QUATERNARY STRATIGRAPHY AND LATE HOLOCENE FAULTING
ALONG THE BASE OF THE EASTERN ESCARPMENT OF STEENS MOUNTAIN,
SOUTHEASTERN OREGON

by

Mark A. Hemphill-Haley

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Approved by the Master's Thesis Committee

Chairman

Approved by the Graduate Dean
ABSTRACT

Two late Wisconsin pluvial high lake stands and at least one late Holocene faulting event have been recognized within the Alvord Valley along the base of the eastern range front of Steens Mountain in southeastern Oregon. The timing of the pluviations was estimated by correlation with the established chronology of Lake Lahontan about 80 km to the south. The older lake reached an elevation of 1310 m from approximately 24,000 to 16,000 yr B.P. A spillover and subsequent channel incision occurred during this lake stand at Big Sand Gap in the Tule Springs Rim on the eastern side of the valley. The lake level was lowered to 1280 m. The valley was occupied once again by a lake during a maximum pluviation from 16,000 to 12,000 yr B.P.

Large-scale geomorphic evidence of recent tectonic activity along Steens Mountain includes: 1) a steep, rugged range front escarpment accompanied by numerous landslides and knickpoints, 2) numerous geothermal springs, and 3) recently formed fault scarps.

A complex zone of high angle faults, the Steens Fault Zone (SFZ), is located along the base of the Central Steens. Numerous youthful appearing fault scarps are present within the SFZ. A segment of the SFZ, the Alvord fault, is expressed as a 19 km-long, north-northeast trending sharp-crested scarp. The scarp is the result of a single movement along the fault, probably as recently as late Holocene time. Evidence for the recent activity of the Alvord fault includes
displacement of the 1280 m Lake Alvord shoreline, a youthful scarp in
the lake playa that has been subjected to little wave modification,
and a basal graben formed during faulting that is preserved in an
older alluvial fan deposit. In addition, an analysis of the degraded
fault scarp provides further evidence that the most recent faulting
event occurred within the last few thousand years.

The SFZ is located within the transition between the Basin and
Range province and the Columbia River Plateau province. Faults
within the SFZ may be related to a north-northeast trending zone of
Holocene tectonic activity along the northwest margin of the Black
Rock-Carson Sink zone of extension in Nevada.
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INTRODUCTION

Investigations of the geology of the central, western and eastern Basin and Range physiographic province (Donath, 1962; Walker and Repenning, 1965; Stewart, et al., 1975; Lawrence, 1976; Hedel, 1980; Wallace and Whitney, 1984; Wallace, 1984; Hanks and Wallace, 1985; Minor, 1986; Minor, et al., in press a, b) have led to increased understanding of the Cenozoic stratigraphy and structure of that region. However, portions of the province remain to be investigated in detail regarding local Quaternary deposits and tectonic features, and their relationships to the better understood areas. Steens Mountain, in southeastern Oregon, is situated in the northwestern-most extent of the Basin and Range province (Lawrence, 1976; Figure 1). This investigation was concentrated along the eastern escarpment of the central portion of Steens Mountain and the adjacent western margin of the Alvord Valley. The objectives of the study included: 1) mapping and describing the Quaternary deposits along the range-front and western valley margin; 2) locating and mapping recent tectonic features and describing their association with mapped deposits; and 3) estimating the timing of the most recent movement along the principal range-front fault.

Location

The study area is located in southeastern Oregon between 43°N
STEENS FAULT ZONE

BASIN AND RANGE TECTONIC PROVINCE DIVIDED INTO ZONES OF LATEST FAULTING.
Pre-Late Quaternary: PLQ;
Late Quaternary: LQ;
Holocene: H.
After Wallace (1984)

Figure 1
Location map of Steens study area and other zones of recent seismicity.
and 42°15'N latitude and 118°15'W and 119°W longitude. Burns, the nearest population center, is 80 km northwest of the northernmost extent of Steens Mountain. Denio Junction, on the Oregon-Nevada border, is approximately 29 km to the south.

Geologic Setting

Four northwest-trending right-lateral strike-slip fault zones to the west of the study area have been identified to be the result of diminishing crustal extension relative to the central Basin and Range province (Stewart et al., 1975; Lawrence, 1976). One of the zones, the Brothers fault zone, terminates at the Steens Mountain scarp (Lawrence, 1976).

The eastern frontal face of Steens Mountain follows an average trend of N25E for approximately 100 km. Fuller (1931) described the range as comprising three segments: 1) the Northern Steens, which reach an elevation of approximately 2400 m; 2) the Southern Steens, also reaching approximately 2400 m; and the Central (or High) Steens which rise to 2967 m along a 1748-m-high escarpment.

Steens Mountain is a west-tilted upfaulted horst block (Figure 2). Minor (1986, personal communication) has recently proposed that the mountain is the result of listric normal faulting in a style similar to that discussed by Gans and Miller (1983). Steens Mountain has been subjected to at least two periods of late Wisconsin
Figure 2
Panoramic view of the eastern range front of Steens Mountain.
glaciation, as evidenced by deep glacial valleys along the north, south and western slopes, and by small cirques which are situated on the eastern side of the mountain crest (Bentley, 1970; Lund and Bentley, 1976). The mountain is composed of a thick sequence of late Oligocene (?) and Miocene basaltic flows, pyroclastic and sedimentary rocks (Davis, 1905; Minor et al., in press a, b). The 1000 m-thick Steens Basalts were accumulated possibly in as little as 2000 years (Gunn and Watkins, 1970). Remnants of ashflow tuffs which were deposited about 9.3 myr B.P. are preserved along the lower and middle western dip-slope of the mountain (Walker and Repenning, 1965), thus providing a maximum age for the initiation of major uplift along the mountain range.

Alvord Valley, which bounds Steens Mountain to the east, is a graben with a maximum width of 13 km. Shorelines cut into the surrounding mountain fronts and alluvial deposits attest to the presence of lakes that filled the valley to depths of up to 100 m in the late Pleistocene.

Previous literature

The Steens Mountain-Alvord Desert area has been the subject of little previous research. I. C. Russell (1884, 1903), Davis (1905), Waring (1908), Smith (1927), R. J. Russell (1928) and Walker and Repenning (1965) provided generalized descriptions of the area.

The volcanic stratigraphy of the Steens has been described by
Fuller (1931), Wilkerson (1958), Fryeberger (1959), Gunn and Watkins (1970) and Minor et al. (in press a, b).

The structure of the region was discussed by Donath (1962) and Lawrence (1976). Cleary et al. (1981 a,b), in addition to conducting gravity surveys in Alvord Valley, discussed the geochemical properties of thermal springs located near the mountain range. Most recently, Minor (1986) described the stratigraphy and structure of the Trout Creek Range to the southeast of the study area.
DEPOSITS

Tv - Tertiary rocks

The Tertiary volcanic stratigraphy has been described by Fuller (1931); Wilkerson (1958); Fryeberger (1959); Walker and Repenning (1965); Gunn and Watkins (1970); and Cleary et al. (1981 a,b). Minor et al. (in press a, b) have provided the most detailed and recent information about the Tertiary rocks exposed along the eastern escarpment within the Alvord Hot Springs and Wildhorse Lake 7.5' topographic sheets. Descriptions of Tertiary rocks in this report are derived primarily from this source.

The late Oligocene (?) to early Miocene Alvord Creek formation, the oldest unit in the study area, is located in the northern portion of the study area along the lower slope of the eastern escarpment. The relatively erosive tuffaceous sediments, opaline cherts and conglomerates attain a maximum thickness of 300 m. South of the mouth of Dry Creek they are buried by landslide debris and alluvium near the base of the escarpment.

