Chapter 2

Introduction to Cuesta Landforms and Sapping Processes on the Colorado Plateau

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INTRODUCTION

Terrestrial channels are commonly composite channels, formed jointly by runoff and groundwater sapping processes. Schumm and Phillips (1986) describe similarities between composite channels along the New Zealand coast and valley networks on Mars. The objective of this book and much of the recent research related to groundwater sapping processes is to determine what geomorphic criteria can be used to distinguish between channels formed dominantly by runoff and those which are sapping dominated. Kochel and Piper (1986) showed, for example, that morphometric parameters describing Hawaiian valleys could distinguish between these two channel types. These remote observations have been supported by field observations in Hawaii by Howard and Kochel in 1985.

The landscape, lithology, and structure of the Colorado Plateau have similarities to areas of Mars, leading geomorphologists to examine the region for possible analogs to such features as volcanoes, wind-deposited and eroded landscapes, and fluvial landforms. In particular, Ingersoll and Malin (1983) have suggested that groundwater sapping processes have been instrumental in sculpting certain valley networks in sandstone. Features of these valleys, such as theater head deposits, nearly constant valley width downstream, short, stubby tributaries, steep valley headwalls, and strong control of size and orientation by rock structure are consistent with the observation of seepage-related erosion localized at headwalls. The valleys display a morphological similarity to some martian valley networks, leading to the suggestion that similar processes led to their development. The Ingersoll-Malin study provided the focal point and null hypothesis model for the field conference described here, which addressed several themes related to the role of groundwater in creating the landscape of the Colorado Plateau.

1. The major conference theme was an evaluation of the relative roles of groundwater sapping and runoff processes in erosion of layered rocks in the Colorado Plateau, and the effects of sapping on valley morphology.

2. Almost all weathering and mass-wasting processes occurring on the Colorado Plateau involve some penetration of water into exposed rocks to produce the physical and chemical changes leading to rock breakup and transport. Thus confusion and semantic controversy can occur until a consensus definition of sapping and seepage erosion emerges.

3. During the field excursion, a wide range of weathering and erosional features were visited that may be martian analogs. Therefore, the field guide discusses the major processes and landforms occurring on the layered rocks of the Colorado Plateau, excluding volcanic, alpine, glacial,olian, and large-scale structural features. The morphology of sandstone escarpments is emphasized.

4. Interpretation of landforms is complicated to the degree that present features have resulted from processes different from those presently acting. Several authors have suggested that certain landscape elements may be relics from Pleistocene fluvial climates. In a similar vein, we discuss the variation of scarp morphology with arid variations in the present climate.

5. We hope that the discussions held during the field trip and the present document will suggest field or experimental research that can answer questions regarding sapping processes and landforms.
STRATIGRAPHY AND STRUCTURE

The Colorado Plateau is underlaid by layer-cake sequences of Paleozoic and Mesozoic sediments, primarily shale and sandstone, with some formations or interbeds of limestone, gypsum, or more soluble evaporites. The beds are generally nearly flatlying, but are locallyfaulted or warped into monoclines, domes, and basins. Largely the sediments are intruded by Tertiary laccoliths, plugs, sills, and dikes or capped by lava flows and volcanic cones. Near Moab, large-scale faulted anticlines have been created by flow and intrusion of evaporites and their solution at depth (Doelling, 1985). General geologic references for the area include Rhyol (1976, 1977), Hinzen (undated), Chorin (1983), and Baars (1983). Geologic maps at a scale of 1:25,000 have been published by the U.S. Geological Survey as follows: Moab Quadrangle, Misc. Invest. Ser., I-160; Grand Junction, I-736; Cortez, I-629; Shiprock, I-345; Gallup, I-881; Salina, I-591; Escalante, I-744; Marline Canmore, I-1003.

Sandstones

The rock types of primary interest during the field conference were sandstones and interbedded shales; particularly the massive sandstones of the Navajo Formation and the slickrock member of the Entrada Sandstone. The sandstones generally contain more than 80% by weight of quartz sand, with variable percentages of orthoclase or clay, and generally weak cementation by hematite, calcite, or, less commonly, silica. Even the massive sandstones contain occasional thin beds or partings of shale, chert, and limestone. For more detail on the sedimentology of these units and relationships to groundwater flow, see chapters 3 and 4. Similarly, some layers have considerably greater or much less than average amounts of cementing hematite or calcite. Schumm and Chorley (1966) give unfractured compressive strengths in the range of 2500 to 7200 psi, which are in the low range for sandstones, but strong enough to permit massive vertical cliffs. Porosity commonly ranges from 20 to 35%, and permeability ranges from about 2 x 10-10 to 1 darcy, with the Entrada near the lower end and the Navajo near the upper end of this range (John, 1962; Cooley et al., 1969). Such permeability is at the high end of the range cited by Freeze and Cherry (1979, p. 28) for sandstones and is equivalent to that of unconsolidated silty sand. Curiously, Schumm and Chorley (1966, p. 133) considered the sandstones to be of low permeability while at the same time conducting experiments that demonstrated the ability of water to penetrate 1/2 to 3 inches into Colorado Plateau sandstones within 20 minutes after the beginning of surface wetting (p. 24). Due to high porosity, permeability, and weak cementation, the sandstones weather easily by several mechanisms, including grains-by-grain surface removal, separation along bedding planes, and crumbling. Schumm and Chorley (1966) subjected samples of representative sandstones to two winters and one summer of precipitation and freeze thaw. The samples typically lost 1 to 3% of their initial volume in this exposure, with a high of 23% for Entrada Sandstone samples. This indicates a potential for rapid surficial weathering and erosion under present climates. Schumm and Chorley also note the presence of loosened weathered sediment on bare rock slopes following snow melt and freeze thaw, with slopes on Entrada Sandstone becoming slippery. Wind and rain rapidly remove such accumulations. Variable amounts of weathering have occurred in mass within the sandstones following lithification and prior to their exposure, both by deeply circulating groundwater and by circulation related to present-day topography (Corley et al., 1969; Hamilton, 1984). Corley et al. (1969) suggest that permeability (and, by implication, weatherability) of sandstones increases with the length of time the rocks are exposed near the surface (that is, beneath other erosional surfaces).

Shales

Shales of the Colorado Plateau range from massive marine shales, such as the Mancos Shale, of dominantly silt and illitic clay composition, to sandy lacustrine shale with montmorillonitic clays, such as shales in the Morrison Formation. All of these shales weather readily upon surface exposure; with rock break up aided by shrink-swell due to changes of moisture content. The montmorillonitic shales break down rapidly as a result of simple wetting, with remarkable volume increases. The illitic sand weather more slowly, yielding yellowish lagoons of reduced salts that must be carried away in solution to continue the weathering process. Despite its weatherability, solid shale is generally encountered within a few inches of the surface on badlands slopes due to the impermeability and self-sealing characterics of the weathered shale and the restricted supply of precipitation.

MORPHOLOGY AND EROSIONAL PROCESSES ON SANDSTONE CUESTAS

Sandstone units on the Colorado Plateau are generally exposed as bare rock slopes except where mantled with eolian sands or talus. However, areas of very low relief, such as the tops, or back-slopes, of gently dipping cuestas of sandstones where overlying shales have been stripped away, may be mantled with sandy, cobly, poorly horizontal soils and crumbly vegetation. Two morphological end members characterize the sandstone exposures: low to moderate relief “slickrock” slopes and cliff or escarp
slopes develop where the sandstones are being undermined. Figure 1 shows terminology used here for the form elements of an escarpment. These two slope types differ significantly in morphology as well as type of weathering and erosional processes and are discussed separately. An intermediate landform type, termed "segmented cliffs" by Oberlander (1977), is also discussed.

**Slickrock Slopes**

The most striking and unusual landform type occurring on desert sandstone exposures is a long, generally rolling relief on bare rock slopes (figs. 2 and 3). Hill forms are generally convex to convex-concave and rather irregular due to the prevalence of small-scale structural and lithologic controls exerted by the exposed rock on weathering and erosional processes. Slickrock slopes occur most commonly on the tops and crests of sandstone cuestas, being exposed by erosion of weak overlying shales. Erosion rates on these cuesta backslopes are relatively small compared to the rate of sandstone erosion on the scarp faces (Howard, 1970), with correspondingly low relief. However, slickrock topography of moderate and locally high relief occurs on thick sandstone units where long exposure and less effective scarp backwasting contribute to the development of steeper relief. The thick Navajo Sandstone best exhibits this higher relief form of slickrock slopes, such as at Zion National Park (fig. 4). Other descriptive terms applied to such topography include "beehives," "haystacks," "wheelbacks," "slipfaces," "bathtubs," and "cornstalks" emphasizing bedding exposure and shearing fractures.

Where strong structural control by jointing or faulting occurs, the fractures tend to be eroded into tunnels or valleys, and the sandstone landscape takes on a reticulated or "maze"-like appearance (fig. 3). Doelling (1985) notes that sandy colliulmum collecting along depressions developed on joints accelerates weathering of the sandstone by providing a moist environment; thus the influence of fractures on the topography is enhanced by positive feedback on weathering rates. Small-scale horst-and-graben development associated with extensional movement has created the "needles" section of Canyonlands National Park (McGill and Stromquist, 1975).

Slickrock slopes are "weathering-limited" (Carson and Kirkby, 1972, p. 104-106) in that transport processes are potentially more rapid than weathering processes. That is, loose debris is removed from the slopes as fast as it is produced by weathering so that little or no loose residuum covers the bedrock. This contrasts with "transport-limited" slopes with a residual soil in which the rate of erosion is limited by the rate at which slope wash or mass movement can move the soil. Many badland slopes on shales are transport limited despite their very thin residuum.

On slickrock slopes the bedding is emphasized by the grain-by-grain loosening or disintegration of thin surface crusts or whole layers of the sandstone exposed on these weathering-limited slopes (figs. 5 and 6), particularly on exposures of the massively cross-bedded Navajo Sandstone (fig. 7). Coarser sand layers, with fewer grain-to-grain contacts weather and loosen most readily, aiding differential surface expression of minor lithologic variations (Hamilton, 1984). Despite these microscopic lithologic controls, the slickrock slopes generally show only minor form control by
bedding, and the fairly planar to rounded slopes cut across bedding planes (figs. 8 and 9). This propensity for smooth, rounded slopes on a weathering-limited landform is noteworthy and in need of explanation. By contrast, the lateral and downslope grading of transport-limited slopes is well known and readily explained as a result of a direct relationship between slope gradient and the rate of downslope mass movement.

One reason for the development of smooth, generally convex slopes is the development of exfoliation or shearing fractures (figs. 10 and 11) in massive,
Figure 4. Steep slickrock slopes in Navajo Sandstone along Utah Highway 9 west of Deadhorse, Utah.  

Figure 5. Slickrock surfaces in Navajo Sandstone along Utah Highway 12 west of Bradley. Left: Black macro-text rock fragments are 3/16 inch in length. Note the scalloped nature of the surface. The eroded surface suggests the former presence of bodies on the flaky surface.  

Figure 6. Disintegration of surface layers at Navajo Sandstone on slickrock surface near Utah Highway 12 west of Bradley, Utah. The large disintegrating pieces are about 50 cm in size.  

Figure 7. Expression of bedding in slickrock slopes in Navajo Sandstone near Utah Highway 12 west of Bradley, Utah.  

Figure 8. Slickrock slopes in Navajo Sandstone in Zion National Park, Utah, showing smooth scaling of slopes around bedrock.  

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poorly jointed (referring here to preexisting regional or systematic jointing) sandstones such as the Navajo Sandstone (Bradley, 1983). The exfoliation is primarily due to stress-relief fracturing (Bradley, 1963) which produces lenticular sheets of thicknesses ranging from an inch to a few feet, with more widely spaced fractures below these, fading out within 10 to 20 m from the surface. Thus the exfoliation joints form "a crude, somewhat subdued replica of the surface form" (Bradley, 1963, p. 521). The tendency of exfoliation jointing to encourage development of a broadly rounded, domal topography is well known, particularly in granite (fig. 12). However, many of the granite domes, obvious peeling of exfoliation layers (figs. 13 and 14) is the exception rather than the rule on slick-rock topography, where grain-by-grain weathering or peeling of thin (<1 cm) weathered rinds seems to be dominant. Thus, although exfoliation joints are common, their control of surface erosion at present seems generally slight; however, they may exert subtle, long-term effects on weathering rates. Control by exfoliation joints may have been more pervasive during pluvial episodes, when deep freezes-thaw cycling may have caused separation and breakup of shearing layers. If this is true, the dome-like large-scale hill forms may be inherited from the Pleistocene.

