La Sals. The largest cornices persist

throughout most of the summer after winters of high snowfall.

The winter of 1981 was marked by low total snowfall and plentiful sun. Resultant sun crusts produced conditions of considerable instability with frequent avalanches after the infrequent, but often heavy, snowfalls occurred (G. Williams, Sidewinder Expeditions, oral commun., 1981).

EXTREME SNOW AVALANCHES

Snow avalanches are exceedingly common in most snowy high mountain areas and, as Mears (1979, p. 1) has noted, only occasional large avalanches may be effectively delineated on maps. In the La Sals we analyzed only extreme events that affected the forest below timberline because above timberline no direct observations have been made and no evidence of past avalanches remains. In general, however, the largest known avalanches occur in the major forested cirques below the highest peaks.

The return interval for large avalanches is of interest, particularly on the longest and widest tracks. Fraser (1966) has shown that four to five centuries commonly occur between extreme events in the Alps, and that high snowfalls usually produce the largest or longest avalanches. Such events tend to clear their path of obstacles that impede flow and therefore can facilitate avalanches in the same areas in later years. On the other hand, if the return interval is long enough, vegetation may become reestablished and eventually retard later movement. As is normal, we observed a great variety of return periods for avalanches of different sizes in the La Sal Mountains (tables 3 and 4). Considering the large number of potential controlling factors — meteorology, terrain, vegetation, aspect, altitude, etc. — it is not surprising that return intervals are irregular. Two long tracks, selected from among many, show the salient information: (1) Mt. Mellenthin - Dark Canyon snow-avalanche track (fig. 1); and (2) Manns Peak - Beaver Basin snow-avalanche track (fig. 1).

We observed the Mt. Mellenthin - Dark Canyon area in the summers of 1977, 1979, and 1980. During the winter of 1976-77 there was so little snow (fig. 2) that either no avalanches occurred there, or they were so small that they melted away quickly. In fact there was so little moisture that we observed no active springs or streams in many places that summer. By contrast, in the winters of 1978-79 and 1979-80, major snow avalanches descended into old forest (greater than 200 yr.) causing considerable destruction of trees (fig. 17).

TABLE 3. *Snow avalanche dates derived from corrasion scars on trees at the lower end of Mt. Mellenthin - Dark Canyon avalanche track.*

ID No.	Species	Snow Avalanche
DC-NC-9	*Picea engelmannii*	1915-16
DC-NC-C	*Picea engelmannii*	1915-16
DC-NC-5	*Picea engelmannii*	1949-50
DC-NC-5b	*Picea engelmannii*	1953-54*
DC-NC-2b	*Picea engelmannii*	1968-69
DC-NC-7	*Picea engelmannii*	1968-69
DC-NC-8	*Picea engelmannii*	1968-69
DC-NC-A(1)	*Picea engelmannii*	1968-69
DC-NC-A(2)	*Picea engelmannii*	1968-69
DC-NC-B	*Picea engelmannii*	1968-69
DC-NC-2a	*Picea engelmannii*	1973-74
DC-NC-6	*Picea engelmannii*	1973-74
DC-NC-4	*Pinus flexilis*	1974-75*

TABLE 4. *Germination-heartwood and corrasion-scar dates from trees at lower end of Manns Peak - Beaver Basin snow avalanche track.*

ID No.	Species	Date
BB-Core 2a	*Pinus flexilis*	1900
BB-Core 2b	*Pinus flexilis*	1913
BB-WC 4	*Picea engelmannii*	1914-15*
BB-Core 3	*Pinus flexilis*	1930
BB-CC 1	*Pinus flexilis*	1940

*Date probably off one year because of false or missing rings.

The month of the 1979-80 avalanche is unknown, but B. Sherman (oral commun., 1980) confirmed that the largest usually occur in spring. Also, two local residents independently noted about 3 m of snow in both Miners Basin and Dark Canyon in midwinter. Because the most recent avalanche directly affected exposed ground in several places and left basal scars on trees only about 1 m above the ground, we conclude that this event was in late winter or early spring after either considerable compaction or melting of the snow, or both. Inasmuch as large trees up to 1 m in diameter were uprooted, a heavy damp-snow avalanche typical of spring conditions seems most likely.