The early Miocene Pike Creek formation consists of rhyolitic to acidic flows, minor tuffs and tuffaceous sedimentary rocks (Minor et al., in press a, b). It attains its greatest thickness, 600 m, in the southern part of the study area, but pinches out north of Little Alvord Creek.

The middle Miocene Steens Mountain Volcanics conformably overlies the Pike Creek formation to the south of Big Alvord Creek,
and unconformably overlies the Alvord Creek Formation to the north. Rocks within this sequence include interstratified andesitic flows, flow breccias and pyroclastic rocks. Minor et al. (in press a, b) report partially eroded cinder cones within the pyroclastic rocks. The location of maximum thickness for the Steens Mountain Volcanics, 1350 m, is north of Big Alvord Creek.

The Pike Creek formation and Steens Mountain Volcanics are disconformably overlain by the late Miocene Steens Basalt, a 1300 m thick series of olivine basalt flows with minor tuff interbeds (Cleary et al., 1981 a, b; Minor et al., in press a, b). K-Ar ages of 15.5 myr B.P. for an upper flow and 21.9 myr B.P. for a lower flow in the sequence have been reported (Minor et al., in press a, b). The Steens Basalts form the upper, precipitous part of Steens Mountain, including the crest. The combined Steens Basalts and Steens Mountain Volcanics form the majority of the eastern escarpment adjacent to the study area.

Welded tuffs of the late Miocene Danforth formation represent the uppermost Tertiary volcaniclastic rocks exposed on Steens Mountain. Although not exposed in the study area, the importance of the Danforth formation to this study is in the location of erosional remnants along the lower and middle western dip slope of Steens Mountain which show only a minor discordance with the underlying Steens Basalt (Bentley, 1970). One member of the formation, the Devine Canyon Ash Flow Tuff, has a reported K-Ar age of 9.3 myr B.P. This establishes a maximum age of the inception of uplift of
Steens Mountain (Walker and Repenning, 1965; Minor et al., in press a, b).

Quaternary Deposits

Quaternary deposits were initially studied and unit names assigned on the basis of aerial photographic interpretation. Preliminary geologic contacts were then verified by field reconnaissance mapping.

Older lacustrine deposits - Qld₁

These are light to medium brownish-gray lacustrine silty sands and medium to coarse-grained sand and gravels, exposed south of Indian Creek along the lower flank of the Steens Mountain range front between 1310 m and 1280 m elevation. Bedding is generally well developed and nearly horizontal. Individual beds are up to 4 cm thick and commonly exhibit cross-bedding. These coarse-grained sediments were deposited in a relatively high energy beach environment of pluvial Lake Alvord at the base of the mountain front. They are the highest-elevation lake sediments, and represent the oldest exposed lacustrine deposits in the study area. The finer-grained silts and clays of the distal lake facies have been buried by later lake deposits below an elevation of 1280 m. The uppermost extent of these deposits, at 1310 m, is accompanied by a poorly-
developed shoreline. A colluvial veneer, up to 25 cm thick, of poorly-sorted angular volcanic clasts caps the beach deposits in some locations closest to the steep range front, but thins away from the escarpment.

Older barrier bar unit - Qbb₁

These deposits consist of moderately to well-stratified, poorly to moderately-sorted alternating sands, silts and gravels, with bedding dipping up to 30 degrees to the northeast. Gravel and coarse sand beds are moderately to well cemented by CaCO₃, while finer grained sands and silts are poorly indurated. Fine grained sediments are medium to light gray while coarse grained deposits are light gray with a reddish-yellow silty-sand matrix.

These sediments are located within an arcuate, low relief barrier bar situated about 4.5 km south of the mouth of Wildhorse Canyon. The bar is 0.7 to 1.0 km wide at a maximum elevation of 1280 m, 14 m above the surrounding lake deposits. The barrier bar has undergone a relatively long period of modification by erosion, as evidenced by its low relief, rounded crests, and slope angles of less than 5 degrees. As further evidence, Wildhorse Creek has dissected a broad channel more than 300 m wide but less than 9 m deep across the width of the barrier bar. The channel banks have been degraded to low gradient slopes of less than 5 degrees.

This barrier bar is the result of deposition of a large volume
of sediment into the low-gradient beach environment associated with the lacustrine deposits (Qld$_1$) of pluvial Lake Alvord. It was probably situated with its crest at approximately stillwater (half-wave) level, or just slightly above that point, as evidenced by its broad, planar summit (Price, 1968). The bar also may have been formed as a submerged longitudinal bar, being subsequently planated during a lower lake stand (see below: Pluvial History).

Older alluvial fan deposits - Qaf$_1$

These deposits occur as erosional remnants of alluvial fans along the eastern margin of the range front at two locations: 1) north of Indian Creek and 2) south of Pike Creek. They are topographically higher than younger alluvial fans. The deposits are composed of poorly-sorted, moderately-stratified, matrix-supported angular to sub-angular volcanic sands and gravels up to 2 m in diameter. They are moderately cemented by CaCO$_3$ and illuvial clays.

The fan remnants are characterized by a subdued, rounded surface and are overlain at the Pike Creek site by landslide deposits of unit Q1s. The distal portion of the fan at Indian Creek has been modified by wave erosion and overlain by intermediate lake deposits (Qld$_2$) up to an elevation of 1280 m.
Intermediate lake deposits - Qld$_2$

The intermediate age lake deposits consist of medium to light brownish-gray silty to medium and coarse-grained sands and gravels, grading to fine sands, silts and clays away from the range front. Bedding is well developed, ranging in thickness from a few centimeters near the outer edges of the deposit to less than a centimeter in exposures closer to the interior of the basin. Bedding planes dip up to 4 degrees, varying in strike from northeast to southeast near the southern end of Alvord Point, where approximately 4-5 m of lake sediments have been exposed by erosion (Figure 3). The upper extent of these deposits at 1280 m coincides with the prominent shoreline formed in the older Qld$_1$ unit. Below 1222 m they are buried by the most recent lacustrine sediments. Exposures are found as far north as Tuffy Creek, after which they are buried by alluvial fan deposits.

Younger barrier bar deposits - Qbb$_2$

The younger barrier bars consist of deposits of well stratified, moderately sorted, light to medium gray sands, silts and gravels. Bedding planes dip between 54 and 40 degrees to the northwest. Beds up to 1 m thick are commonly cross-bedded. Coarse grained beds are slightly cemented by CaCO$_3$ while finer grained sands are poorly indurated.
Figure 3

Intermediate lake deposits (Qld$_2$) exposed near the southern end of Alvord Point.
These deposits form distinct elongate, arcuate or straight mounds, up to 500 m wide and 3 km long, that are located near the mouths of major sediment sources such as Big Alvord Creek, Cottonwood Creek, Pike Creek and the narrow northern Alvord Valley. In most cases the range-facing limbs of the bars have been partially buried by alluvium or gently slope back toward the range front (Figure 4). In contrast, the valley-facing slopes are steep and stair-stepped by recessional shorelines. Maximum heights of the bars relative to adjacent lake deposits vary from 3 m near the range front in the northern part of the valley to 18 m at locations away from the range front. The largest of these are the Cottonwood Creek and Pike Creek barrier bars. At locations generally adjacent to the range front or near the mouth of a creek (e.g. between Pike Creek and Little Alvord Creek), only remnants of a bar exist, having been almost completely buried by alluvium or eroded.