Locally, planar slopes on sandstone may have resulted from exposure of sandstone formerly mantled by talus from overlying cliffs, as discussed below. However, most slickrock slopes lack the association with overlying cliffs and are either too gentle or too steep to have been talus slopes.

Other weathering and erosional processes may aid the development of planar slopes. Strictly surface erosional attack generally does not produce planar surfaces. For example, surface solution of limestones results in irregular surfaces with a tendency toward pointed projections and concave depressions (such surface solution has been termed "uniform denudation" by Lange, 1959) as well as an intricate expression of bedrock solubility differences (fig. 15 shows bottom denudation applied to a contact of scarp planar forms, showing pointed outliers and smooth, concave

![](Image)

**Figure 12.** Diagrammatic sketch of exfoliation joints in massive Colorado Plateau sandstones. Arrows show inferred directions of expansion. (A) An exfoliation dome. (B) An exfoliation core. (C) An overhanging exfoliation plate in a meander-like figure (figure and caption from Bradley, 1963).
Figure 13. Shale beds near Step 1 at the Inscription House Area (figs. 45 and 61) showing control of weathering and erosion by shearing fractures. The rounded exposures of Sasso Sandstone with strongly dipping bedding planes are sharply cut at their base by a gently dipping shearing fracture. The beds below the shearing fracture (where figures are walking) evidently are less subject to weathering than the overlying rock, perhaps due to less influx of water along fractures.

Figure 14. Shale beds near Step 1 at the Inscription House Area (figs. 45 and 61) showing control of weathering and erosion by shearing fractures that cut across the bedding planes (see fig. 45) for further explanation.
reentrant). However, any weathering process that acts through some depth from the surface will tend to eode away projecting masses due to the greater surface area relative to volume. Such processes may include solution of cement, weathering of feldspars and clays, and disruption of the rock along microfractures and between grains due to differential volume changes produced by temperature changes, freezethaw, or shrink-swell of clays. Since such weathering is necessary below surface, gravity removal can occur, a "grafting" of surface slopes should result, with a characteristic scale of action of the same order of magnitude as the depth of weathering (presumably a few centimeters to a few meters). The occurrence of such weathering processes is evidenced by the development of shallow fracture networks on many sandstone exposures. In massive sandstones these surface cracks create a network pattern with a scale of 1 to 5 m that clearly wraps around existing topography (fig. 16), creating an "elephant hide" pattern. Where strong layering is exposed in cross-section, the cracks follow bedding planes and also create fractures cutting across the bedding, forming a "checkerboard" or "waffle" pattern (fig. 17). The effective depth of these fractures is probably about 1/5 to 1/2 of their lateral spacing. Hamilton (1984, p. 32-34) suggests the fractures result from cyclic near-surface volume changes resulting from thermal cycling, wetting and drying, or freeze-thaw.

Weather Pits

Solutional removal of calcite cement has locally created karstic landforms on slickrock slopes. The most striking features are enclosed depressions ranging from a few tens of centimeters to tens of meters in diameter and a few centimeters to over 3 m in depth (fig. 18). These "pits," "shallow," "pans," "potholes," "waterpockets," "tanks," or "weather pits" differ from typical karst sinkholes in that most do not drain into obviates solutionally enlarged joints (although some do, as in fig. 19) and are developed on bare rock rather than in surface soil or alluvium. Weather pits also occur on exposed basalt and granite. Solution pits are a similar feature developed on limestones. Weather pits occur primarily on divides but may occur locally on slopes of 10 to 20 degrees, as discussed below. Sometimes they cause inversion of relief on divides (fig. 20). The important mechanism leading to weather pit development on sandstones is the moderate rate of solution of calcite cement, which
means that standing water has a greater ability to dissolve the calcite than rainfall and run off over a sloping surface, where the runoff never approaches saturation with calcite. This leads to a positive feedback which tends to deepen and enlarge chance surface depressions. Twidale (1976) attributes the development of weather pits on granite to enhanced weathering by standing water. Many of the depressions act as traps for mineral sand and silt, or for sand transported by run off. The weather pits generally retain water for hours to weeks after rain storms. The water may be lost by evaporation, by flow into fractures, or by in-
irregular percolation into the sandstone. The relative roles of evaporation and downward seepage are uncertain and probably locally variable. The morphological and process similarity to weather pits on granites suggest that infiltration is not a prerequisite to their development on sandstones. A visit to an area of weather pits 2 to 3 days after appreciable rains revealed that some of similar size and surface catchment area were dry, whereas others contained appreciable amounts of water. This suggests that at least some of the weather pits serve as recharge points to the shallow groundwater circulation systems in sandstones that may reemerge as seepage or spring flows along cuesta scarp. Calcite dissolved from the sandstone may be carried with the percolating water or evaporated on the surface between rains, and renewed by the wind along with the loosened clastic grains. Locally shallow checkerboard or reticulate fractures develop along the edges of the weather pits, presumably due to enhanced weathering from the prolonged presence of water.

George Billingsly (personal communication, 1985) has suggested that the contraction of water frozen in weather pits may pull rock flakes from the walls of the weather pits and help to enlarge and deepen them. Ephemeral channels draining the sandstone are commonly interrupted by similar weather pits, often resulting in a “beaded” drainage pattern (fig. 21). Even small weather pits (10 cm) develop pits (figs. 22 and 23), which, because of their location along channels, are generally called potholes. The beaded pattern is generally limited to slopes less than about 10 to 20 degrees and is replaced by narrow furrows or “gutters” on steeper slopes (figs. 24 and 25). Conventional wisdom ascribes the potholes to casual erosion in plunge pools, and this mechanism may be important along some of the larger washes. However, the potholes grade imperceptibly into weather pits near divides, so that solution by standing water between runoff events may also be important along the larger channels. The enhancement of solution rates by fixed turbulent eddies during runoff may also be as or more important than direct abrasion in potholes on calcite-cemented sandstones.

Not all slickrock surfaces exhibit prominent solutional features. The reasons for their variable importance presumably involve differences in climate, lithology (particularly in the amount of calcite cement and rock permeability), slope steepness, ability of wind or running water to remove loosened sand, and degree of mantling of the surface by windblown deposits.

On a larger scale than the solutional features discussed above, Young (1986) draws an analogy between haystack-shaped slickrock slopes in Australia and tafoni of lower alpine environments, emphasizing the role of solution of the calcite cement in these sandstones in producing the rounded slope forms. He also notes that case hardening of rock surfaces plays a role in slope evolution. Young apparently discounts a dominant role for exfoliation fracturing in these sandstones. Also important locally has been the hardening of sandstones adjacent to joints by silica deposition, which results in these hardened zones forming reticulate hillcrests, just the opposite of the typical haystacks of the Colorado Plateau such as fig. 3), where joints are zones of weakness.

Figure 21: Bedding disruption and roll-on Navajo Sandstone north of US 160 west of Kanab, Utah. The slope gradient is about 15 to 25 degrees.

Figure 22: Potholes along a channel in Navajo Sandstone near Newspaper Rock of US 89 and 191 north of Moab, Utah.
Coesta Scars

The exposure of weaker strata (generally shales or highly fractured sandstones) beneath massive sandstones causes undermining of the sandstone, leading to cliff development and rapid cliff backwasting. A far greater volume of rock is initially broken up by scarp retreat than by erosion on slickenside slopes when considering average rates over large areas. Because of the rapid retreat of scarp slopes, the cliffs generally cut back into preexisting slickenside slopes. Figure 26 shows a situation in which uplift exposure of shale near the stream level (right side of figure) has caused the development of cliffs, and their backwasting into slickenside slopes, such undermining was discussed by Albritt (1986) and Oberlander (1977).

The relative rates of erosion on different parts of a coesta can be illustrated by considering, as a first approximation, that the form elements maintain a constant gradient and a constant position relative to the stratigraphic layers through time. These assumptions require a constancy of both stream erosion and slope processes through time which probably approximates the long-term average behavior of scarp erosion but not the short-term changes due to climatic fluctuations. The assumption of a constant position of slope elements relative to the stratigraphy (fig. 27B) is clearly a closer approximation to scarp evolution than the assumption that slope elements retain a constant position through time (i.e., a constant rate of vertical erosion on all elements of the scarp (fig. 27A)).

In horizontal stratified rock, these assumptions predict a rate of vertical erosion proportional to the slope tangent, whereas the horizontal rate of erosion (lateral backwasting) is idealized on all slope elements (fig. 27B). This implies an infinite rate of downwasting for a vertical cliff, which is an artifact of considering cliff retreat as continuous rather than as discrete events such as rockfalls. Therefore, as mentioned above, on a typical escarpment the downwasting of the slickenside slope on top of the caprock is very slow compared to both cliff retreat and to vertical erosion below the rim (fig. 28).

The relative rates of erosion on various slope elements are also affected by the structural dip. All form elements erode at an equal rate parallel to the structure that is, in a downwarp direction with constant gradient, then the instantaneous rate of vertical downwasting, \( V \), is given by the structural dip, \( d \), the slope angle, \( s \), and the rate of downwarp backwasting of the escarpment, \( D \). Where the slope is inclined with the dip (fig. 29A):

\[
V = D \cos d \cdot \cos s \sin d \quad \text{for}\quad 90^\circ > d > 0^\circ.
\]

And (fig. 29B):

\[
V = D \sin s \cos d - \sin d \quad \text{for}\quad 90^\circ > d > 0^\circ.
\]

Where the slope opposes the structural dip (fig. 29a):

\[
V = D \sin d - \cos d \tan s \quad \text{for}\quad 90^\circ > d > 0^\circ.
\]

During continued downcutting by streams draining the escarpment, the relief should adjust until downwarp
exposure of new caprock and upsip removal by backwashing are roughly balanced. Therefore, the rate of vertical reduction of the rim should be independent of the dip maintaining a constant relief through time, whereas horizontal retreat of the escarpment would be inversely proportional to the tangent of the dip, and the volume of caprock eroded per unit time would be inversely proportional to the sine of the dip. The very rapid rate of horizontal retreat predicted for low dips does not occur because the escarpment becomes segmented by erosion along drainage lines into isolated mesas and buttes whose local relief, distance from the main escarpment, and rate of backwasting increase through time. Nevertheless, these considerations imply that, in general, a greater volume of rock must be eroded per unit time from gently dipping scarps than from steeper ones. The gradients and total relief on a given scarp should increase where the structural dip decreases to maintain relatively constant rates of vertical reduction of the rim. Figure 20 compares relief of escarpments on two sandstones in the Henry Mountains area, Utah, as a function of the reciprocal of the sine of the dip, showing that there is, in fact, a relationship of the type predicted. Wasting of caprocks occurs primarily by rockfall, undercutting, slumping, and fretting. Rockfall includes events ranging from calving of individual blocks to the failure and fall of a wide segment of the face, resulting in a rock avalanche on the scarp rampart. Some sandstone caprocks (especially in the Morrison Formation) are underlain by block-by-block by weathering and erosion of the underlying shale without rapid fall of the undermined blocks (figs. 31 and 32). The blocks may be repeatedly lowered with lit-
Shale. Slumping is prevalent on relatively few escarpments, where it may dominate as the mechanism of scarp retreat (figs. 33 and 34). The Toreva block slumps are a classic example (Reiche, 1937). Conditions leading to slumping failure have not been firmly established, but a low shear strength of the unweathered subcaprock unit is probably the major factor. Low shear strength can result from low bulk strength or a high degree of fracturing and/or abundant bedding plane partings. Other factors in some cases may be deep weathering of the sub-caprock unit by groundwater flow and high pore water pressures. Rockfall is the most common form of scarp retreat (fig. 35). Over time, a rough balance is maintained between the production of debris at the scarp face and its removal from the rampart. Debris produced
by rockfalls with high potential energy may result in powdering of a large percentage of the original rock (Schumm and Cholette, 1966). On many scarp slopes, the coarse debris produced by the rockfalls must be weathered and eroded before further scarp retreat can occur (fig. 36). Weathering processes acting on the debris are similar to those occurring on slickrock slopes, including splitting or shattering, granular disintegration, and solution of cement for the rock masses in the case of limestones. The necessity for weathering of scarp-front debris before further erosion of the sub-caprock unit leads to a natural episodic nature of rockfalls and scarp morphology, as outlined by Koons (1955a) (fig. 37). Where caprocks are eroded primarily by large rockfalls, continued erosion of the sub-caprock unit at the margins or base of the rockfall eventually raises the debris blanket into relief, sometimes forming subsidiary small escarpments where the
Figure G. Portion of escarpment of Torrey Sandstone near Wanton Shale on North Canyone Mesa near Bunkerville, Utah, along U.S. 93. (A) North-western deposits of mudflat deposits that are dissected by continuing erosion at base on the rampart. The escarp in the above extends into the underlying Marmor Shale. A steep edge line present at the base of the Torrey Sandstone in the above.