This avalanche path (fig. 5) starts in a col about 3645 m altitude on the south shoulder of Mt. Mellenthin between Dark Canyon and Horse Creek. Track length was about 1 km until the winter of 1978-79, then was extended more than 100 m and the same distance again the following winter. It now measures about 1.3 km in length.

FIGURE 17. Lower part of major snow-avalanche track in Dark Canyon. The area in the foreground and middle was completely forested in 1962 (date of aerial photographs), but several subsequent avalanches have removed the trees and extended the track several hundred meters downslope from here. This area of former forest is the uppermost and youngest moraine in Dark Canyon (see figure 5 for location of moraine crossing avalanche track). The moraine surface has been partially modified by snow avalanche erosion and deposition across it.

Wedge and cross-cut sections taken from trees downed in this most recent track extension show a number of prior avalanche dates (table 3) when snow moved far enough into the forest to debark a few trees and cause corrasion scars. As usual in these studies, the many false and missing rings, so common to callous-margin regrowth over scars, necessitated procedural caution and replication (Shroder, 1978; 1980) to establish accurate and reliable dates.

The Manns Peak - Beaver Basin snow-avalanche track provided only limited section sampling (table 4). Prior to about 1915 the track extended about 1.2 km northeast from Manns Peak (altitude 3741 m). After an extreme snow-avalanche event, probably in the high-snow year of 1915-16 (fig. 2), the track was 600 m further into the valley to an altitude of about 3145 m. This lower track extension is now a jumble of uprooted trees and a thicket of new growth. The wide variety of dates obtained from tree heartwood in this area shows the imprecision implicit in using germination date as an indicator of avalanche time. Young trees may be either overridden or germinate long after the event. On the other hand, the corrasion-scar date (1914-16) is far more precise, both from the context in which the tree was extensively scarred at an abrupt steepening of slope, and in several replicating counts made on the sample. It is probable that a false ring occurs, however.

Comparison of a limited sample of extreme snow-avalanche dates (tables 3 and 4) with available snowfall information (fig. 2) reveals close correspondence between large avalanches and years of high snowfall, as one might expect. This is in agreement with the observation of Schaerer (1972, p. 218) that avalanche frequency (and presumably also magnitude) is determined more by climate, including snowfall amount, than by terrain characteristics. Extreme avalanches are defined loosely herein as those with greatly extended runout zones, but determination of exact controls of runout-zone length is a problem, and other studies have shown the feature to be controlled by di-

verse phenomena (Bovis and Mears, 1976; Butler 1979; and many others). No simple understanding of runout-zone length has emerged, probably because of the large number of potential controlling variables; the predictability of both return interval and runout-zone length based on our limited study in the La Sals is thus not possible.

GEOMORPHIC EFFECTS OF SNOW AVALANCHES

Erosion and deposition produced by snow avalanches are generally difficult to analyze because of: (1) the lack of quantitative information about return intervals; (2) the problems related to accurate measurement of debris transport per event; and (3) the common variety of other processes that contribute to the various deposits at the bottom of avalanche tracks (Rapp, 1959; White, 1968). Peev (1966, p. 358-359) specifically recommended stratigraphic studies, as well as short-term quantitative analyses of seasonal movement of debris, in order to obtain a more complete picture of snow avalanche erosion and deposition.

Numerous excellent analyses of seasonal movement have been conducted elsewhere (Potter, 1969; Luckman, 1971; Gray, 1973; and many others), but most techniques for measuring of short-term debris movements were judged by us to be inappropriate because they are potentially too unrepresentative in the context of probable long return intervals of avalanches in the La Sals. Nevertheless, a few observations of such phenomena, as well as of long-term deposition and some landforms, are warranted because several appear unusual. These features are: (1) smooth hyperbolic- or "U"-shaped bedrock chutes; (2) plentiful debris erosion and transport in roots of uprooted trees; and (3) the effects of snow load and associated meltwater on rock-glacier movement.

Bedrock chutes: Furrows, gullies, or chutes eroded into slopes by snow avalanches are ubiquitous in alpine areas and have been described by a variety of workers (Matthes, 1938; Davis, 1962; Peev, 1966). In most cases, however, it has not proved possible to differentiate adequately between the effects of snow avalanches, static nivation, stream flow, or various other types of mass movement such as rock fall, torrential-rain-induced debris avalanches, or the like. In general such discrimination in the La Sal Mountains is not possible either, except in a few locales high on peaks where an unusual configuration of chutes shows their erosive control by joints, freeze/thaw, and snow avalanches only.