In contrast to the older barrier bar (Qbb1), these bars appear morphologically youthful, with summits bounded on the valley-facing side by sharp-crested, steep slopes inclined up to 18 degrees. Steep-walled channels (up to 20 degrees) are formed by stream dissection. Barrier bar summits consistently terminate at 1280 m, forming continuous planar surfaces. The exception is in locations where burial or extreme erosion has taken place. This 1280 m elevation is consistent with the shoreline elevation and upper extent of the intermediate lake deposits (Qld2).
Figure 4
Eastern face of barrier bar (Qbb$_2$) at the mouth of Pike Creek. Intermediate alluvial fill deposits (Qaf$_2$) appear in middle background of photograph. Undifferentiated alluvial fan and intermediate lake deposits (Qfu/Id$_2$) appear in foreground.
Landslide deposit - Qls

Landslide deposits consist of unsorted, non-stratified volcanic clasts derived primarily from poorly-consolidated, relatively erosive, tuffaceous sediments of the Alvord Creek and, to a lesser degree, Pike Creek formations. The sediments commonly exhibit a clast-supported texture.

The landslides are commonly expressed as slightly sinuous, elongate features with narrow upper reaches, broad lower expanses and lateral margins that are dotted by numerous cold springs. The slide surfaces are hummocky and commonly contain closed depressions, frequently containing water. The landslides are predominantly found within the canyons of the principal drainages north of Tuffy Creek. To the south, the Steens Basalt is the predominant rock type and much less susceptible to failure. Several of the landslides cover areas up to 2 km long and 500 m wide.

Generally, landslide activity in this area is indeterminable. However, a landslide at Indian Creek extends below the 1280 m intermediate lake stand shoreline, is void of evidence of wave modification, and thus indicates a post-pluvial lake age. In addition, a large, fresh head scarp in the massive slide mass north of Big Alvord Creek observed during the field reconnaissance in 1984 (Figure 5) was reported by Mr. Ed Davis, (owner of the Alvord Ranch), as not being present the previous year. Toe segments of the majority
Figure 5

Photograph of a landslide headscarp which was formed in 1984 (arrow) north of Big Alvord Creek. Note boys for scale.
of the landslides, with exception of the Indian Creek locality, have been truncated by alluvial fans.

Undifferentiated alluvial fan and lake deposits - Qafu/ld₂

Less than 1 km north of Alvord Hot Springs, medium to fine sand and silt of the distal portions of alluvial fans overlie and are interbedded with silts and clays of the intermediate lake deposits (Q₁ld₂). These undifferentiated deposits are typically concealed by vegetation and marshes. Several trenches, approximately 2 m deep, were excavated parallel to the range front in locations progressively closer to the center of the valley, beginning at about 2 km east of the mouth of Big Alvord Creek. The trenches revealed alternating coarse sands and silty clays, with the fine-grained sediments predominating valleyward. Water from infiltration of channelized surface runoff was confined to and flowed from the coarse layers. The presence of numerous alternating sand and silt layers probably indicates minor short-term fluctuations in the level of the pluvial lake.

The interlayered alluvial fan and lake deposits are not exposed above 1280 m, coinciding with the upper extent of the intermediate lake deposits (Q₁ld₂).

Intermediate alluvial fan deposits - Qaf₂

These deposits consist of light yellowish-red to yellowish-gray poorly-sorted, sub-angular to sub-rounded clasts ranging in size from
silty sand to boulders. Basalt boulders up to 5 m diameter litter the fan surface, and are common components of the deposit near the mouths of streams. A matrix-supported texture predominates near fan apexes; a clast-supported texture is more common in the distal portions of the fans. Lithologic composition and aerial extent of the fans vary from south to north within the study area. From Alvord Point to Tuffy Creek, the fans are small and composed almost entirely of basalts and associated rocks of the relatively resistant Steens Basalts. North of Tuffy Creek the fans are larger, extending much farther from the range-front, and have a higher percentage (up to 50%) of clasts from the more erosive Alvord Creek formation, Pike Creek formation and Steens Mountain Volcanics. The increase in fan size to the north can also be attributed to the larger catchment basins of the Central Steens block.

These deposits exist as moderately dissected, relatively youthful-appearing alluvial cones (Figure 6). Small, non-integrated cobble-filled channel segments cross all portions of the fan surface, and are probably active during high discharge storm events. The majority of alluvial fans south of Tuffy Creek emanate from small, closely-spaced drainages, and coalesce near their apexes. The fans north of Tuffy Creek, associated with larger, more widely-spaced drainages, are large, more discrete and coalesce at their distal extents.

In the central and northern part of the study area the
Figure 6

View to the north of the intermediate alluvial fan deposits (Qaf₂) of Little Alvord Creek (arrow). Intermediate barrier bar deposits (Qbb₂) located in the foreground.
intermediate alluvial fan deposits overlap and partially bury the younger barrier bar \((Qbb_2)\) and intermediate lake deposits \((Qld_2)\).

Younger alluvial fan and channel deposits - \(Qaf_3\)

These deposits consist of yellowish-gray to gray poorly-sorted, sub-angular to sub-rounded clasts ranging in size from silty-sand to small boulders. A clast-supported matrix is characteristic of these deposits. Clast lithologies are similar to the intermediate alluvial fan deposits, with predominantly basalt of the Steens Basalts in the south, to up to 50% tuffaceous sediments north of Tuffy Creek. The deposits are generally unweathered and non-indurated.

The sediments are found in active channels and fan lobes. The channels generally incise the intermediate alluvial fan deposits \((Qaf_2)\) and continue valleyward where they deposit the younger sediments as fan lobes. These deposits are found in centralized channels associated with the perennial streams of the Central Steens and the seasonal streams south of Tuffy Creek. In addition, they are found as layers less than 50 cm thick in small, shallow distributaries that radiate across the surface of the intermediate alluvial fan deposits.

Thermal Spring Deposits - \(Qts\)

These deposits consist of light gray to white fine silts and
clays that form a powdery veneer overlying intermediate lake deposits (Q1d₂). A large fan-like plume of these sediments is located at Alvord Hot Springs. The plume emanates from a fault scarp at the base of the mountain range and continues for a distance of 2 km to the present lake shoreline. It extends a distance of 1.5 km at its greatest width.

Younger lake and playa deposits - Q1d₃

These deposits consist of light-gray coarse sands and gravels along the beaches associated with the present lake in the Alvord Desert that has persisted since 1982. Sediments from the center of the lake, sampled from a canoe, are comprised of light-brown to light-gray fine silts and clays.

These lake sediments are confined to a maximum elevation of 1222 m, below the present lake shoreline, and cover almost 130 square km of the valley floor.

Sand dunes - Qsd

Dunes consisting of coarse to medium grained light-brown unconsolidated sands are found primarily in the southern and southwestern portion of the study area. They generally overlie intermediate lake deposits (Q1d₂) on the valley floor. They occur as dune fields covering areas up to 6 km² along the southern margin of the present Alvord Lake, and as large individual concentric dunes in
the eastern part of the valley.

Based on artifacts and $^{14}$C dates, Mehringer and Wigand (in press) cited dune stabilization along the eastern margin of the Catlow Valley (at the western edge of Steens Mountain) at 7200 yr B.P. Assuming similar climatic and physical conditions for nearby Alvord Valley, this age might be used as an initial estimate for the period of dune stabilization in the present study area.
PLUVIAL HISTORY

Pluvial lakes respond to changes in climatic conditions such as precipitation, evaporation and runoff, however the relationship between lake level and the ratios of these three principal variables is not clearly understood. In addition, it is believed that pluviations result from the same climatic conditions responsible for glaciations, but it is unclear what the response time is for pluviations or glaciations to the changing climatic conditions. Indeed, considerable debate has risen as to the timing of past pluviations in relation to the timing of glaciations (Benson, 1981, 1978; Brakenridge, 1978; and Snyder and Langbein, 1962). Therefore, these problems were not addressed in this study, but in order to estimate the timing of the most recent lake stands, the assumption was made that pluvial Lake Alvord acted in a manner similar to the much-studied lake systems of Lahontan and Bonneville. In those basins, pluviations have been assigned age designations which reflect the closest temporal period of glaciation (e.g. late Wisconsin high stand).