(B) An extensive all deposits blanket is strongly dissected at its base. The presence of the deposits blanket is further supported by the cutting of the beds is more readily and marked. The deposits blanket is probably a mudflat sedimentation represented by massive deposit. The sandstone in the middle distance is formed from the erosion of a mudflat depository extending from the escarpment to a mudflat gravel. Some evidence along the Fremont River behind the photographer.

Figure H. Recent mudflats in Navajo Sandstone showing abundant mudflat deposits. (A) Rockfall southeast of Kanab, Utah. (B) Rockfall in escarpment near Sheep (or Sheep) Jasper, Ariz. 65' and 66'.
debris blanket is subject to further mass wasting (fig. 33b). Thus old rockfalls stand well above surrounding slopes of both exposed caprock unit and younger rockfalls. The relief and steepness of scarps eroded by rockfalls is controlled not only by necessity to cause failure of the caprock unit, but also by the length of time and relief necessary to weather and erode rockfall debris.

Fretting is used here to refer to surface attack of caprock and subsurface units by salt tretting, freeze-thaw, cement dissolution, and similar processes occurring at zones of seepage discharge. Because of the concentrated focus of attack, fretting commonly results in accompanying spalling and collapse development typical types of rockfall. Fretting processes are discussed more fully below.

Most prominent scarps on the Colorado Plateau are formed of massive sandstone underlain by shale or other easily weathered rock, so that backwasting is caused by loss of bulk strength of the underlying layer accompanied by erosional attack of the scarp rampart. However, some of the incompetent layers producing scarps are strong in bulk but are eroded primarily because of denser fracturing relative to more massive but not necessarily stronger overlying sandstones (Oberlander, 1977; Nicholas and Dixon, 1986).

Segmented Scarps

Many areas of moderate to great relief on sandstones on the Colorado Plateau exhibit a complex topography embodying elements of both slickrock morphology and scarps. Such landscapes developed in the slickrock member of the Entrada Sandstone at Arches National Park are the object of a comprehensive study by Oberlander (1977). In this area, slickrock slopes are interrupted by nearly vertical cliffs which Oberlander terms "slab walls" due to their erosion by failure along sheeting (anti-loading) fractures parallel to the scarp face. The slab walls terminate at their base at indentations developed in thin weak zone (partings) whose weathering and erosion cause the slab wall backwasting (figs. 38 and 39). Partings that readily weather ("reflective partings") are either closely spaced bedding planes with high fractured sandstone.
sandwiched in between, or are one or more thin (2 to 2 cm) layers of tafoni terrigenous shale (fig. 40). The partings commonly are of limited horizontal extent, so that slab walls die out laterally (fig. 39A). Some partings may have more than one slab wall where partings occur at two or more levels (figs. 38B, 38C, and 39). Oberlander presents convincing evidence that slope erosion occurs by both erosion of slickrock slopes and slab wall backsliding. This, coupled with interaction of new partings and lateral dying out of other partings during slope retreat, leads to progressive changes in slope profile form (fig. 41). In Oberlander’s model, the gradient of slickrock slopes below effective partings depends largely on the relative rates of scarp backsliding by parting erosion and the rate of weathering and erosion on the slickrock slopes, with gentler slickrock slopes associated with rapid parting erosion. Sometimes backsliding at a parting may cease, due to playing out of the parting or to local conditions less conducive to parting erosion. In such cases, continued erosion of the slickrock slope below the parting leads to the development of a near-vertical slope below the parting; such slopes are called “secondary walls” by Oberlander. Such inactive slab walls also commonly develop alveolar weathering (fig. 38B), discussed further below. An important conclusion of Oberlander’s study is that thin partings in otherwise massive bedrock cause a complicated slope form (in particular the slab wall) so that slope breaks are not necessarily an indication of lithologic differences above and below the slab, but imply only a thin discontinuity.

Similar slope forms occur in other sandstone units, especially the Navajo Sandstone and the Cedar Mesa Sandstone’s Natural Bridges National Monument (fig. 38). In those formations the slab wall is often strongly overhanging into a thin two-dimensional arch or alcove presumably backwashed along sheeting fractures. One puzzling aspect of these prominent indentations is a general paucity of mass-wasting debris on the lower floor (figs. 38B and 38C). Schumm and Chorley (1964) cite the steady breakup of the wasted debris as an explanation, but slab wall failures from relatively short cliffs yield abundant debris (fig. 42), and the alcoves are a relatively protected environment. Another possible explanation is present-day inactivity of
Figure 41. Development of scarp forms in massive sandstone through time and space. For simplicity, a constant ground level is assumed during scarp retreat, along with equal thicknesses at removal from major slab walls in each unit of time. At A, the planar scarp is preserved due to upward erosion of the bedded substrate. At B, the thick-bedded substrate passes below ground level, so scarp retreat stops, and effective initiation of scarp retreat is not possible. Effective scarp retreat is possible at C, E, F, G, H, and I, leading to local slab wall separation and movement into slickrock. Parthus that open at D, F, G, and I initiate growth of new slab walls. Note that the effects of further parths at B and D that have been removed continue to be expressed in the form of slickrock ramps and concave-zone slopes. Erosion of ground level during this lowering would cause large lobe extensions upward from major contact (caption and illustration from Oberlander, 1977.)

Figure 42. Aerial view of double alms in Navajo Sandstone in the Lodore Lassen Plateau (fig. 45), showing sandstone debris created by recent rockfall.

parting erosion and resulting slab failure due to aridity. The effects of climatic change on scarp morphology are discussed further below.

A different type of slab wall characterized by lack of an obvious parting layer occurs locally on massive sandstones. The lower contact between the slab wall and the lower slickrock slope varies in vertical position (fig. 43). One explanation is that the slab walls are not backwasting at present, whereas the lower and overlying slickrock slopes continue to erode. Thus the vertical wall is similar to the secondary walls of Oberlander. A lack of mass-wasting debris below the slab wall is consistent with this interpretation. The lack of obvious resistant layers or parths makes it unclear why the slab walls developed. However, the present inactivity of the slab walls is consistent with their poor exposures to weathering processes and the lack of undermining.
Role of Sapping Processes in Scarp Erosion and Morphology

Various geomorphologists have suggested that rock weathering and erosion at zones of groundwater discharge have contributed to the backwasting of scarp forms, and valleys in sandstones exposed on the Colorado Plateau (Gregory, 1917; Bryan, 1928; Ahnert, 1960; Campbell, 1973; Lehty and Malin, 1983). Lehty and Malin (1983) define sapping as "the process leading to the undermining and collapse of valley head and side walls by weakening or removal of basal support as a result of enhanced weathering and erosion by concentrated fluid flow at a site of seepage." Higgins (1984) distinguishes between "spring sapping," caused by concentrated water discharge and "seepage erosion," resulting from diffuse discharge at lithologic contacts or other lithologic boundaries. This discussion addresses the general question of the role of groundwater in slope erosion on the Colorado Plateau and the specific question of the role of groundwater in erosion of deeply incised valleys in sandstone that bear a morphological similarity to some martian valleys.

The definitions of sapping and seepage erosion given above are likely to occasion semantic arguments about marginal situations. Scarp erosion processes that are clearly not sapping erosion include plunge-pool undermines and rock weathering by moisture delivered to the scarp base by precipitation, condensation, or absorption of water vapor. However, other circumstances are not as clear-cut. For example, water penetrating into tensional and exfoliation joints close to cliff faces and causing rockfalls as a result of freezing or water pressure would probably not be classified as sapping by most geomorphologists. Similarly, corrosional erosion of shale beneath sandstone by water penetrating along wide fractures (Ahnert, 1960) terms this "subterranean wash" is similar to piping, but probably should not be included as a process of groundwater sapping. On the other hand, rockfall caused by weathering of shale beneath a sandstone scarp in which water is delivered by water flow along joints within the sandstone is more likely to be considered sapping, even in the absence of obvious water discharge along the scarp face. Weathering processes resulting from intergranular flow within sandstone would generally be considered sapping. Seepage and sapping weathering and erosion are defined in the context of this paper as discussed in this paragraph.

Groundwater flow plays an uncertain role in the weathering of the shales and weakly cemented layers whose erosion causes scarp retreat in overlying sandstones. Oberlander (1977) mentions spring sapping as a process of scarp retreat, but felt it is limited to scarp near the top of slopes where flow paths through the sandstone are short. He apparently also felt that flow occurs primarily through fractures. In the 1977 paper it is unclear what processes Oberlander thought were responsible for the erosion of the partings that result in the segmented slopes. However, in a per-
sonal communication, discussed further below, he suggests that weathering can occur by surface proc-

esses unrelated to sapping. Schumm and Cheleski (1966), while providing experiments and observations on weathering of caprock units, essentially avoid the issue of processes of scarp retreat. Koons (1955) is similarly vague about the undermining processes.

Ab wzgl (1960) clearly feels that sapping processes are of great importance in scarp retreat in sandstone/shale scarp areas of the southwest. But he provides little evidence. Leith and Malin (1982) suggest that disrup-
tion of surface-exposed salt crystal growth where seepage emerges and sloughing of thin sheets of the bedrock are the major processes of erosion in massive sandstones, and that sapping is usually con-
centrated in thin zones above low permeable boun-
daries within or below the sandstone. This back-
ward and undermining of the overlying sandstone then occasions development of slab failure and, locally, alcove development associated with the development of exfoliation jointing, as outlined by Bradley (1963). Leith and Malin discuss primarily spring sapping processes occurring at canyon head-
walls, and the degree to which they feel sapping or seepage erosion occurs more generally on sandstone scarp is uncertain. Observations reported below sug-
gest a fairly important role of shallow groundwater circulation in scarp retreat in sandstone/shale sequences of the southwest under present climates.

Positive and Negative Evidence for Surface-

Directed Erosion

Oberlander (1977) feels that active backwasting is occurring along the study "effective partings" dis-
cussed earlier (fig. 40). Leith and Malin (1982), in a recent personal communica-
tion, Oberlander clarifies the processes that he feels cause removal from the partings:

The active partings I have looked at show no evidence of seepage or any recent time. I think that very weak and coherent seepage would be evident from conspicuous vegetation, moss or fungal growth; or staining of rock by white salts or dark. Merrich desert varnish similar to that seen in streams on rock faces. Varishlstroming bacteria do best in spots that are occasionally wetted and the extremely dry spots that some gorgions cannot colonize and "accl" the niche. If wetting is more frequent, fungi and vegetation will show up--as they clearly do at the many ac-
tive seeps in the area. Where there is seepage to-
day, I feel certain that there are open routes for downward penetration of water, and that joints or exfoliation fissures play a dominant role. I find it hard to accept downward penetration through the pores of a hundred or more feet of sandstone under the present climate of the Plateau. Accord-

ing to studies of downward penetration of water re precip amount (in soil), it just doesn't work—
otherwise there would be seeps all over the place!