These chutes seem to develop best on the upper parts of strongly jointed, homogeneous diorite intrusives of the Middle Mountain Group. Several well-developed sets occur on the north peak in the Mt. Tukuhnikivatz complex (fig. 1). Active freeze/thaw in the strongly jointed intrusive produces blocks generally about 25-50 cm in diameter and presumably provides plentiful rubble for removal by snow avalanches.

Mt. Peale has four large chutes and several smaller ones that pass snow avalanches and rock fragments out over a 200-m-high cliff and down onto the largest of the Dark Canyon rock glaciers (fig. 5). These chutes tend to have smooth straight floors, a wide hyperbolic- or "U"-shaped cross section, and are separated by narrow knife-edged ridges 1-10 m high. Four chutes measured have rectilinear slopes of 35°, 37°, 37°, and 38°.These are almost exactly like those described in the Sierra Nevada (Matthes, 1938; Davis, 1962), but missing from so many other areas. In the La Sal examples the broad hyperbolic shape is less pronounced or missing in heterogeneous rocks or in chutes where snow exists throughout summer; variations in lithology and structure, along with nival and fluvial processes, possibly reduce smoothness.

Another set of such chutes on the igneous rocks of a peak adjoining Mt. Tukuhnikivatz conduct snow avalanches down into a narrow "V"-shaped ravine onto non-igneous rocks in Gold Basin (fig. 1). The irregular shape of this ravine, in contrast to the avalanche chutes above, suggests the influence of variable lithology and structure, together with erosion by running water rather than that of the more uniform and spreading flow of snow avalanches.

Although an oversimplification of rheology, some snow avalanches in the La Sals may act as pure Bingham substances in which the snow has some initial strength until shear stress is sufficient to cause failure. After that the material will flow viscously as long as the initial shear strength is exceeded. Johnson (1970, p. 537-571) has shown how the flow of a Bingham substance will convert a channel of "V"-shape or triangular cross section into one with a "U"-shape because low shear stress and lower flow velocities occur at the apex of the "V". Velocity, and thus also erosion, are greatest midway up the sides of the channel, and a "U"-shape will therefore be produced in a channel medium that is homogeneous or uniformly heterogeneous. M. Bovis (written commun., 1981) has stated that midwinter avalanches are fully turbulent and that viscous effects would therefore be negligible at that time of year, but this is probably not relevant because we suspect greater erosion of exposed bedrock by heavy damp-snow avalanches in spring. Also snow avalanches tend to override old

snow in gullies and thus will erode laterally in their track to produce a wider "U"-shape.

Dark Canyon erosion and deposition The large snow-avalanche track in Dark Canyon (fig. 5), whose main source area is the col above Horse Creek basin, has two prominent erosion and deposition features associated with it. On its upper slopes an unusual debris-burst cavity was produced in the extreme avalanche of 1979-80, and further down considerable forest was extensively uprooted so that large quantities of rock and earth were transported with the roots and deposited on the 7° slopes.

The debris-burst cavity resulted when the avalanche ran against an uphill-facing stream bank (fig. 18). This bank slopes at an angle of 21° and we presume it must have been at least partly exposed to receive the impact. A meter or two from the downhill edge of the stream bank a cavity was produced in the 12° slope by bursting of turf, earth, and debris to a depth of about 50 cm. The cavity measured 6.5 m long, 3.5 m wide, and had a 5-m-long crack exposed uphill from it. Turf blocks, recognizable as coming from the cavity, were scattered down the hill, with the distance of least transport being 173 m and the greatest 213 m. Judging from the height of removal of freshly detached branches on nearby standing trees, the avalanche was about 4 m deep when slightly uphill from the debris-burst cavity and about 6 m deep after it passed beyond the lower edge. Scour or impact pits have been reported in the Sierra Nevada, in Asia, and in Norway (Davis, 1962; Peev, 1966; Corner, 1980), but they appear much larger, smoother, have raised rims downstream, and seem mainly the result of erosion at a strong slope concavity below a steep slope. The debris-burst cavity described herein occurs on the crest of a slope convexity and is irregular. In any case, none of these various erosion features seem to be particularly common, and their formation probably requires rather special circumstances.