Evidence of pluvial Lake Alvord

The Alvord Basin was occupied in the late Pleistocene by pluvial lakes. Evidence for ancient lake occupation includes: 1) stationary and recessional shorelines etched high into the sides of the mountain range; 2) gravel bars deposited near the range front; 3) sand spits
near the gravel bars; and 4) a wind gap, Big Sand Gap, which was carved into the Tule Springs Rim on the east side of the valley. This happened when water from the highest lake stand overtopped a low point in the mountains, and downcut through the rocks as it flowed eastward into the adjacent Coyote Lake valley.

Pluvial events

At least three periods of lake occupation occurred within the Alvord Basin. They include: 1) an older, possibly late Wisconsin, stand; 2) an intermediate lake stand; and 3) the historic lake stand present during the course of this study.

Older lake event

The older lake stand has been recognized in the Alvord Valley from the following evidence: 1) a poorly defined wave-cut bench located south of Tuffy Creek and along Alvord Point at an elevation of about 1310 m elevation (Figure 7); 2) a large eroded barrier bar situated south of the mouth of Wildhorse Creek; and 3) lake deposits located directly below the 1310 m shoreline.

The shoreline is the highest pluvial feature found in the study area. It exists as a barely perceptible wave-cut notch which crosscuts bedding in the Steens Basalts, most notably around the perimeter.
Figure 7

View to the south of the 1310 m shoreline notch visible along the skyline of the eastern edge of Alvord Point. The more prominent 1280 m shoreline is also visible.
of Alvord Point (Plate 1). It generally lacks coarse grained deposits such as gravels and cobbles. This evidence, in conjunction with the poorly inscribed shoreline, indicates that the lake was maintained at this elevation for only a brief period of time.

The large barrier bar is 3.2 km long and up to 1 km wide. It is expressed as a broad, slightly arcuate mound surrounded by an apron of lacustrine sediments. Dissection by Wildhorse Creek, probably shortly after or during dessication of Lake Alvord, has left a 300 m wide channel through the barrier bar. This channel displays degraded side walls and a broad flood plain, characteristic of relative maturity.

The crest of the barrier bar at 1280 m is angular and planar, and appears youthful relative to the morphology of the rest of the bar. The crest is probably the result of wave modification by a lower lake stand than the one that was present during deposition of the barrier bar.

A wind gap, Big Sand Gap, located in the Tule Springs Rim which bounds the eastern portion of the Alvord Basin, was identified during the field reconnaissance (Figure 8). A gentle sag is formed in the Tule Springs Rim, which reaches a lower elevation of 1372 m near the gap. However, the top of the gap itself is at an elevation of only about 1310 m, which coincides with the elevation of the highest old shoreline. Big Sand Gap formed when ancient Lake Alvord filled to the point of overflow at the structural sag existing in the Tule Springs Rim, and water spilled eastward into the valley containing
Figure 8
The wind gap, Big Sand Gap, can be seen as a notch in the Tule Springs Rim along the eastern side of the Alvord Valley. An intermediate barrier bar deposit (Qbb2) appears in the foreground.
Coyote Lake. The gradient in the gap channel supports continued eastward flow as downcutting progressed. The elevation of the channel at the base of the gap defined the maximum limit of subsequent lake levels. The gap was incised down to an elevation of 1280 m, approximately 55 m above the present valley floor. This elevation is equal to that of the planar crest of the older barrier bar deposit. Incision of the gap may have been rapid, as evidenced by the poorly-developed shoreline at 1310 m and faint recessional shorelines down to 1280 m.

Intermediate lake event

The maximum elevation of subsequent lake stands in the Alvord Valley was controlled by the 1280 m base of the Big Sand Gap channel. Evidence for later lakes includes: 1) morphologically younger gravel bars (Qbb₂) with sharp crests and upper surfaces at 1280 m; and 2) a well-developed morphologically-youthful shoreline at 1280 m which is cut into the older barrier bar deposits (Qbb₁) (Figures 9 and 10). Indirect evidence supporting this lake stand is derived from the Lake Lahontan pluvial record which indicates the presence of a sustained major lake occupation in that basin (Benson, 1978). The barrier bars associated with the intermediate lake event (Qbb₂) can be found along the western margin of the valley floor adjacent to the Central Steens, a source of large volumes of sediment for the lake margin.
Figure 9
Prominent 1280 m shoreline along eastern flank of Alvord Point. The 1310 m shoreline is barely visible above it.
Figure 10
Composite air photographs of Alvord fault and other members of the Steens fault zone. Broad arrows indicate Alvord fault. Smaller arrows point to other young fault scarps.
Lacustrine silts and clays, at least 4.5 m thick, were deposited away from the range front near the southern end of Alvord Point. Prominent recessional shorelines were formed as the lake level slowly dropped during the decline of the intermediate pluvial event. These shorelines can be easily identified from aerial photographs at the north end of Alvord Point, around Serrano Point and along the eastern flank of the younger barrier bar (Qbb₂) east of Cottonwood Creek (Figure 10). The recession of the lake was also accompanied by progradation of alluvial fans. The oldest alluvial deposits found along the eastern range front which bury the 1280 m intermediate lake shoreline are the remnants of the older alluvial fan deposits (Qaf₁) located at Indian Creek and Pike Creek. Successively younger fans (Qaf₂ and Qaf₃) have further obscured shoreline features of this lake stand, and have partially buried the barrier bars located east of the Central Steens.

Historic lake stand

The Alvord Desert has been partially inundated by water since 1982. By 1985 the lake was 3 m deep, having reached an elevation of 1222 m. Several long-time Alvord Valley residents reported this to be the highest level of the lake in 90 years. This increase in lake level since 1982 coincides with the similar recent rising of other Basin and Range pluvial lakes (e.g. the Great Salt Lake, Mono Lake, and Warner Lakes).
Figure 11
Chronologies for three Pleistocene lakes in the western United States:
A) chronology for Lake Alvord presented in this paper (rising and falling limbs are speculative and based on Lahonton chronology);
B) chronology for Lake Lahonton (Thompson et al., 1986);
C) chronology for Lake Bonneville (Scott et al., 1983).
Correlation with Lake Lahontan

Benson (1978), Scott (1983) and Thompson et al. (1986) have provided strong evidence that the lake levels of the two largest pluvial lake complexes of the Great Basin province (Lake Lahontan and Lake Bonneville) rose and fell in generally synchronous manners (Figure 11). This indicates that climatic and hydrologic regimes can be considered similar across the width of the Great Basin, and that the physical characteristics of the individual basins controlled the magnitude of the lake levels.

The assumption is made that the chronology of lake stands for Lake Alvord is probably similar to that of Lake Bonneville, but even more so to that of Lake Lahontan, whose northernmost extent lay about 80 km south of the study area. Therefore, the well-documented pluvial history of Lake Lahontan was used to explain the behavior of Lake Alvord.

Benson (1978), Davis (1983) and Thompson et al. (1986) have reported that Lake Lahontan was at an intermediate and probably rising stage of pluviation from about 24,000-16,000 yr B.P. based on tephrochronology and $^{14}$C dates from wood debris and tuff. During this period, Lake Alvord was presumably also at an intermediate stage with an overall rising trend (Figure 11). It is impossible to determine exactly when the spillover event occurred, but it probably happened in the interval 24,000-16,000 yr B.P. as the lake was
rising. The lake rose slowly toward the crest of the sag in the Tule Springs Rim as evidenced by the large size of the barrier bar (Qbb1), at the mouth of Wildhorse Creek which required an extensive period of deposition. Therefore, it might be assumed that the spillover event took place at the latter part of the rising 24,000-16,000 yr B.P. interval. Refinement of the age of this event is not possible without further evidence.