So how does removal at thin effective partings occur? The finger-thick layer of shale in the shallow recesses between the overlying and subjacent-
course massive sandstones seems solid. Yet I presume its wetted by water working down to it through fissures. Somewhere the shale is work-
ing out of its recess and onto the cliff face. Here

and there on the base of the subjacent layer are a few mm-size flakes of the shale. I presume that these thin and discontinuous shale interbeds started out as fine sediments washed and blown onto flat surfaces areas in the ancestral \,e, where they were quickly buried by advancing

tunes. The lack of depth of these deposits pro-
vides a measure of the rapidity of dune advance:
Such material would have a high content of salt that were plays or subflats anywhere nearby. The shale could be analyzed for its salt-
ity (I'd bet it's high). If the stuff is rather saline, its exposed edge would absorb water from the windblown air, especially during drying periods. This edge swells, slakes, and crumbles. How it gets scattered out was I can watch for the winters there being beautiful, but icy—very icy.

On the other hand, many scarp slopes in the southwest are composed of cliffs that extend well below the caprock unit into the underlying shales (fig. 35A). These cliffs are remarkably stable, having persisted and grown vertically (downward) throughout the Holocene (and late Pleistocene), showing the efficacy of even a small overhang in restricting surface weathering of shales in an environment (see fur-
ther discussion below). The widespread development of tall pedestals of sandstone supporting sandstone capstones (hoodoos or damoselites) is further evidence of such protection from weathering pro-
cesses and the relative inefficiency of windblown rain or snow, or moisture condensation in shale (or sand-
stone) weathering beneath overhangs of more than 2 or 3 feet.

Steep streamcutting by corrosion or plume pool ac-
tion is also of questionable importance in scarp retreat, at least for washes with drainage areas less

than a few square miles. Washes in sandstone in sandstone generally occupy only a fraction of the total scarp width. In addition, the scarp is

Currently overhung when developed in massive sandstone, and plume pools are rare and small below the

waterfalls (figs. 44 and 47). Even steep streams

on this sandstone feels sandwiched between shales exhibits overhangs considerably wider than the

stream bed and show little development of plume pools (figs. 46 and 47). As mentioned previously, small washes developed on saddle slopes above scarp are commonly interrupted by solution pits, and

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Positive Evidence for Groundwater Sapping

Cavernous Weathering and Alcove Development

Weathering and erosion of sandstone by the effects of crystal growth occur at a variety of scales in sheltered locations. Steep slopes and scarp's in sandstone are frequently interrupted by rounded depressions, often overlapping, which intersect sharply with the general slope (figs. 48-50). Such alveolar weathering, or "talonts," occurs not only in sandstone, but also in granites, feldspar, and other massive rocks (Moore, 1982, 1983). Both accelerated erosion in the hollows, and case hardening at the exposed portions of the slope (Greece and Rosenau, 1982) may contribute to the development of talonts. Salt accumulations are often quite apparent in the cavernous hollows (fig. 57), and the bedrocking results in spalling of sheets of weathered rock up to a few centimeters in thickness (fig. 52). Mostor (1983) notes high soluble cation contents in the spall debris in talonts and the presence of the mineral gypsum. Lathy (1981) and Lathy and Maley (1985) find a large deposition on spilling walls. The mechanisms by which such mineral deposition may contribute to the spalling include pressure exerted by crystal growth, thermal expansion, and contraction of the crystallite-filled rock, and expansion and contraction due tohydration of deposited minerals.
Figure 46. Steep stream in sandstones at Colorado National Monu-
ment, Colorado, showing poor definition of channel, absence of
plunge pools, and numerous short overhanging scree.

Figure 47. Steep gully in interbedded shales and sandstones of the
Sepal Formation on the north rim of the Grand Canyon, Arizona. Note
the absence of plunge pools and the undermining of sandstone
ledges along the gully.

Figure 48. Abnormal weathering, or talus, developed in Navajo
Sandstone at Capital Reef National Monument, along Utah 24. Note
the highway reflector for scale. Also note unweathered ribs where sur-
face wash occurs.

Figure 49. Abnormal weathering at the same general location as that
shown in fig. 48.

(Cooke and Smalley, 1968). Freeze-thaw dissection
on the moist seepage faces may also contribute, and
spalling may be aided by the weight of accumulated
winter ice (Lalib and Malin, 1965).

A surface protected from surface runoff is a neces-
sary condition for talus and alcove development.
On steep sandstone scarp, surface runoff commonly
flows as sheets down the scarp, held by water ten-
sion on slightly overhanging slopes. Such runoff paths
are commonly accentuated by desert varnish, where
such runoff paths, cross-zones of talus development,
back-washing is inhibited, and the talus are separated
by columns that often resemble flowstone columns in caves (figs. 48, 49, and 53). Surface rainwater might infiltrate salt faxetting simply by solution and removal of salts brought to the surface by evaporating groundwaters or more actively by case hardening of the exposed surface by deposition of clays or calcite (Can-ica and Rowan, 1982).

Two intergrading types of sapping landforms develop on massive sandstones. The more incipient form is the development of talus on steep scarps and on large talus blocks. Such talus may literally riddle certain steep slopes (figs. 48-50), with the talus concentrated along certain beds that are either more susceptible to the salt faxetting or receive greater groundwater discharge. Talus blocks generally develop talus on their lower, overhanging portions. The concentration of talus development at the base of such blocks may be due to the protection from surface wash as well as upward "wicking" of salts from underlying soils or shales, a process that contributes to weathering of the bases of tommiesites (Mussie, 1983; Hamilton, 1984). Oberlandier (1977) points out that talus develop most strongly on scarps initially steepened by basal undermining but presently no longer backwashing because alluviation or talus deposition covers and protects the basal backwashing face. Thus generalized talus development on a scarp indicates relative inactivity of backwashing by surface attack or basal undermining.

On the other hand, large alcoves are common in massive sandstones and are often actively retrograding as a result of sapping erosion (figs. 54 and 55). However, direct sapping is usually localized to zones less than 2 m thick along permeability discontinuities where the discharge of groundwater is concentrated (fig. 56), although at major valley heads the seepage zone may be
20 to 25 m thick (fig. 57). These sapping zones generally backwaste by processes similar to those of talus and locally talus zones are superimposed on the sapping face. In addition to salt fretting, backwasting by groundwater discharge can also occur by cement dissolution and by weathering of shale beneath or interbedded in the sandstone. The retreat of the active zone of sapping undermines the sandstone above, with the result that occasional rockfalls occur (figs. 36B and 42). In massive sandstone the undermining occasions the development of exfoliation sheeting fractures, resulting in large arches or alcoves, with the deepest parts of the alcoves presumably corresponding to the minor rapid sapping attack (figs. 36B, 43A, 54A, 55, 56A, and 37B). In well-jointed sandstones, such as the Wingate Sandstone, arches and overhanging cliffs are less common, and the role of scarp retreat by sapping processes is not as obvious but may be just as important (fig. 58).

The major aqulculde for the Navajo Sandstone is the underlying Kayenta Formation, and the major saps develop at this discontinuity (figs. 54 and 58). However, thin shales and limestone interbeds (interdunal deposits) create minor aqulculde within the Navajo Sandstone (see chapters 3 and 4), leading to frequent development of multiple levels of saps and associated alcoves at valley headwalls (fig. 59).

A distinction may be made between "wet" sapping, with a damp rock face and an efficient discharge and "dry" sapping, face generally encrusted with mineral salts (Levy and Malin, 1983). In general, talus is associated with dry sapping because of their localized development, whereas large alcoves are generally associated with a more regional groundwater flow and exhibit facies that are at least seasonally wet. This distinction, like most, admits of many intermediates and transitions from one to the other type as a result of seasonal or long-term climatic fluctuations. For example, dry sapping can lead to alcove development (fig. 60).

Neither active seepage nor deposition of mineral crusts on protected sandstone walls are necessarily correlated with rapid weathering and backcutting of the sandstone walls. For example, the Weeping Wall at Zio National Park is an impressive scep emerging from the Navajo Sandstone, but the associated alcove and canyon are relatively small. Many other examples of fairly high discharge rates but only minor or nonexistent alcoves can be found throughout the Colorado Plateau. Too rapid a seepage may in fact discourage deposition of salts. Although rapid seepage can also cause backwasting by dissipation of calcite or gypsum cement, this would occur only if the groundwater were undersaturated. Similarly, many examples of thick mineral incrustations at seeps lacking evidence of backwasting can be found on sandstones throughout the southwest. Several factors control whether minor-
als deposited by evaporating seepage are deposited intergranularly within the rock (encouraging exfoliation and granular disintegration) or at the rock surface (with little resulting sapping), including the type and concentration of salts, the average and variance of water discharge to the surface, the distribution of pore sizes and their interconnectivity, the presence and size of textures, the temperature regime at the rock face, and the frequency of occurrence of wetting of the rock face by rain or submergence (if along a stream or river). The interaction of these factors is poorly understood. Discrepancies between the size of the alcove or sapping valley and the magnitude of the seep can also result from differences in the length of time that sapping has
been active. Despite these cautions, it is reasonable to expect within a specific physiographic, structural, and stratigraphic setting that the degree of alcove development and the degree of headward erosion of seepage valleys would correlate with the size of the seeps involved (Lar Ny and Malin, 1985).

In summary, most evidence suggests that sapping processes are common in sandstones of the Colorado Plateau and the processes result in the widespread development of talus and alcoves. The major caution is the evidence cited by Oberlander suggesting that groundwater sapping is unimportant in erosion of effective partings at thin shale layers in the Entrada and other sandstones. There are several possible explanations for this difference in interpretation of the major process of scarp retreat. The effective shale partings may, in fact, be so readily weathered by atmospheric moisture due to high soil content or other factors that seepage is not required for such partings as it is for cavernous weathering and alcove development elsewhere. The Entrada Sandstone exhibits considerable variation in lithology and cementing both vertically and areally, so that in the Arches National Park the Slick Rock Member may be relatively impermeable. However, in the Noliguet Canyon and talus with evidence of active salt fretting are found a few tens of miles south in the same member (e.g., Figs. 51, 52 and 60). Another possibility is that backwasting along effective partings is not very active under present climatic conditions, but was more active during the late Pleistocene. The evidence for relatively inactive scarp retreat (and, by extension, seepage erosion) under present climatic conditions is discussed below.

Large-Scale Morphological Indicators of Sapping Erosion. Groundwater sapping is probably an important process in scarp retreat throughout the Colorado Plateau, as suggested by Ahnert (1960). However, the landform assemblage closest to martian valley systems is the deep, narrow canyon networks of the type discussed by Lar Ny and Malin (1985).

The planiometric form of canyons and escarpments is the most obvious signature of the erosional processes involved in scarp retreat in layered rocks. In the absence of concentrated erosional attack, erosion of caprock units would be by uniform decrescence (Fig. 13), resulting in scarp profiles with sharp projections and broadly concave resistant members (Dutton, 1983, p. 238-239; Davis, 1901, p. 178-180; Lange, 1959). Attack at an escarpment by uniform decrescence would gradually make embayments more shallow and the planform of the scarp face closer to linear. Almost all escarpments in gently dipping rocks exhibit deep reentrants and, as erosion progresses, a breaking up into isolated mesas and buttes. This indicates erosion concentrated along generally linear zones of structural
weakness, fluvial erosion, or sapping (often acting in combination). One evidence for the role of either or both fluvial erosion and groundwater sapping is the asymmetry of scarpas in gently tilted rocks. The planform of segments where the scarp faces up dip is generally similar to that expected by uniform decay resistance, since little drainage passes over the scarp, and segments facing down dip ("back scarps" of Aiken, 1983) are deeply indented as the result of fluvial or sapping erosion (fig. 61) (Aiken, 1969; Lally and Malin, 1985). Since both fluvial erosion and sapping produce reentrants, the mere presence of deep reentrants is not conclusive evidence for sapping erosion.