On the lower part of the track, the uprooted trees left behind pits each about 1 m deep. Older examples of this phenomenon there show that as the trees decay the root-bound debris leaves a mound. The combined effect produces a hummocky topography of a few meters maximum relief. This topography is common on many of the lower tracks of snow avalanches in the La Sals and thus shows a long history of this activity.

The extreme avalanche of 1979-80 in Dark Canyon moved about 25 large trees, most of which measured 0.5-1 m in diameter. Each tree had about 1.5 m^3 of debris bound in the roots, which indicates

FIGURE 18. Snow-avalanche chutes on north face of Mt. Peale.

about 35-40 m^3 of debris was moved up to several hundred meters downslope by the one event. In time this debris is likely to compact to at least half its present thickness, which will result in a discontinuous or hummocky layer probably 25-50 cm thick.

Debris entrainment in the roots of trees uprooted by snow avalanches is an obvious and active form of erosion and deposition by extreme events in the La Sal Mountains, but its past history is unknown. On the other hand, this mechanism may have been more prevalent here than in many other mountain ranges for several reasons: (1) snow depths tend to be low, thus allowing trees to be more easily uprooted than snapped off; (2) the climatic regime is not harsh, which allows trees at high altitude to attain large size and therefore greater resistance to breakage; and (3) recent high snowfalls have produced several extreme-event avalanches, which tend to be of large size and therefore do a great deal of erosive work in a short period of time. Over the long run, however, the smaller annual avalanches and other processes probably also make a significant contribution.

McCORMICK PARK DEPOSITION

The west slope of Pilot Mountain in the North Mountain Group has a prominent avalanche track that extends for 1-2 km down valley (fig. 1). On the north side of the track at an altitude about 3290 m, near the Alamo prospect (Hunt, 1958, plate 40), a new mine-road cut has exposed a 20-m-long organic layer 80-90 cm below the surface. At the present time the site is covered with a heavy growth of large Engelmann spruce (*Picea engelmannii*) that are one to two centuries old, based on diameter comparison with other nearby cored trees of the same species. Some have been knocked over recently by avalanche activity but, as a whole the site does not appear to have had a major disruptive avalanche since the time of burial of the organic horizon because the present generation of trees is rooted in the thin soil above it. Material above and below this horizon is unsorted colluvial rubble and fine clastics that we interpret from the topographic and stratigraphic context to be avalanche deposition. Radiocarbon dating of wood and soil gave figures of 1145 ± 55 and 1775 ± 155 C^{14} B.P. The disparity in dates, which were carefully double-checked during lab analysis, is probably coupled to differences between the original tree age and soil organic matter of living or reworked material. These dates provide estimates of 50-70 cm/millenium deposition rate; a figure commensurate with episodic long-return-interval activity as envisioned by us.

Assuming this is a similar deposition rate at exactly the same altitude as that on the Mt. Mellenthin - Dark Canyon avalanche path, described previously as 25-50 cm/event, this works out to one or two major deposition events per millenium. Thus a return interval of major avalanche events here would be only once in 500 years. This is not a high rate, but is in agreement with most long-term direct observations of such phenomena (Fraser, 1966). For example, our analysis of the work of Madsen and Currey (1979) on the Snowbird avalanche site in the Wasatch Mountains of Utah shows an average deposition rate of about 39 cm/millenium for the last 12,000 years and a minimum of seven major forests probably destroyed by avalanches during that time.

SNOW-AVALANCHE EFFECTS ON ROCK GLACIERS AND LANDSLIPS

Rock glaciers and landslips in the La Sal Mountains appear to be polygenetic, mechanically interrelated in some places, and in many cases they move intermittently. In addition we suspect that the intermittent activity may be out of phase from one feature to another, at least in nonglacial times as is now the case. All of the active rock glaciers there are ice-cored or ice-cemented diorite rubble with interstitial fine clastics (Shroder and Giardino, 1978; Shroder et al., 1980; Shroder and Sewell, 1980). Over half in the La Sals have accumulated on unstable Morrison Formation and Mancos Shale, and have subsequently loaded both to failure. We suspect that water, much of it supplied by melting avalanche snow, becomes trapped beneath the ice of the rock glacier or landslip, increases hydrostatic head, and thus facilitates movement. Delivery of snow via avalanching onto any slope has the same effect as a local increase in precipitation. Thus if a regional increase in precipitation produces increased slope movement, as in the High Plateaus of Utah (Shroder, 1978), an increase of avalanche snow on similar lithology and landforms in the La Sals is likely to do the same. Many of the rock-glacier-like features in the La Sals that receive plentiful avalanche snow at their heads, like the Blue Lake example (fig. 5), have large slump blocks at their heads and grade downslope into massive landslips on unstable bedrock. Steep fronts at the angle of repose also occur where basal-shear resistance is sufficient to produce the characteristic oversteepened front (Warnhaftig and Cox, 1959); elsewhere basal push lobes develop or the fronts are more gently sloping where basal slip of the landslip type is more active.