Following the spillover event the lake rapidly declined to 1280 m, a 30 m drop in lake level. In their study of Lake Bonneville, Scott et al. (1983) cited evidence of a rapid decline of the lake following the overflow at Red Rock Pass that included the absence of recessional shorelines between the Bonneville and Provo shorelines. There are only very faint recessional beach features in the Alvord Valley between the 1310 m and 1280 m shorelines.

In the Lahontan basin, the period from 16,000 to 12,000 yr B.P. is associated with the highest stand of the lake (Thompson et al., 1986). Although the climatic conditions necessary for a high lake stand would have also existed in the Alvord Basin, Big Sand Gap prevented a rise in lake level.

Again employing a correlation with Lake Lahontan (Benson, 1978; Davis, 1983; Thompson et al., 1986), the stand of Alvord Lake which had been maintained at 1280 m began to decline rapidly at about 12,500 yr B.P., possibly diminishing almost completely by 10,000 yr B.P.
LATE QUATERNARY TECTONICS

The northwest portion of the Basin and Range physiographic province within the southeastern Oregon has been regarded as an area of tectonic quiescence for the late Quaternary (Wilkerson, 1958; Thenhaus and Wentworth, 1982). A lack of seismicity and little physical evidence of youthful fault-related surface features have been cited as the reasons for this conclusion. This study provides evidence of late Quaternary faulting along the eastern front of Steens Mountain, with the latest major surface rupture event occurring in the late Holocene.

Long-Term Uplift

Gravity investigations (Cleary et al, 1981 a, b) indicate that the Steens Basalts occur as basement rocks in the western Alvord Valley adjacent to the Central Steens, at a depth of about 1000 m below the playa surface. The basement is overlain by about 1000 m of valley-fill alluvial and lake sediments. When added to the 1748 m difference in elevation between mountain crest and valley floor, a total of 2748 m of displacement along the eastern escarpment of the range is calculated. If uplift was initiated near the time of deposition of the Danforth formation (9.3 myr B.P.), then a minimum
long-term uplift rate of approximately 0.29 mm/yr can be calculated. However, by projecting a line along the erosional surface of the Steens Basalt, at the top of the mountain, toward the east to a point which intersects the projected frontal fault of Steens Mountain, an actual displacement of 3131 m is calculated. Assuming that the initiation of uplift was the maximum 9.3 myr B.P., a long-term slip rate of 0.33 mm/yr is derived (Figure 12). If in fact the inception of uplift was more recent than 9.3 myr B.P., the long term uplift rate would be higher.

In summary, the maximum total displacement along the Steens fault zone is 3131 m. The precise timing for the inception of uplift is not known, but assuming the Danforth formation was uppermost unit prior to uplift, its 9.3 myr B.P. age can be used as a maximum estimate for the beginning of uplift.

Geomorphic and Associated Evidence of Recent Tectonic Activity

Large-scale geomorphic evidence of recent tectonic activity is present along the Steens Mountain range front. This includes: 1) the morphology of the range-front scarp; 2) presence of springs; and 3) faults. Although these geomorphic features do not provide a quantitative measure of the rate of activity, they do allow qualitative assessment of the recency and magnitude of tectonic movement in the area. In addition, regional and local seismicity
Cross-section of Steens Mountain and the Alvord Valley near Little Alvord Creek. Projected and present summits of the mountain are shown at 3350m and 2967m respectively. Basalt basement located about 1000m below playa surface (Cleary et al., 1981 A & B). Long-term slip-rates calculated from estimated displacements along range front fault since inception of faulting approximately 9.3mya.
provide information on tectonic activity of an area. Only recently has it been postulated that areas devoid of pronounced seismicity are not necessarily inactive, but rather are possible sites of relatively long recurrence intervals for seismic events related to regional tectonic clustering of earthquakes (Wallace, 1984). This is characterized by shifts in seismicity and surface faulting from one area to another in response to a non-uniform stress field. Therefore, it is important to consider the seismicity of the surrounding region with respect to the local geologic and seismic conditions.

Range-front scarp

The eastern frontal escarpment of Steens Mountain is a rugged, steep feature formed in relatively resistant basalts and erodable volcaniclastic sediments that rises 1748 m above the Alvord Valley in a distance of 8 km (Figure 13). Incision by east flowing creeks with small drainage basins has formed deep, precipitous canyons. Numerous knickpoints, represented by waterfalls and boulder runs, are located along the courses of these drainages. Many of these knickpoints are formed as a result of differential erosion of various rock types in the mountain front. However, the large number and distribution of these features implies that uplift of the range has been too rapid for stream gradients to achieve equilibrium. The lower elevations of
Figure 13
View to the north of the rugged eastern escarpment of Steens Mountain from Alvord Point.
the range front north of Tuffy Creek feature numerous large landslides up to 2 km long and 0.5 km wide. These landslides, which originate in the highly-erodable tuffaceous sediments of the Alvord Creek formation, have created a hummocky, ponded topography (Figure 14).

Springs

The Alvord Valley has been the site of investigation for geothermal development. Three major thermal springs (Mickey Hot Springs, Borax Springs, and Alvord Hot Springs) have been studied for their thermal energy potential and geochemical composition (Cleary et al., 1981 a, b). They maintain a relatively constant discharge with only minor seasonal fluctuations. Borax Springs, near Alvord Lake, is at the center of the narrowest part of the valley. Mickey Hot Springs discharges from the eastern front of Mickey Buttes, a narrow north-trending horst located in the northern portion of the Alvord Valley. Alvord Hot Springs flows directly from a fault scarp along the base of the range front and out onto the adjacent playa (Plate 1; Figure 15). Gravity surveys conducted by Cleary et al. (1981) located a large gravity low near Alvord Hot Springs which they concluded to be a fault zone. They also suggested, based on geochemical analysis, that Alvord Hot Springs is the result of meteoric waters that have percolated through the volcanic rocks of Steens Mountain, heated at
One of many large landslides found near the base of the Central Steens. This one is located to the south of Big Alvord Creek.
Figure 15
Alvord Hot Springs flowing from the scarp base of the Alvord Fault. Road is built on the scarp original surface. Arrows point along trend of fault.
depth by an elevated geothermal gradient and recirculated along the fault zone.

Cold springs are also common along the range front. Many of these are located within or at the base of the numerous landslide masses, and are probably the result of meteoric water moving through the volcanic rocks within the range and becoming concentrated along the base of the less permeable landslide material. Other springs are found along the eroded margins of alluvial fans, and may be controlled by fan stratigraphy. Some springs are distinguishable only by an increase in vegetation, while others maintain an appreciable flow.

Seismicity

Seismic activity in the Alvord Valley can be characterized as historically quiescent with no instrumentally-located epicenters being reported. Valley residents near Borax Springs reported a small swarm of tremors in 1918 (Cleary et al., 1981 a, b), but shaking was not felt in Fields, 11 km to the south. However, a significant increase in discharge was reported to have occurred at Borax Springs at that time (Cleary et al., 1981 a, b).

A swarm of earthquakes was reported in the Warner Valley, about 135 km east of the study area, in 1968 (Couch and Johnson, 1969). The largest earthquake was reported to have a 5.1 Richter
magnitude. Twelve other events with magnitudes greater than 4.0 occurred within four days. Couch and Johnson (1969) reported that eight of the events had first motions caused by normal west-side-down movement along a north-south trending fault. Two events had first motions consistent with strike-slip movement on an obliquely-trending structure possibly related to northwest-trending strike-slip faults of Stewart et al. (1975) and Lawrence (1976).