Nicholas and Dixon (1986) emphasize the role of variable density of fracturing of incompetent and caprock layers in controlling the rate of scarp retreat and the development of reentrants, projections, and isolated buttes. Their evidence suggests that variable density of fracturing is important on certain scarps, but it probably controls primarily the small-scale planform features. Variable density of fracturing cannot account for the remarkable asymmetry of gently dipping scarps.
Figure 59. Scarp in Wingate Sandstone. Scarp at the left and right are typical of the well-jointed sandstone. The smooth cliff face at the center are fairly unusual for scarps in this formation.

Figure 59. Valley headwalls with seeps at multiple levels. At least three levels of seeps are visible along the Canyon walls (figs. 64 and 65). The seeps occur on horizontal layers of interbedded shale and sandstones within the Navajo Sandstone (see chapter 3 and 4). The headwall in the photograph is approximately 10 m thick. On the canyon headwall in Tooele County, Utah, an eastern tributary of the Virgin River has cut a large, nearly dry, hard sediment with three major levels of seeps and alcoves. The headwall is about 200 m high. See fig. 66 for location and fig. 67 for a general view.
Figure 6.1. Plan view of an escarpment of gently dipping Wingate Sandstone showing drainage, nearly straight bluff to upper side and densely embayed lower side. Wingate also dips about 1.5 degrees due west. The escarpment is south of Utah 95 between the Colorado River and Natural Bridges National Monument.
Ahern (1966) and Laity and Malin (1985) suggest that fluvial erosion will produce V-shaped canyon heads that widen consistently downstream, whereas sapping produces U-shaped, or theater-headed canyons of relatively constant width downstream (the terms U- and V-shaped refer here to valley planform, not valley cross-section). Laity and Malin (1985) suggest that the difference in canyon planform for updip (V-shaped) and down-dipping (U-shaped) valleys dissecting the Navajo Sandstone (figs. 54, 62, and 63) is an indication of a prominent role of sapping erosion in the latter and fluvial erosion in the former. However, this comparison is weakened by the difference in dip orientation between the two sets of valleys. The updip-draining valleys do not cut through the resistant sandstone into the underlying shaley Kayenta Formation, so that valley deepening by fluvial plunge-pool action and canyon widening by undermining due to...
non-sapping-related erosion of the shale would not occur, but could be operating in the downdip-draining valleys that do expose the Kayenta Formation. Furthermore, even where updip- and downdip-draining valleys expose the same stratigraphic sequence, headward canyon migration is inhibited in the former by the plugging of the resistant caprock and enhanced in the latter by the updip increase in caprock elevation. A later comparison would be between valleys in the same structural setting (same dip and valley orientation), with similar over- and underlying rock units with comparable compressive strength and structural integrity, but having different permeability or different cementing agents (e.g., calcite versus less soluble silica). Finding such a comparative situation may be difficult or impossible.

Fortunately, many of the other diagnostic features cited by Laity and Malin (1983), such as canyons with large theater heads having alcoves and seeps, are much more convincing of a dominant role of sapping in canyon excavation and widening. The types of features indicative of a sapping origin proposed by Laity and Malin are discussed below, using as an example the box-canyon system tributary to Navajo Creek on
the Navajo Indian Reservation near the Inscription House Ruin (figs. 64–67). The rocks in the mapped area slope gently to the southwest with dips of 1:80 to 1:40. The center of the mapped area is the crest of a very broad anticlinal flexure superimposed on the regional dip. The canyons are eroded into the Navajo Sandstone. The overlying shales and sandstones of the Carmel formation have been stripped from the Navajo over most of the mapped area except along the southern portions of the divide followed by the Navajo Mountain Road. The underlying flaggy shilstones, mudstones, and sandstones of the Kayenta Formation are exposed in the bottom of all canyons, indicating that the primary locus of backwasting is near this contact. The central sections of Geshi and Far End canyons have a secondary scarp in the center of the valleys developed in Wingate Sandstone exposed along the crest of the anticlinal flexure.

The Inscription House area exhibits the same types of features cited by Lathrop and Malin (1985) avidence of backwasting processes in a similar area about 50 km north along the Escalante River. These features include theater-shaped heads of first-order tributaries (fig. 39), relatively constant valley width from source to outlet (fig. 64–67), high and steep valley sidewalls,
pervasive structural control, and frequent hanging valleys. The most direct evidence of sapping processes are the numerous alcoves, both in valley heads and along sidewalls (figs. 59 and 60). Springs are numerous on the valley walls and bottoms. Although many valley headwalls occur as deeply eroded alcoves, some terminate in V or half-U shapes with obvious extension along major fractures (fig. 61). Although sapping action is fairly evident where alcoves occur at valley heads, sapping may also occur in fracture-controlled headwalls. Evidence for erosion by stream-passing over the valley headwalls is slight, and plunge-pools are not obvious. The valley extension along fracture traces suggests control by groundwater flowing along the fractures. The regional joint directions are a strike-oriented NNW-SSE and a sub-parallelizing NNESSW set. The latter fracture system even the strongest control on the drainage network, with a tendency for valley development to extend uphill such as in the northeastern ends of Gods and Tealoshunda Canyons (fig. 65). In fact, a preponderance of the major valley heads are oriented uphill, which is consistent with sapping by a regional downslope groundwater flow.

One striking feature of the canyon network is a frequent discontinuity between the extent of broadland canyon growth and the relative upland area contributing drainage to the canyon head. For example, at A in figs. 63 and 64, a relatively large upland drainage area is associated with a minor canyon. The situations at B, C, and D are even more pronounced, since the upland drainage covers the side rather than the end of the canyon (figs. 63 and 66). Such circumstances suggest that groundwater flow, rather than surface runoff, controls headward extension of the canyons. In fact, such discontinuities are better illustrated in the Inscription House area than in the area studied by Laity and Malin.

Another indication of sapping control is the extension of canyons right up to major divides, such as at stop II and III (fig. 67). In fact, the divide has probably migrated slightly southeastward of it as a result of canyon extension. In contrast to surface runoff, groundwater flow can locally cross divides. Divide migration can also occur in surface runoff drainage basins, but is associated with the presence of slopes that are well-groomed from stream to divide. In the case of headward migration of canyons, streams above the scarp are precariously perched on the upper hillside slopes, so that base level control is very indirect.

The pattern of valley development and the relative contribution or sapping processes versus fluvial erosion is influenced by many structural, stratigraphic, and physiographic features. Infiltration may be restricted by overlying aquitards where not stripped from the sandstone (Laity and Malin, 1985; chapter 4). The porosity of the sandstone is influenced not only by primary minerals, but by diagenetic cements and overgrowths. These secondary minerals vary considerably from layer to layer and location to location (Laity, 1983; chapter 4; Kochl and Kiley, chapter 11). Laity and Malin (chapter 4) discuss the important role of thin aquitards, generally interlayered deposits, and bedding striations and slabs in controlling groundwater flow through the sandstone. Primary and diagenetic bedding features are important controls on groundwater migration (Laity and Malin, 1985; chapters 3 and 5). On the other hand, dense primary
effecting, such as in the Wingate Sandstone, restricts development of alcoves and probably diminishes weathering by cement solution and salt weathering, chapter 4). Sapping processes may still be important, but the spectacular alcoves commonly developed in the Navajo Sandstone are lacking. Relatively flat uplands on sandstones should encourage infiltration, whereas steep slickrock slopes probably lose much precipitation to runoff. Weather pits and thin covers of surficial sand may also encourage infiltration.

EFFECTS OF CLIMATIC CHANGE ON SCARP MORPHOLOGY AND SAPPING PROCESSES

Most recent discussions of scarp morphology and associated weathering and erosional processes have implied that the morphological elements can be explained by presently active processes (e.g., Koons, 1973; Schumna and Chorley, 1966; Olmstead, 1977). However, some escarpments in the Colorado Plateau region appear to be subject to considerable less caprock erosion under the present climate than during the Late Pleistocene (Reese, 1937; Ahner, 1966; Howard, 1970).

The scarpment of Emery Sandstone near Caneville, Utah, is a case in point. Extensive old-rock debris deposits on North and South Caneville-Milles Goosebed at their margins as much as 60 m were interpreted by Howard (1970) to be fluvial lake equivalent in age (Illinoian) because some of the debris blankets interfinger with pediment deposits of Bull Lake age, whereas the remnants extend no further downslope than the level of the former pediment surface (figs. 35b, 70, and 71). The old debris blankets have been dissected by a similar amount and project concordantly laterally along the escarpment. Much of the scarp rampart is now devoid of extensive debris, so that extensive badlands have developed on the underlying Marcus Shale. In places, the cliff capped by the Emery Sandstone extends downward as much as 50 m into the underlying shale (fig. 11A). Koons (1973) demonstrates that the development of cliff faces in the reservoir subarea unit is a normal occurrence in the cycle of a Navy, debris blanket erosion, and erosion of the sub-arch unit which triggers the next northward. However, for reasons outlined above, extensive cliff development in the subarcb unit may also indicate stagnation of caprock rates washing, whereas erosion of the subarch unit continues on the scarp rampart.

An outstanding example of such morphology is afforded by the escarpment capped by the lower units of the Morrison Formation overlying the thin bedded sand and shale beds of the Summerville Formation near Hazelwood, Utah. Much of the escarpment is capped by a thin caprock layer which in places appears to support vertical cliffs up to 10 m high in the Summerville Formation (fig. 2C). The caprock is eroded into small cutouts and incised in the underlying Summerville Formation (fig. 71), indicating that the caprock protects the vertical face from weathering and erosional processes. The rampart generally consists of debris-rich badlands likewise covered in the Summerville Formation, which meet the vertical face at an abrupt angle. The rampart is flanked downslope by dissected flat-topped Bull Lake drift debris, pediments, and aggradations derived from the caprock, giving evidence that abundant debris was transported from the escarpment during that time. When a caprock of caprock is present, even a thin sequence of overlying sandstone shows little debris on the rampart. However, where the caprock unit is discontinuous or missing (fig. 71), the overlying sandstone wastes (undermining and rockfall). However, during the Pleistocene climate the caprock unit evidently occurred in the Caneville and occasional screeing (fig. 72). The scarp is illustrated in photos 1 and 22 of Schumna and Chorley (1966) are probably examples of such presently inactive scars.

In canyon areas, a scarp in a massive sandstone is often underlain by a nearly bare rock slope with a gradient of about 10 to 40 degrees. No obvious lithologic break separates these slope elements. In canyons, scarps of Caneville, and escarpments can be found where such bare rock slopes yield laterally to rock
Figure 79: Major rock debris landslides on the Brevi Sandsheets adjacent to North and South Caneville Mesa. In this lower, debris deposits, with the general glacial surfaces of Brevi Lake age, these debris are identical to the larger debris. However, their present distribution suggests that they are illusory. The younger rock debris deposits are probably largely Wisconsinan. The older rock debris, indicating a generally strong ended, are not such. This is viewed from the west side of the Caneville Mesa, looking down the major divide. Major divide

Figure 71: Photograph illustrating the rockslide movement on Brevi Sandsheets near Masters Ranch on South Caneville Mesa near Brevi Lake. Note the slope, direction of the former end of the rockslide. Note the current rate of movement and the direction of the movement, which is not visible, supported on the left.

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slopes of similar gradient but mantled with rockfall debris (fig. 75). A likely explanation is that the bare rock slopes were formerly mantled with rock debris produced by cliff retreat and that the debris mantle protected the underlying rock from weathering and erosion. The cover of debris on such slopes is now obviously inadequate to protect the rock from weathering and erosion, and the implication is that scarp backwasting is now relatively inactive, which has permitted the weathering and stripping of the former debris mantle. Nonetheless, similar features can be found in the areas of segmented scarp surfaces at Dinosaur Provincial Park (the area studied by Olierander, 1977) (fig. 76). Olierander interpreted the low-gradient stickrock slopes to be a normal feature of erosion of segmented scarms (fig. 41), but it is possible that some of the stickrock slopes feature vertical cliffs may have formerly been mantled by a debris blanket under a different climate with weathering of debris can be seen in fig. 76).