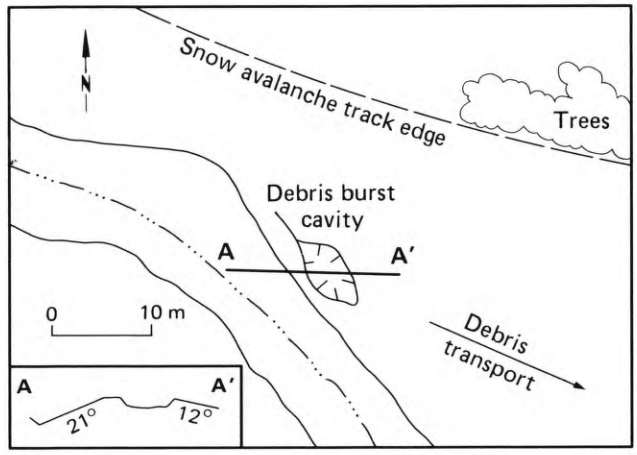


FIGURE 19. Debris-burst cavity in plan and profile, Mt. Mellenthin - Dark Canyon snow avalanche of winter 1979-80.

The role of water in landslips is well known, and a similar relationship in rock glaciers is becoming increasingly apparent (Barsch and Hell, 1975; Shroder, 1971, 1978; Johnson, 1978; Haeberli et al., 1979). Thus the relation of snow avalanches to rock glaciers

is significant in a number of ways: (1) additions of rock debris and ice mass, produce some of the characteristic internal structures of boulders, ice cement, and frozen matrix; (2) increased load on the rock glacier overcomes resistance to shear; and (3) provision of spring and summer meltwater also decreases shear resistance. These factors are all apparently involved in the recently renewed movement of the large rock glacier below Mt. Peale in Dark Canyon that is discussed below (figs. 5 and 20).

This rock glacier occurs at an altitude of 3185-3320 m and measures about 800 m wide, 400 m long, and 30-40 m thick. It has three main lobes, each with a depression (easternmost depression not visible on fig. 5) at the head of the rock glacier below the talus cones at the cliff base. These depressions may be former ice-core-meltout zones that now receive most of the snow avalanches, although in years of extreme events such as 1979-80 some of the avalanches traveled as far as the oversteepened front. The westernmost depression has some Engelmann spruce, the oldest of which germinated before 1750 (D.R. Currey, written commun., 1979). The avalanche chutes shown in fig. 18 occur directly above this westernmost lobe and depression. The central depression marks a shear zone developed within the rock glacier as a result of differential movement caused by greater rock debris and snow avalanches down the east side of Mt. Peale. Talus above this depression is marked by a rapid-debris-flow channel with small levees produced prior to 1962, probably by torrential rain. The depression has considerable fine clastics, turf, and evidence of occasional ponded water.

Water was heard running at many places inside and was seen issuing, at temperatures near 0°C, from beneath the front as various springs. In places in Gold Basin and Dark Canyon, liquid water temperature was measured and found to be as much as -0.7°C. White (1971, p. 50) and others have also observed water at subfreezing temperatures in rock glaciers elsewhere. Dissolved salts have been suspected, but release of water under pressure from beneath the ice may be more likely.

Basal-push lobes of rock fragments about 2-3 m high occur toward the east of the Mt. Peale rock glacier and small turf rolls about 10-25 cm high also are located in direct contact with rocks at the base of these (figs. 5 and 20). Both features are interpreted as the result of differential basal slip, although some of the turf rolls could be the result of loading of fine

FIGURE 20. Front of Mt. Peale rock glacier, July, 1980. Basal-push lobe of rock fragments and lower turf rolls are overriding logs.

clastics. A rotten log, overridden for about 25 cm by a basal push lobe (fig. 20), was dated at 690 ± 80 C^{14} B.P., and indicates that advance in the last few centuries has been rather slow.