Steens Fault Zone

During this investigation, a complex system of normal faults was mapped (Plate 1) and named the Steens fault Zone (SFZ). These faults are expressed at the surface as predominantly east-facing scarps that displace late Quaternary lake and alluvial fan deposits, lake strand lines and Tertiary volcanic rocks (Plate 1). The most prominent faults have been assigned informal names that refer to geographic names (Plate 1).

The area where late Quaternary displacement of the SFZ is most pronounced lies almost entirely within the southern portion of the study area, to the southeast of Wildhorse Canyon. An exception to this is the Alvord fault, which can be traced northward along the range front to the mouth of Dry Creek Canyon (Plate 1). Fault segments of the SFZ trend in three directions and have been grouped accordingly: Group 1 - predominantly north, with east-facing scarps; Group 2 - northwest, with northeast-facing scarps; and Group 3 - east-northeast, with northwest facing scarps (Table 1).

The following fault scarp relationships exist within the SFZ:
<table>
<thead>
<tr>
<th>Name</th>
<th>Location</th>
<th>Length (km)</th>
<th>Trend</th>
<th>Average Trend</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alvord fault</td>
<td>T34S,R34E, sec 28,29,32</td>
<td>19</td>
<td>N18W to N12E</td>
<td>NON</td>
</tr>
<tr>
<td></td>
<td>T35S,R34E, sec 4,5,8,17, 20,29,32</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>T36S,R34E, sec 5,8,17</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Unnamed</td>
<td>T36S,R34E, sec 17,20</td>
<td>2.0</td>
<td>N20W to N05W</td>
<td>N15W</td>
</tr>
<tr>
<td>Unnamed</td>
<td>T35S,R34E, sec 7</td>
<td>1.1</td>
<td>N18E</td>
<td></td>
</tr>
<tr>
<td>Unnamed</td>
<td>T35S,R34E, sec 7</td>
<td>0.6</td>
<td>N15E</td>
<td></td>
</tr>
<tr>
<td>Unnamed</td>
<td>T35S,R34E, sec 7</td>
<td>0.6</td>
<td>N19E</td>
<td></td>
</tr>
<tr>
<td>Smyth Wells fault</td>
<td>T35S,R34E, sec 30,31</td>
<td>7.7</td>
<td>N18E to N10W</td>
<td>NO4W</td>
</tr>
<tr>
<td></td>
<td>T36S,R34E, sec 6,7,18</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Serrano Springs fault</td>
<td>T35S,R34E, sec 31</td>
<td>1.6</td>
<td>N13W to N55E</td>
<td>N25E</td>
</tr>
<tr>
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<td>T36S,R34E, sec 6</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Serrano Point fault</td>
<td>T35S,R34E, sec 30,31</td>
<td>3.9</td>
<td>N54W to N29W</td>
<td>N50W</td>
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<tr>
<td></td>
<td>T35S,R33E, sec 25</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Name</td>
<td>Location</td>
<td>Length (km)</td>
<td>Trend</td>
<td>Average Trend</td>
</tr>
<tr>
<td>-------------------</td>
<td>--------------</td>
<td>-------------</td>
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<td>---------------</td>
</tr>
<tr>
<td>Embayment fault</td>
<td>T35S,R34E</td>
<td>2.4</td>
<td>N75W to N62W</td>
<td>N71W</td>
</tr>
<tr>
<td></td>
<td>sec 30,31,32</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>T35S,R33E</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>sec 25</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kueny Ditch fault</td>
<td>T35S,R34E</td>
<td>4.0</td>
<td>N68E</td>
<td>N68E</td>
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<tr>
<td></td>
<td>sec 20,29,30</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>T35S,R33E</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>sec 25,35,36</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Unnamed</td>
<td>T35S,R34E</td>
<td>1.0</td>
<td>N51E</td>
<td>N51E</td>
</tr>
<tr>
<td></td>
<td>sec 19,30</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wildhorse Creek fault</td>
<td>T35S,R33E</td>
<td>0.5</td>
<td>N54W</td>
<td>N54W</td>
</tr>
<tr>
<td></td>
<td>sec 24</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
1) Group 3 faults are truncated by Group 1 faults (e.g. the Kueny Ditch fault is truncated by the Smyth Wells fault); 2) Group 3 faults are also displaced by Group 2 faults (e.g. the Kueny Ditch fault is displaced by the Serrano Point fault); and 3) Group 1 faults are not disturbed. Therefore, the oldest faults within the SFZ are those belonging to Group 3, trending northwest. Even after aerial photographic interpretation and field reconnaissance, the complex relationship between faults of Group 1 and Group 2 could not be determined.

The Alvord fault

The Alvord fault, a major structure within the Steens Fault Zone, is located primarily within the unconsolidated sediments that flank the lower range-front of the eastern Steens Mountain escarpment. It is expressed as a continuous north-trending, east-facing scarp, for approximately 17 km. This fault was studied in detail in order to determine the age of the most recent faulting. In addition to collecting geologic data, topographic profiles were measured across the scarp along selected segments of the fault for slope degradation analysis.

Geomorphology of the Alvord fault

The Alvord fault can be clearly identified on aerial photographs as a slightly sinuous trace paralleling the eastern range front of
Steens Mountain, the trend varies from N18W to N12E, with an average of NO1W. It is expressed on the surface as an east-facing scarp with a maximum height of 2.5 m.

At its southern end the fault crosses the valley floor playa where it branches into a zone up to 700 m wide of faults that disrupt intermediate lacustrine (Q1d2), and possibly dune deposits (Qsd) (Plate 1; Figure 11; T36S, R34E, sec. 5,8). Along the valley floor, scarp crests are sharp on aerial photographs and show no sign of wave modification. Numerous springs occur at the scarp base along this segment.

Further north, the Alvord fault displaces intermediate lake deposits (Q1d2) and the strandline associated with these deposits at 1280 m (Plate 1; Figure 16; T35S, R34E, sec. 32). Tertiary volcanic rocks have been uplifted and backtilted to the west, forming a prominent spur informally named Black Dog Spur.

Northward for a distance of about 77 km (Plate 1; T35S, R34E, sec. 29, 20, 17, 8 and 5), the fault displaces deposits of the intermediate lacustrine unit (Q1d2) and alluvial fan sediments deposited along the eastern flank of Alvord Point (Qaf2) (Plate 1, Figure 11). Geomorphic evidence of vertical offset of the fan deposits includes: 1) deep channel incision above the fault, resulting in abandonment of the Qaf1 surface; 2) perched stream terraces above the scarp; and 3) formation of alluvial cones (Qaf2) at the new channel mouth formed below the scarp (Figure 17). Springs are numerous and characteristically occur as marshy, thickly
Figure 16
Intersection of Alvord fault and 1280 m shoreline along eastern flank of Alvord Point. Intermediate lake deposits (Qld$_2$) lie below the shoreline.
Figure 17

Alluvial fan (Qaf_2) formed below the Alvord fault on the east side of Alvord Point.
vegetated areas at the base of the scarp.