Almert's (1963) historical interpretation of scarp morphology consists sharply with the "dynamic equilibrium" interpretations of Schumm and Childrey (1958 and Oilerander (1977)). Almost suggested that scarp retreat occurs by erosion of studies and other weaker resistant layers by carving-related processes that have only been active during periods characterized by abundant, gentle precipitation. By contrast, he felt that stickrock slopes are eroded largely by sheetwash, which has been active during the same but intense rainfall of interglacials. Almost felt that the generally sharp breaks between stickrock slopes and adjacent scarp was evidence of the non-contemporary origin of the two features, whereas Olierander and the present authors feel that such sharp breaks can result from contemporaneous stickrock erosion and scarp retreat. Almost's assumption that stickrock slopes are eroded primarily by sheetwash ignores the importance of the initial weathering that must occur on these weathering-limited slopes. This weathering, which makes debris capable of slope transport, is probably not optimal under the present climatic conditions of brief, intense rainstorms, but is ensured by winter precipitation and freeze-thaw effects. Despite these objections to Almost's interpretations, the examples presented previously suggest that there have been changes throughout the Pleistocene and
Holocene in the relative rates of the major processes producing coastal landforms. Sarp undermounding by sapping, undermining, or surface-controlled weathering, weathering and erosion of scarps, and locally, slumping and landsliding. Changes in the relative importance of these processes have produced landforms that in many cases can only be understood by knowledge of the temporal process changes (i.e., coastal landforms). These examples of remnant effects of past climates add a complicating element to the interpretation of scarps' morphologies. The difficulties in unraveling the geographic history of coastal landforms are compounded by the paucity of stratigraphic evidence, because, unlike fluvial systems, less deposits carry records of past events and climates. The exception is rockfall debris, which contains few
fossils, pollen, organic carbon deposits, or archeological remains, and which is very difficult to examine for stratigraphic relationships.

The occurrence of climatically related landforms and differences between present and glacial landform morphology have been discussed in a general way. The question is now directed more specifically to sapping processes and landforms, and to the relative importance of sapping at present compared to the glacial epochs.

The greater rainfall during the glacial maxima would suggest greater sapping activity then, but that same moisture supply would also contribute to salt-weathered weathering, particularly freeze-thaw weathering. Also, more abundant groundwater does not necessarily produce more sapping in sandstones, because if seepage is sufficient and humidity high enough to permit runoff of the seepage, little salt accumulation will occur. However, other groundwater-related weathering processes would be enhanced by...
greater available moisture, such as hydration and leaching of shale interbeds. Thus during the pluvial epochs we might expect more debris production by freeze-thaw, less algal weathering, but greater backwasting at the contact between shales and underlying shales and along shale interbeds. The present paucity of rockfall debris, rather than indicating nearly complete breakdown and rapid weathering of the debris, as suggested by Schumm and Choate (1966), may simply indicate the relative inactivity of many scars in present climates. Certainly there are numerous examples of abundant rockfall debris produced by modern rockfalls (figs. 36 and 42).

The limestone House area offers possible examples of reflect lambdoid elements. In some areas, the Navajo Sandstone escarpments give the appearance of weathering and erosion having cut through preexisting talus- and now backcutting along limited portions of the scarp, producing alcoves (fig. 77). This is the correct interpretation, that the localized sapping, alerosing, and algal weathering occurring at present contrasts sharply with pluvial period scars with rapid talus production and more general backwasting. The pluvial-epoch backwasting processes may also have been strongly influenced by groundwater seeping, but primarily by shelf weathering and sapping undermining. Another prominent feature of the theater headwall at the end is the paucity of rockfall debris. This contrasts with abundant rockfall debris along the sides of the valley downstream from the escarp (figs. 43B, 43A, and 48). Lamy (chapter 4) suggests that both initial pulverization of rockfall debris and rapid weathering in the moist environment of the headwall at the escarp can account for the paucity of debris. However, recent rockfalls in alcoves generate abundant debris, and the rockfall debris along valley walls also suggests that the backwasting of scarp walls produces a considerable volume of coarse debris that must be further weathered before it is transported from the rampart and backwasting can continue. Erosion or pulverization from plunge pool activity also seems inadequate to account for the small volume of debris in the alcoves, because little plunge pool development occurs and the alcoves are much wider than the streams passing over the headwall. Thus, as suggested by Lamy, particularly rapid weathering at the most headwalls causes rapid disintegration and removal of rockfall debris relative to the rates occurring on the drier valley-sidewall scars. However, it is unclear why the rapid rockfall debris weathering and erosion are not accompanied by a similar enhancement of headwall sapping and alcove deepening by spalling. Nonetheless, different processes may be responsible for weathering of rockfall debris than those which produce the primary sapping erosion at the same time. The weathering of the rockfall debris is likely to involve upward working of soluble salts from the seepage flowing at the interface of the rock and the underlying debris. This would produce an active zone of dry salt tretting within the debris-the overhang of the alcove often protects the debris from rainfall and the removal of the soluble salt. However, at the seepage face the constant supply of water would not allow deposition of strongly soluble salts, and the backwasting would have to occur by either deposition of less soluble minerals such as calcite (Lamy, 1988) or solution of

Figure 77. Multiple alcoves in a theater-headland escarp in Navajo Sandstone in the limestone House area (fig. 63). Note the vegetation in the alcoves, the absence of fresh debris within the alcoves, and the presence of weathered rockfall debris at possibly earlier deposits on the finger-like projections between alcoves. These projections are probably bedrock crested with thin mantles of rockfall or talus mantles.
the high-angle, flat, and the igneous rocks would have less salt weathering and more through-thinning water flow being salts, whereas deeply weathered rocks on the steep sides of the rock walls might encourage more salt weathering. However, it is possible that the scarcity of recent rockfall weathering is due to the present inactivity of the sapping erosion that usually accompanies the interpretation of the recent rockfall. A frozen climate might encourage more rapid sapping erosion while causing less enhancement of debris weathering.

The number and extent of active seeps in alcoves in the San Juan Sandstone appear to be less than would be expected for active vadose development by sapping processes. Many major alcoves are presently dry or have small seeps that are in only part of the alcove. Most of the alcoves in which cliff dwellings were built 800-1400 years ago by the Anasazi have few or no seeps. Although a dry environment may have been a factor in alcove selection, seeps must have been active at some time in the past to create the alcove. Similarly, low cliffs have occurred near only a few cliff dwellings, even since their occupation, and most of these are small. The paucity of recent rockfall and the general absence of alcoves on the Colorado Plateau has already been discussed.

Most active seeps in alcoves are watered into the San Juan Sandstone from local springs located near the tops of the cliffs. These seeps are generally derived from local rock sources; many of the seeps and associated alcoves occur immediately below the surface. Many of these alcoves are 45 and 51N. Many of these high-level seeps are not associated with strong alcove development for example, the high-level seeps at the north edge of figure 8B. The seeps associated with major alcoves at canyon heads at the contact with the overlying Kaiparowits Formation are commonly smaller in volume than the overlying seeps, or are presently dry, suggesting that present volumes of discharge are insufficient to sustain a strong regional groundwater flow, although they are sufficient to maintain perennial streams in many of the canyon bottoms. Many of the present seeps occur either below the alcove into the stream or its alluvium, or below the layer that defines the farthest zone of backwasting in the alcove. However, Holocene hydrogeology has buried the lower portions of most alcave seeps (fig. 8B); this burial may also be important in restricting backwasting. Also, many of the seeps at present do not extend across the full width of these alcave seeps. A wetter past climate would have favored the deeper regional groundwater flow and would have contributed to thicker and more extensive seeps.

These observations on reduced levels of seep activity, as compared to a past fluvial climate at the time of major alcove development and valley extension, pertain primarily to eastern alcoves in the San Juan Basin in the northern Arizona portion of the Navajo Indian Reservation, such as around the Aspersion House area (fig. 64). Seeps at canyon heads, in tributaries to the Escalante River described by Laby and Malin (1935) appear to be more active than described above. However, Laby and Malin note that many alcove alcaves are presently dry, and Laby notes in chapter 4 that sapping processes may have been more active in the past. The deeper canyons (and thicker San Juan Sandstone)
in the Inscription House area, as compared to the Escalante River area, may contribute to the difference. Since "typical" flows will be interrupted by more intermittent aquasalutes, the few canyons carved into the Navajo Sandstone are so difficult to traverse that oblietion regional characterization of such activity is impractical except locally, such as the intermittent access afforded by Lake Powell to the canyons studied by Laity and Malin (1965).

The difficulties involved in interpreting the geologic history of slope processes will probably mean slow progress toward direct resolution of the question of scarp landforms. Another approach might be to study the comparative morphology of scarps in different present-day climates. Such comparisons have their own problems in equivalence of lithology, structure, and relief. As an example of the potential usefulness and difficulties associated with this approach, Navajo Sandstone scarps, northwesterly of Kanab, Utah, exhibit much less alluvial weathering and slope development and more evidence of coarse physical weathering and talus production than in the Inscription House area, which has a similar elevation and in which the scarps have similar relief (fig. 79). It is uncertain whether these differences are due to climate, structural, or lithologic factors.

Other methods might be used to assess the history of slope processes on the Colorado Plateau and the degree to which sagging landforms might be relict. Detailed study of alluvial chronology in association with rockfalls (rockslides) burning, ablation, and恋爱 could be used to estimate the relative frequencies of rockfalls (fig. 80). Similarly, the occurrence of rockfalls onto an abandoned site in a species can give an indication of the degree of recent rockfall activity in scarps.
UNRESOLVED QUESTIONS AND RESEARCH OPPORTUNITIES

From the preceding discussion it should be evident that our understanding of the processes involved in sapping erosion is fragmentary, and that it is difficult to make conclusive statements about the past and present roles of sapping processes in scar formation on the Colorado Plateau, although there are strong indications that sapping processes are important and perhaps dominant on some scar. Any studies of the processes or history of scar processes would be welcome, particularly those involving the role of sapping and seepage.

The problems of interpretation become even more acute when we attempt to extrapolate to Mars from possible terrestrial analogs, because many of the morphological details indicative of sapping, such as alveolar weathering and alcove development, are not visible at the resolution of martian imagery. Furthermore, the great age of the fluvial features, together with landform modification by eolian and impact processes, means that many large-scale but subtle features, such as shallow drainageways on uplands, may have been obliterated. Even some of the large-scale features cited by Leith and Malin (1985) as being indicative of sapping processes must be regarded as tentative, such as the udip and diploid valley comparisons discussed previously.

Quantitative morphometric characterization of scar and canyon morphology may help to distinguish sapping-related features from those produced by runoff processes (Kochel et al., 1985). As an example, relatively constant down valley canyon width should be easily distinguishable from gradually increasing width downstream by plotting valley width versus order or drainage area. Another characteristic of scarps backed by sapping mentioned by Leith and Malin is that the rate of backwasting should in general be positively correlated with the drainage area contributing to the scar face; this would not be true for scarps backed by flood by surface attack. Conversely, backwasting by sapping implies that the removal of the last remnants of a caprock should be more difficult than earlier stages. Many gently dipping escarpments on the Colorado Plateau have a wide zone of thin salts, small buttes, and needles (fig. 61), with a classic example being Monument Valley. A high proportion of such forms may characterize sapping retreat of scar fronts. Parts of the fretted terrain on Mars have a similar appearance. However, the research of Nicholas and Dixon (1986) demonstrates that variable density of rock fracturing affecting rock strength may also be an important factor leading to the development of isolated buttes; some of the fretted terrain on Mars has clearly developed by erosion localized along fracture zones (although sapping dominated by groundwater flow through the fractures could be the controlling process).