In 1979 we noted that the front of the rock glacier was at the angle of repose and somewhat unstable but could be climbed in several places without hazard. In 1980, after the extreme snow-avalanche activity of the previous winter, almost all of the front of the rock glacier became re-activated. Large quantities of newly broken and overturned rock fragments occurred all along the front except in the notch between the westernmost lobe and the other two. Some of these rock fragments may have been eroded from the upper cliffs or rock-glacier top by avalanches that passed to the front, but the newly oversteepened front, which had fallen in several places, and freshly overturned earth at the base also indicated renewed movement. Inasmuch as the 1980 snow persisted into the fall, we speculate that avalanche meltwater contributed to the renewed movement of the rock glacier through the summer. Advance at the base could not be accurately measured, but probably was less than a few centimeters in aggregate. Forward movement seemed to occur in an irregular fashion across the front, both by collapse of the oversteepened top and by basal slip.

CONCLUSION

Analysis of mass movement in the La Sal Mountains requires differentiation of diamictons produced by glaciation or other processes. Unsorted, unstratified, coarse-clastic deposits are best assigned genetic attributes best only where substantiating data are available. Sedimentologic data alone have been shown by others to be equivocal with the result that we rely primarily on geomorphology for differentiation. In some cases the landforms are so similar that such discrimination is not possible. In this first detailed mapping of landslips in the La Sal Mountains, glacial deposits have been excluded from the central analysis because other comprehensive studies with accepted new methods must be performed on them in the future. Nevertheless, our reconnaissance suggests only a Bull Lake - Pinedale sequence with no unequivocal evidence for pre-Bull Lake glaciation and only minor post-Pinedale glacial activity.

On the other hand, development of rock glaciers in post-Pinedale time has been active, although the usual ideas of rock-glacier activity resulting from action solely of internal ice apply only in a few cases. Instead, many of the features result from accumulation of plentiful igneous-rock rubble on unstable Morrison Formation or Mancos Shale that results in variable landslip activity. Interstitial ice in the rubble interstices probably provides an impermeable barrier to upward movement of groundwater, which further aids in the movement of these rock glacier/landslips.

We identified 20 major areas of landslips, all of which involve the Brushy Basin Member of the Morrison Formation, or the Mancos Shale, and most of which were mapped previously as glacial moraine. Other large slope failures occur in the area, but an exhaustive study would be impractical at present. Landslips appear to be of all ages in late Pleistocene and Holocene, with retrogressive main scarps and reactivation in old failure zones being common. Many major movements are likely to have begun as permafrost was increasing or decreasing in conjunction with prior glaciations, because of the attendant groundwater fluctuations. The influence of water on mass movement is well known and precipitation changes coupled with long-term temperature fluctuations are likely to have had considerable effect on large slope failures. At the present time, most large landslips are inactive and no major activity is probable in the immediate future. Nevertheless, widespread forest clear cutting or mining, major road building, or increased water-diversion projects could have some effect in specific places where unstable bedrock and steep slope coincide.

Snow-avalanche activity in the La Sal Mountains is relatively unstudied, but our reconnaissance shows need for additional information relative to increased wintertime land use of the area. Snow avalanches are also significant geomorphic agents in the area because they produce considerable erosion and deposition of debris, much of which later becomes involved in landslip or rock-glacier activity. Perhaps most significantly, the deposition of snow by avalanching on certain slopes produces a local, but major increase in slope water and consequent renewed movement of rock glaciers or landslips.

Much work remains to be done on surficial deposits in the La Sal Mountains. The establishment of a new, detailed chronology awaits refinement of the glacial sequence and integration of that with the mass-movement events. The whole spectrum of new relative-age-dating techniques (Burke and Birkeland, 1979; Miller, 1979) coupled with ample radiocarbon dating and dendrochronology will be necessary to provide temporal control. Such work will add considerably to our understanding of geologically recent changes in eastern Utah, and will thus better enable

prediction of effects of either future climate shifts or changing human land use in the La Sal Mountains.

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