Continuing toward the north for a distance of about 5 km (Plate 1; T35S, R34E, sec. 5 and T34S, R34E, sec. 32, 29, 28 and 20), the Alvord fault bounds the base of the highest portion of the range, the Central Steens. Here the youngest trace coincides with more numerous and larger alluvial fans. Immediately north of Tuffy Creek, the Alvord Hot Springs flow from the base of the fault scarp (Plate 1; Figure 13; T35S, R34E, sec. 32). Approximately 400 m north of the hot springs the Alvord fault splays into two segments: 1) a barely perceptible scarp trending northeast into the valley and becoming indistinguishable in the grass covered, marshy distal fan/lake unit (Qfu/ld2); and 2) a distinct east-facing scarp trending northwest toward the range front. Along the northwest trending segment, approximately 600 m from the hot springs, the fault displaces the older alluvial fan sediments associated with Indian Creek (Qaf1). A well preserved fault-line graben, possibly related to gravity settlement of the fan sediments during surface rupture, exists along the length of the fault within this unit (Figure 18; Appendix A, profiles 7, 8, 9, 17, 18, and 19). Preservation of this features provides evidence of the recency of the faulting, because generally grabens of this nature are eliminated by sedimentation in less than a few hundreds of years (David Schwartz, 1986, personal communication). The persistence of the graben at this location is probably due to the lower erodability of the moderately to well-indurated coarse sands and gravels of the older alluvial fan deposits.
Figure 18
View to the west of a basal graben preserved along the Alvord fault within the older alluvial fan deposit (Qaf₁). The depth of the graben is about 2 m at this site.
compared to the non-indurated silts, clays and sands of the lacustrine deposits where no graben exists.

Northward along the northwest-trending segment, the fault once again bifurcates and can only be clearly distinguished for another 0.5 km to the north (Plate 1). The two traces then cut landslide debris (Q1s). It is not apparent whether the fault ends at this site or continues higher into the volcanic rocks of the range front.
An analysis of the eroded scarp of the Alvord fault was conducted to provide further evidence of the recency of activity. Wallace (1977) and Bucknam and Anderson (1979) reported that the angle of the degraded slope is dependent on the age and the initial height of the scarp. Furthermore, the profile of the eroded scarp resembles the error function, the solution to the diffusion equation for step-like initial conditions (Hanks et al., 1984; Hanks, 1986, personal communication). Therefore, the profile of the scarp can be measured in the field and analyzed by a mathematical model to provide a good first estimate of the age that faulting occurred.

Fourteen profiles of the Alvord fault were measured following the procedures outlined by Wallace (1977) and Bucknam and Anderson (1979). The fault profiles were surveyed along the central and northern Alvord fault (Plate 1). Selection of the profile sites was non-random in order to avoid modification of the scarp induced by humans, animals or localized erosion. Also, scarp areas that were less vegetated were chosen in order to minimize measurement error. Of the 14 profiles measured, 7 were located in intermediate lake deposits (Q1d2; Profile #1, 2, 3, 10, 11, 13, and 16; Table 2) and 7 were measured in older alluvial fan deposits (Qaf1; Profile #7, 8, 9, 15, 17, 18 and 19; Plate 1). The 7 profiles measured in the older alluvial fan deposits (Qaf1) included a basal graben. The graben does not allow the slope to conform to the profile that is similar to
TABLE 2

VALUES FOR FAULT SCARPS LACKING BASAL GRABENS AS ANALYZED ACCORDING TO HANKS et al., 1984.

<table>
<thead>
<tr>
<th>Profile</th>
<th>2a (m)</th>
<th>(\theta_s)</th>
<th>(\theta_f)</th>
<th>(X_{84})</th>
<th>(\kappa t)</th>
<th>(t)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>1.5</td>
<td>19</td>
<td>2</td>
<td>2.6</td>
<td>1.69</td>
<td>1500 yr</td>
</tr>
<tr>
<td>2.</td>
<td>2.5</td>
<td>26</td>
<td>5</td>
<td>2.4</td>
<td>1.44</td>
<td>1310 yr</td>
</tr>
<tr>
<td>3.</td>
<td>1.4</td>
<td>21</td>
<td>3</td>
<td>1.4</td>
<td>0.49</td>
<td>450 yr</td>
</tr>
<tr>
<td>10.</td>
<td>2.5</td>
<td>21</td>
<td>8</td>
<td>2.85</td>
<td>2.04</td>
<td>1850 yr</td>
</tr>
<tr>
<td>11.</td>
<td>2.2</td>
<td>20</td>
<td>7</td>
<td>2.8</td>
<td>1.95</td>
<td>1780 yr</td>
</tr>
<tr>
<td>13.</td>
<td>1.78</td>
<td>17</td>
<td>7</td>
<td>2.83</td>
<td>1.99</td>
<td>1800 yr</td>
</tr>
<tr>
<td>16.</td>
<td>1.68</td>
<td>18</td>
<td>5</td>
<td>2.58</td>
<td>1.66</td>
<td>1510 yr</td>
</tr>
</tbody>
</table>
the error function (Hanks et al., 1984; Hanks, 1986, personal communication). Therefore, the analysis could not be performed on these profiles.

Assumptions

In order to derive an age estimate for the last episode of faulting, it is assumed that the scarp of the Alvord fault was formed during a single event. Slope profiles and field evidence support this assumption; there are no multiple breaks in slope which would indicate a compound scarp (Wallace, 1977).

A second assumption necessary for the degradation analysis is that a value for the mass diffusivity constant ($\kappa$) can be estimated. This constant incorporates all variables responsible for the modification of the scarp due to climate, aspect and composition, but excluding time. Hanks et al. (1984) and Hanks and Wallace (1985) estimated that the value of $1.1 \text{ m}^2/1000 \text{ yr}$ was a good initial value for this constant in the Great Basin province. This estimation was based on analyses of shoreline scarps in the Lake Lahontan and Lake Bonneville basins. It was assumed that this value for "$\kappa"$ would provide a good first estimate for the mass diffusivity constant in this analysis of the Alvord fault scarp (Hanks, 1986, personal communication).
Methods developed by Hanks et al. (1984) and Hanks and Wallace (1985) utilizing the linear diffusion model for the analysis of the scarp height, slope angle and age relationship were used in this study. The various geometric components of the fault scarp (Figure 19) have been thoroughly described by Wallace (1977) and Hanks et al. (1984). The slope angle $\theta_s$ is the maximum angle of the degraded scarp face. The far-field slope angle $\theta_f$ is the angle of the original unfaulted surface. The scarp offset $2a$ and scarp height $H$ are equal when the far-field slope value $\theta_f$ is zero; when $\theta_f$ is greater than zero, $2a$ is smaller than $2H$ (Figure 19).

A graphical method of analyzing the degraded fault scarp is discussed in Hanks et al. (1984), and was demonstrated to me by Dr. Hanks (1986, personal communication). It has been observed (Hanks et al., 1984) that most scarps possess a nearly anti-symmetrical profile (Figure 20). The error function predicts the anti-symmetrical profile of the fault scarp by the equation:

$$\xi = \frac{x}{2(\kappa t)^{1/4}}$$

where "$\xi$" is the error function, "$x$" is the location on the abscissa that corresponds to the error function value, "$\kappa$" is the mass
Fault scarp geometry, from Hanks, et al. (1984). 2a is the scarp offset, 2H is the scarp height. $\theta_s$ is the scarp slope angle and $\theta_f$ is the far-field slope. The slope ($\theta_s$) depicted in this sketch is of a fault scarp that has been degraded for a period of time (t).
Figure 20

Graphic method of estimating scarp age \( t \) by analysis of scarp morphology, from Hanks et al. (1984). Using the ordinate value of 0.84 corresponding to the error function value of 1, the abscissa value \( x_{84} \) is found. Using a \( K \) value of 1.1 \( m^2/1000 \) yr and the graphically derived \( x_{84} \) value, \( t \) can be found.

\[
\frac{x_{84}}{2(Kt)^{1/2}} = 1
\]
Figure 22
diffusivity constant and "t" is time. When the argument of the error function is equal to 1, the 84% amplitude point is attained. Therefore, the value for "t" can be obtained when,

\[ 1 = \frac{X_{84}}{2 (\kappa t)^{\frac{3}{2}}} \]

and the diffusivity constant (\(\kappa\)) = 1.1 m²/1000 yr is used.