Recent studies by Kochel and Piper (1986) have shown that morphometric parameters can be used to distinguish between runoff-dominated and sapping-dominated valleys on Hawaii. Figure 81 shows an example of principal component analysis applied to valleys on the islands of Hawaii and Molokai. These plots show a clear separation in most cases between runoff and sapping valleys. Table 1 summarizes the morphologic and morphometric attributes of sapping and runoff valleys formed in Hawaiian basaltic.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Runoff dominated</th>
<th>Sapping dominated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basin shape</td>
<td>Very elongate</td>
<td>Light bulb shaped</td>
</tr>
<tr>
<td>Head termination</td>
<td>Tapered, gradual</td>
<td>Theater, abrupt</td>
</tr>
<tr>
<td>Channel bend</td>
<td>Uplifted</td>
<td>Variable</td>
</tr>
<tr>
<td>Pattern</td>
<td>Parallel</td>
<td>Dendritic</td>
</tr>
<tr>
<td>Junction angle</td>
<td>Low (40°–50°)</td>
<td>Higher (55°–65°)</td>
</tr>
<tr>
<td>Downstream tributaries</td>
<td>Frequent</td>
<td>Rare</td>
</tr>
<tr>
<td>Relief</td>
<td>Low</td>
<td>High</td>
</tr>
<tr>
<td>Drainage density</td>
<td>High</td>
<td>Low</td>
</tr>
<tr>
<td>Drainage symmetry</td>
<td>Symmetrical</td>
<td>Asymmetrical</td>
</tr>
<tr>
<td>Basin area/canyon area</td>
<td>Very high</td>
<td>Low</td>
</tr>
</tbody>
</table>

Experimental development of sapping stream networks also permits qualitative or quantitative comparisons with possible martian analogs, and such an approach is being pursued. Such experiments offer the advantage of controlled conditions and known structure and material composition, but they are somewhat limited by small physical size and problems of scaling to the target (martian) features.

One approach to moving from strictly qualitative comparisons is to model scar retreat numerically, making assumptions regarding the processes involved, structural influences, and lithologies. A fairly heuristic approach to such modeling is discussed in chapter 5.

ROAD LOG

The sapping road log follows the field trip as it occurred. Suggestions for additional stops and side excursions are included. George Bilingly provided valuable information on stratigraphic nomenclature.

Day 1

Mile 0 Junction of US 89 and US 180 at the east end of Flagstaff, Arizona. Proceed North on US 89. The route is shown in fig. 67 by heavy lines. The initial portion of the tour route passes over Quaternary
Figure 81: Principal component analysis of drainage basin morphology on parts of the islands of Mauii: Kohala area and Molokai. Basins containing large deep canyons with abrupt headwalls are classified as "sapping," whereas shallower basins with a more linear profile are classified as "runoff." The variables measured on these basins are the mean junction angle (MSA), the basin shape (KS), the total drainage density (TD), the confluence basin relief measured at mid basin (CR), the first-order channel frequency (FT), and the ratio of the total basin area to the area of the canyon (RC). The measured values of these variables for individual basins are plotted against the principal component score for each basin for the first three principal components. Principal component scores for the first three components are also plotted against each other. The database for the principal component analysis included all available variables of basin morphology, not included in the plots. Although in this analysis the initial classification of the basins into sapping versus runoff was subjective, a distinction between the two basin types is consistently manifested in the principal component scores. Objective statistical classification techniques are now being investigated.
volcanic features associated with the San Francisco Peaks.

Mile 11.6 Sunset Crater loop road. Stay on US 89. (The Sunset Crater area is a nice sidetrip to see young basalt flows and cinder cones.)

Mile 25.9 Other end of Sunset Crater loop road. Stay on US 89.

Mile 35 The road drops off Quaternary basalt flows onto the Moenkopi Formation and Kaibab Limestone.

Mile 46.8 Junction of US 89 and AZ 64. Stay on US 89.

Mile 51.4 Badlands in Chinle Formation.

Mile 61.4 [0] Junction of US 89 and US 180. Turn right onto US 160 and reset the mileage counter. The road now climbs upward in section through Moenkopi and Kayenta Formations onto plateau on Navajo Sandstone mantled with eolian deposits. Over the next 50 miles the road passes near Tertiary and Quaternary geomorphic surfaces (pediments) mapped by Cooley et al. (1969).

Mile 35 The road lies on Navajo Sandstone with eolian mantle. Black Mesa, capped with Cataraxus shales and sandstones, lies to the southeast of US 160 for the next 50 miles.

Mile 49.7 [0] Junction of US 89 and AZ 98. Turn left onto AZ 98 and reset the mileage counter.

Mile 12.1 Junction with NAV 16, the road to Navajo Mountain. Turn right on this dirt road. You are entering the area shown in fig. 65.

Mile 17.5 Inscription House Trading Post. Four-mile loop trip on dirt road to overlooks of Toolevashau Canyon and Inscription House Canyon areas with excellent examples of alcove development. Inquire locally for road shown on map. Discussion themes: slickrock versus undercut morphology; discrepancies between surface drainage and pattern of valley extension; processes of sapping and alcove development; role of fractures, crossbeds, and impermeable interbeds in controlling sapping processes; role of shearing fractures in controlling slickrock topography and alcove development; entrenched valley fills (figs. 9, 13, 14, 36B, 59, 67, and 80).

Mile 22.5 [0] Return to NAV 16 and turn left (north), resetting mileage counter.

Mile 2.9 Brief STOP to look out over deep canyon in Navajo Sandstone. Small alcoves, but valley head not alcoved due to prominent fractures. Discussion question: what processes important in strongly joint-controlled headwalls? (fig. 67).

Mile 5.3 STOP at dry-canyon in Navajo Sandstone. Hike down to valley overlook. Discussion of evidence for seepage in alcoves, nature of sapping processes, development of alcoves, differences in morphology in strongly jointed versus massive sandstones, and role of sapping in strongly jointed sandstones. Discussion of large-scale features indicative of sapping. Backtrack to US 160. (figs. 69-70).


Mile 12.6 [0] Junction of US 160 and AZ 54D. Reset mileage counter.

Optional sidetrip. Turn left onto AZ 64. The road climbs a steep dip slope on top of Navajo Sandstone.

Mile 5.6 Top of monoclinal. STOP at solution pits developed on nearly flatlying plateau of Navajo Sandstone. Discussion theme: processes developing solution pits—factors controlling size—secondary jointing similar to mudcracks—possible role of eolian fill in larger weather pits—effect of surface gradient—"bead-ed" drainage and solutional grooves—relative amounts of rainwater loss through evaporation versus infiltration—role of solution pits to introduce through-flowing water into sandstone, contributing to sapping on sideslopes—why are solution pits developed strongly on some Navajo Sandstone exposures, whereas others nearly flat surfaces have none?

Mile 9.1 Optional continuation to Navajo National Monument with 1-hour roundtrip walk to Betlakin alcove overlook. "Return to US 160. (figs. 16, 11, 24, and 55)."

Mile 11.2 (1.2) [0] Junction of AZ 564 and US 160. Turn left onto US 160 and reset mileage counter.

Mile 0.5 Continuing on US 160, notice large sandslides in Cataraxus shales and sandstones to the right of the road for the next 5 miles.

Mile 4.8 Brief STOP to view beaded drainage and solutionally deepened rills on dip slope of Navajo Sandstone to the left. Discussion theme: relative roles of corrosional and solutional fluvial erosion in rills on slickrock slopes. (figs. 21 and 25).

Mile 9.0 Note entrenched valley fills of Laguna Creek on the left. The valley is not entrenched further upstream.


Day 2

Mile 2.4 STOP at entrenched fills of Laguna Creek. Note extensive piping and futile attempts at control using old automobiles, etc. Some structural grabens formed as a result of undermining by piping. Discussion theme: contrast the corrosional development of
piping with groundwater sapping processes in sandstone. Also note that piping seems limited to clay-rich alluvium, although sapping probably is important in valley-head extension in dissected sandy alluvium.

Mile 10.6 Escarpment of Navajo over Wingate over Chirile. Escarpment primarily backwashes by rockfall rather than alcoying due to pronounced vertical jointing in Wingate. Discussion theme: is some form of sapping important in such rockfall-wasted scarps?

Mile 17.7-18.8 STOP to discuss role of undermining in creating vertical slopes and alcoves in De Chelly Sandstone. Where the sandy Organ Rock tongue is not exposed at the escarp base, the De Chelly has typical slickrock morphology, but this is being eaten back into cliffs where underlying Organ Rock is exposed. Discussion of Ahlbert's climatic change hypothesis. (fig. 26)

Mile 27 Heart of Monument Valley. Possible picture stops. The "monuments" are composed of thin caprock of Shinarump Conglomerate (Chinle Formation) over a thin layer of shaly Moenkopi Formation over prominent vertical cliffs in the De Chelly Sandstone (Cotton Formation) and underlain by ramparts of Organ Rock member (Cutler Formation). Ahlbert suggests that the monuments are not actively backwashing due to a scarcity of debris from the De Chelly Sandstone. Most rocks on the Organ Rock ramparts are the very resistant Shinarump and are commonly on peduncles (damestools). Ahlbert suggests more active backwashing by spring sapping during the pluvials. Notice that small "mites" and the end of linear mesa produce less debris on the ramparts than along straight or indented portions of the scarp, due to a smaller area of cliff per area of rampart.

Mile 30.40 The road gradually passes downhill through Organ Rock Member, Cedar Mesa Sandstone Member, and Halgaito Tongue (shaly) of Cutler Formation and onto the Honaker Trail Formation (formerly the Rico Formation).

Mile 44.0 Bridge over the canyon of the San Juan River at Mexican Hat.

Mile 48.0 (0) Junction of US 163 and UT 261. Left turn onto UT 261 and reset mileage counter. View of dissected monocline flexure to the right exposing lower Cutler and underlying Hemmosa Formation.

Mile 4.9 Junction with road to overlook of San Juan River gorge necks. (Optional side trip to goose neck overlook.) Continue on UT 261.

Mile 7.4 Begin steep climb in valley headwall developed in Cutler Formation (Permian). Lower slopes in Halgaito Tongue, Organ Rock Shale, and caprock of Cedar Mesa Sandstone. The valley is only vaguely theater-headed, probably due to the thin-bedded nature of sandstone layers with shale interbeds and partings, which limits downward percolation.

Mile 8.9 STOP along the road in cliffs of Cedar Mesa Sandstone. Well-developed cavernous weathering and sapping in sandstones at contacts with underlying shaly units. Note salt concentration in crests in talus and evidence of active spalling. Sapping is probably important in backwashing of thin-bedded sandstones despite lack of alcoves and strongly theater-headed valleys. Note the paucity of evidence of active fluval erosion by plunge pools. The profile of the Cedar Mesa scarp is moderately convex, probably due to more rapid wasting at the top due to sapping processes in upper layers (groundwater is restricted from percolating lower due to shale interbeds).

Mile 9.1 Brief stop at overlook. View of monoline and Mexican Hat.

Mile 9.5 [0] Mulpy Point overlook turnout. Reset milege counter.

Mile 3.8 Time-permitting round trip to Mulpy Point. This side excursion can also be taken on the return portion of the trip. STOP at the overlook of Monument Valley and gorge necks of the San Juan River. Pronounced development of solutional pits, some very deep, on sandstones capping escarpment. Good slickrock morphology. Also backwashing of escarpment by calving of very large joint-bounded blocks. (figs. 18-20 and 31)

Mile 5.1 End of the road at Mulpy Point. STOP for magnificent views. Return to UT 261.

Mile 10.2 [0] Turn left onto UT 261. Reset mileage counter.

Mile 22.7 [0] Junction of UT 261 and UT 95. The high ridge to the north is underlain from bottom to top by upper Cutler (de Chelly), Moenkopi, and Chinle Formations (with cliffs in Moabback Member). The "Bear's Ear" at the crest of the escarpment are in Wingate Sandstone. Turn left onto UT 95 and proceed to Natural Bridges National Monument (NBNM). Reset mileage counter.

Mile 1.8 Turn right off UT 95 onto UT 275 toward NBNM.

Mile 18.3 Continue to NBNM and take the loop drive, returning to UT 95 and turning east. Discussion theme: well-developed slickrock and alcoving in canyon walls in Cedar Mesa Sandstone, sometimes resulting in natural bridges. Good illustration of most of the morphological features discussed by Oberlander. (fig. 38)

Mile 20.1 [0] Junction of UT 95 and UT 261. Continue on UT 95 and reset mileage counter. At approximately mile 9 cross the Comb Ridge Monocline and rise rapidly into Jurassic-Cretaceous strata.
Mile 29 [0] Junction of UT 95 and US 163. Turn north and reset mileage counter. The road from here to beyond Monticello generally lies on Cataractus Sandstone, locally modified by pediment and fan deposits from the Abajo mountains to the northwest. Local eolian mantling. Morrison Formation exposed in deeper valleys.

Mile 4 Blending

Mile 21 Monticello. Overnight stop.

Day 3

Mile 21 Continue north on US 163.

Mile 34 Road drops sharply off Cretaceous sandstones, through the Morrison Formation onto a flat plain developed on the Entrada Formation. Steep buttes in the Slick Rock Member of the Entrada Formation are visible on both sides of the road. These are often sculpted into deep alcoves and thin projecting fins.

Mile 39 [0] Junction of US 163 & UT 211. Reset mileage counter and turn left onto UT 211.

Mile 10.9 STOP at alcove along ephemeral stream in massive Navajo Sandstone. Active seep below stream in alcove. Ephemeral stream with weather pits or waterpockets in streambed. The stream bed in the area of waterpockets is steep and devoid of alluvium; although the bed is gravelly alluvium upstream. Discuss the relative roles of macroturbulence (kolkas) and solution in creating the waterpockets. Return to US 163 & 191. (Figs. 22 and 45A)


Mile 6.7 [0] Junction of UT 211 with the road to Needles Overlook. Reset counter and turn left onto Needles Overlook Road.

Mile 7.5 Pull off onto short dirt road leading to strongly alcove-exposed exposure of Entrada Sandstone after passing several examples of alcoves in the same formation on route. Well-developed talus and honeycomb weathering. Good examples of salt crustine and spalling. Good examples of surface wash and possible resulting case-hardening preventing talus development in otherwise ideal circumstances. Discussion topic: role of surface wash in developing alluvial weathering. (Optional 8.4-mile roundtrip to Needles Overlook). Return to US 191 & 163. (Figs. 50-52 and 60)

Mile 15.0 (23.4) [0] Junction with US 191 & 163. Turn left and reset mileage counter.

Mile 2 Casa Colorado Rock in Entrada Sandstone about 2 miles east of the highway (no sign). Pronounced alcove and fin development, with deep natural cisterns on top of rock. Gypsum may be precarious to the development of arches.

Mile 2.17 Good views of alcove and natural arches in Entrada Sandstone

Mile 7.5 Junction with UT 46. Continue on US 191 & 163.

Mile 17.8-33.4 Fault grabens and salt anticlines. Discussion of their origin and possible analogs to mar tian grabens.

Mile 33.4 Moab.


Mile 35.1 [0] Entrance to Arches National Park. Reset mileage counter. Turn right into park and visit Courthouse and Windows sections. Contrast of slickrock slopes on Navajo Sandstone with bladed, alcoves, and arches of Entrada Formation. Talus and spalling in incipient arches. Discussion of Oberlander paper—role of joints, shaly beds in landscape morphology. Comparison of slopes on Entrada and Navajo Formations. (Figs. 39, 40, and 76)

Mile 23.4 Return to US 163 & 191. Turn left onto US 163 & 191. Continue 2 miles to junction with UT 128. (Optional side trips to Dead Horse Point, Grandview Point, or potash workings.)

Mile 0.0 Turn left onto UT 128 and reset mileage counter.

Mile 9.35 The road follows the canyon of the Colorado River, which here flows across a broad alluvial plain to the Colorado River. The road crosses Navajo, Wingate, Chinle, Moenkopi, and Cretaceous Formations and then back upward through the section, eventually into the Cataractus sandstones and shales.

Mile 50 Junction with I-70. Turn east onto I-70. Built in Cretaceous sandstones and shales of Mesa Verde Group to the north of the road.


Day 4

Approximately 28-mile side trip through Colorado National Monument with good examples of theater-headed valleys in Wingate, Kayenta, and Entrada Formations. (Figs. 44 and 49) Return to I-70 and turn west, continuing past the junction with UT 128 (mile 58), past Green River (mile 111), Utah, to junction with UT 24.

Mile 143 Turn south onto UT 24. Pass through badlands in Morrison Formation. Note the spectacular monoclinical east limb of the San Rafael Swell to the west, exposing hogbacks in Navajo and Wingate Formations. Cross the San Rafael River. The road climbs onto broad desert with extensive eolian mantle on Er-
trada Formation (miles 160-180). Most of the sand is probably derived from this formation.

Mile 163 Pass the turnout to Goblin Valley. Goblin Valley is an excellent example of hoodoos in the Entrada Formation.

Mile 188 [0] As Hanksville is approached, note isolated buttes in the Entrada Formation and, to the west of the road, cliffs of the Summerville Formation capped by the lower part of the Morrison Formation. The laccolithic Henry Mountains are visible to the south. Reset mileage counter at Hanksville. At Hanksville continue west on UT 24.

Mile 12 After climbing escarpments in the Summerville Formation capped by Morrison (note occasional slumps), rounded badlands in variegated montmorillonitic shales of the Morrison Formation and prominent escarpment of Ferron sandstone over Tumunk shale (Cretaceous), turn left onto prominent terrace of the Fremont River. Good view of the Fremont River, three levels of pre-Wisconsinan gravel till terraces of the Fremont, and the 800 to 1000 foot high escarpment of the Emery Sandstone over Blue Gate shale (Cretaceous). Note the steep, linear-sloped badlands with knife-edged divides in the illitic Blue Gate shale. The badlands have formed from dissection of a pediment graded to the lowest prominent terrace, of Bull Lake age (Illinoisan). Some remnants of this pediment are visible. The Emery sandstone escarpment primarily backsways by rockfall and avalanching, but the backwasting is less active now than during the Bull Lake pluvial. To the west, the lava-covered plateaus of Thousand Lake and Boulder Mountain are visible. The latter was glaciated during the Bull Lake pluvial. (lgs. 35 and 71)

Mile 19 Continue west on UT 24. Just before the road crosses through the back of the Ferron Sandstone, pull off the road on the right and climb to the top of the prominent terrace of Bull Lake age. Good examples of the form and direction of badlands by dissection of the Bull Lake pediment and of dissected rockfalls of the same age on the ramps of the Emery Sandstone escarpment.

Optional roundtrip to Capital Reef National Park, following upstream along the Fremont River as it passes downward through the geologic column section through prominent eastward-dipping monoclines (the Waterpocket Fold). Excellent examples of talus in exposures of Navajo Sandstone near road level. Also good examples of domal topography in Navajo Sandstone with strong joint control (the “Capitals”). (lgs. 48 and 49)

Mile 38 [0] Return to Hanksville, reset mileage counter, and turn south onto UT 95.

Mile 11 Travel for several miles on exposures of the Entrada Formation mantled with eolian sands. Stop at the view of escarpments of the Summerville Formation capped with Morrison gypsum and sandstone to the west of the road. This escarpment has been largely inactive in backwasting since the last major pluvial. (lgs. 33, 73, and 74)

Mile 26 [0] Continue on UT 95, passing dissected pediments from the Henry Mountains and hoodoos in the Entrada Formation. Turn left onto UT 276 and reset mileage counter. Overnight stop at Ticaboo (mile 27) or Bullfrog (mile 40).

Day 5

Mile [0] Return to UT 95 and reset mileage counter. Turn south on UT 95. The road follows downstream in the canyon of North Wash, descending the geologic column through the Carmel, Navajo, Kayenta, Wingate, Chinle, and Moenkopi Formations, eventually reaching the Colorado River at Hite.

Mile 3.2 Stop along UT 95 and hike to the northeast to the head of a prominent theater-headed alcove with active seep (approximately 1.5 mile roundtrip). Note the impressive alcove, lack of plunge pool of small stream passing over scarp, and paucity of rockfall debris at the base of seep and alcove (as contrasted with abundant rockfall debris along valley sidewalls). Minor theater headwalls with largely dry alcoves are passed along the hike. (lgs. 43B and 56).

Mile 13.9 and 15.1 Continue southeast on UT 95. Turn off for scenic views of Lake Powell.

Mile 20.3 Cross the Dirty Devil River (also known as the Fremont River). The bridge is on the Cedar Mesa Sandstone.

Mile 22.3 Cross the Colorado River.

Mile 23.5 Hite Marina turnout. Continue on UT 95. Over the next 70 miles the road follows a tripped surface on the Cedar Mesa Sandstone. Bluffs to either side of the road expose, in ascending order, the upper Cutler Formation (Organ Rock Member), the Moenkopi Formation, the Chinle Formation, and, on the highest mesas, the Wingate Sandstone.

Mile 37.4 Overlook into canyon carved in Cedar Mesa Sandstone to the left, with good exposures of talus. (lgs. 53)

Mile 46.4 Settlement of Fry Canyon.

Mile 58.5 Junction with UT 263, the Halls Crossing Road. Continue on UT 95.

Mile 66 Junction with UT 275, the Natural Bridges National Monument road. Continue on UT 95.

Mile 67.8 [0] Junction with UT 261. Turn south onto UT 261, retracing an earlier part of the trip as far as the junction with US 163. Reset mileage counter.

Mile 15.8 Pass through Comb Ridge Monocline which successively exposes at the Cutler, Moenkopi, Chinle, Wingate, Kayenta, and Navajo Formations.

Mile 18.1 Junction with US 191. Turn south onto US 191. After crossing the San Juan River, the road lies on sand-mantled Navajo Sandstone, with the bluffs to the east exposing the Carmel, Entrada, Summerville, Bluff, and Morrison Formations.

Mile 47 Turn right (west) at the junction with US 160, passing through Mexican Water. The road still lies on Navajo Sandstone.

Mile 49.6 Turn left (south) onto US 191.

Mile 75.0 Alcoves in Navajo Sandstone are visible to the left. The road now drops in the stratigraphic section, exposing the Kayenta Formation, the cliff-forming Wingate Sandstone, and finally the Chinle Formation.

Mile 85.1 Junction with NAV 12 at Round Rock. Continue on US 191. The road from here to Chinle lies on the Chinle Formation, locally eroded into rounded badlands. Scraps in the Wingate Sandstone lie to the west of the highway, and Black Mesa, developed on Cretaceous sandstones and shale, forms the western horizon.

102.9 Junction with NAV 59 at Many Farms. Continue on US 191.


Day 6

Take 54-mile roundtrip from Chinle on NAV 7 through Canyon de Chelly, New Mexico, as far as Spider Rock overlook and return, stopping at all turnoffs and lookouts. Good views of alcoves, some created by sapping processes, and some by undercutting. Many have associated cliff dwellings. Some of the lower canyon slopes in the De Chelly Sandstone have approximately 30 to 40 degree slopes. Some are debris mantled, whereas others are nearly bare. The latter may be slopes formerly protected from erosion by a debris cover which is now eroded due to lack of cliff retreat. (fig. 75)

Mile 0 Returning to Chinle, reset mileage counter and continue south on US 191. The road continues on the Chinle Formation.

Mile 33.2 Triple junction of US 191 with AZ 264 and NAV 16. Turn southwest on NAV 16, which joins with NAV 15 in 13 miles. The road initially follows the canyon of Ganado Wash, exposing the Wingate, Moenave, Carmel, Entrada and Cow Springs Sandstone Formations on the northwest side, but then the road rises onto the Pliocene lake bed and tuff deposits of the Bidahochi Formation.

Mile 70.6 Junction of NAV 15 with AZ 77 at Bidahochi. Turn south on AZ 77. The road lies primarily on the Chinle Formation to the junction with I-40.

Mile 69.84 Tertiary volcanic flows and cones at the eastern edge of the Hopi Buttes are scattered over the landscape.

Mile 107.2 Junction with I-40. Turn west onto I-40.

Mile 207.4 Flagstaff at US 89 junction. End of trip.

REFERENCES


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