The scarp profiles that did not include a graben were drafted with the point of zero anti-symmetry located at "X" and "Y" equal to zero (Appendix A). Points of convexity along the profile are assigned positive "X" and "Y" values, while the concave portions of the slope receive negative values. The \(X_{84}\) value, or the vertical value equal to 84% of the height of the convex slope, is graphically estimated. A horizontal line is drawn through the slope, and the "X" value (horizontal distance from zero along the scarp) is calculated. This \(X_{84}\) value is then entered into the error function equation (Figure 20). The solution to this equation is the product:

\[ \kappa t = X_{84} \]

for which "t" can be found by supplying the given "\(\kappa\)" value of 1.1
m²/1000 yr (Hanks et al., 1984).

Results

The "t" values for the scarp profiles were all less than 2000 yr (Table 2). Profile #3 resulted in an anomalously low "t" value of 450 yr. A re-evaluation of the site found that the scarp is adjacent to a recent channel in the alluvial fan deposits (Qaf₂) associated with Indian Creek. Therefore, the young value is probably the result of fluvial modification by channelized water traversing the base of the slope. The values for the scarp offset 2a and slope angle (θₘ) are plotted in Figure 21 along with values from other areas (Hanks et al., 1984). The steep slope of the best fit line for the Alvord fault profiles is an indication of the youthful character of the scarp (Hanks, 1986, personal communication).

The results of this slope analysis provide evidence that the Alvord fault scarp is young, and that the fault which produced the scarp was subjected to movement probably in late Holocene time.

Limitations

Evaluation of the degraded scarp of the Alvord fault must be viewed with the limitations of the study in mind. First, the value of 1.1 m²/1000 yr is only a general first approximation for the mass diffusivity constant "κ". It does not take into account differences in erosivity variables associated with this study area, such as
Figure 21

Plot of relationship between slope offset (2a) and scarp slope (θs) for four fault scarps in Utah from Hanks et al. (1984). Best fit lines are plotted for each scarp. In addition, data from Alvord fault is plotted as "t" with best fit line. Line slope is steeper than Utah data as expected for younger scarp. Arrow points to anomalous value for profile No. 3.

○ Fish Springs Range, Utah, Kt = 3.3 m², t = 3000 yr B.P.
○ Oquirrh Mountains, Utah, Kt = 35 m², t = 32,000 yr B.P.
□ Sheeprock Mountains, Kt = 58 m², t = 53,000 yr B.P.
△ Drum Mountains, Utah, Kt = 5.8 m², t = ~8700 yr B.P.
aspect of the fault scarp, precipitation associated with latitude of
the area, the type of material comprising the scarp, the original
morphology of the scarp face and numerous other factors. Differences
in opinion about the role and magnitude of the diffusivity constant
are numerous (Nash, 1980; Colman and Watson, 1983; Mayer, 1984).

The second limitation involves measurement error. The method of
using an Abney level and stadia rod to measure the fault scarp is not
precise. The angle of the rod is affected by surface features and
vegetation along the slope profile. These sources of measurement
error cannot be avoided without modifying the scarp. In addition,
any unforeseen modifications by animals or fluvial processes would
affect the results of the analysis (see Profile #3, Appendix A, Table
2).

Finally, if a fault scarp is the product of multiple movements
over a long period of time, and the assumption that it is a single
event scarp is made, then the interpretation of the results will be
incorrect.
Discussion

The Steens Fault Zone is located within the zone of transition between the Basin and Range province to the east and south, and the relatively unfaulted Columbia River Plateau to the north and west (Figure 22). Therefore, fault orientations, slip directions and activity may be related to features within both provinces. Stewart et al. (1975) defined the 750 km-long Oregon-Nevada lineament, a belt of closely-spaced en echelon faults extending from central Oregon to central Nevada, and traversing Steens Mountain. Lawrence (1976) renamed the lineament the Brothers fault zone, and defined 3 other major northwest lineaments in Oregon (Figure 22). He concluded that these structures are right-angle strike-slip faults that were formed to compensate for diminished extension in the northwestern Basin and Range province. In addition, Lawrence reported that the Brothers fault zone was truncated by the Steens Fault Zone.

The northwest-trending faults located within the Steens Fault Zone show no apparent lateral component of movement. However, they do not extend eastward beyond the Alvord fault, which is in agreement with Lawrence (1976).

Donath (1962) identified a prominent fault trend of N20E to N30E in the Summer Lake area of Oregon to the west. These faults are associated with a N30W to N40W-trending fault group which he suggests represents conjugate faults formed during north-south compression.
Figure 22
Although northeast-trending faults are found within the SFZ, it is unclear if they are related to compression.

Wallace (1984) and Wallace and Whitney (1984) have defined a belt of historic and late Quaternary faults within central Nevada referred to as the Central Nevada-Eastern California Seismic belt (Figure 22). Four large faulting events associate with earthquakes of $M > 6.8$ occurred within this belt. In addition, Wallace (1984) reported that no large scarps younger than 300 yr B.P. have been found within this belt. Instead, only fault scarps with ages on the order of thousands of years are found. Wallace (1984) suggested that the historic faulting may be a pulse of activity that had taken place after a period of quiescence. It is also noteworthy that surface rupture did not take place along a continuous fault, but instead occurred in discrete segments. The ruptured segments were separated by non-ruptured areas up to 40 km long.

Wallace (1984) described two tectonic structures within Nevada, the Black Rock-Carson Sink zone of extension, and the Central Nevada downwarp (Figure 22). Both structures are zones of extension with the Central Nevada Seismic Belt defining the boundary between them. Wallace (1984) suggested that the higher degree of Holocene activity within the Black Rock-Carson Sink zone may indicate a migration of activity toward the northwest. The northeastern termination of the Black Rock-Carson Sink zone of extension and the Central Nevada Seismic Belt is defined by the Oregon-Nevada lineament, and their
southwestern margins are represented by the Walker Lane. Wallace (1984) also reported that the northwestern boundary of the Black Rock-Carson Sink zone of extension is defined by a finger-like zone of denser faulting. The projection of this zone northward along strike roughly coincides with the SFZ (Figure 22). Evidence of late Holocene faulting within the Steens Fault Zone in conjunction with Wallace's (1984) assertion that tectonic activity may be migrating toward the northwest may indicate a zone of Holocene faulting parallel and northwest of the Central Nevada Seismic Belt that has been historically quiescent.
SUMMARY AND CONCLUSIONS

In summary, late Holocene faulting occurred along the base of the eastern escarpment of Steens Mountain and the western margin of the Alvord Desert in southeast Oregon. Evidence is provided by faulted pluvial and alluvial deposits as well as fault scarp geomorphology.

Stratigraphic relationships of lacustrine features, including lake beds, gravel bars and shoreline terraces, suggest two pluviations prior to the present occupation of Alvord Valley, with the older high stand at an elevation of 1310 m and the younger at 1280 m. The lacustrine deposits are in contact with alluvial fan units that were deposited either during periods of low lake level or complete dessication within the valley. Big Sand Gap, located at the eastern margin of the valley, served as a spillover outlet during the first pluvial lake stand and lowered the lake level by 30 m to an elevation of 1280 m.

Fault scarps associated with a complex zone of tectonic rupture, named the Steens Fault Zone (SFZ) in this report, are located along the southeastern range front of the Central Steens. Fault orientations within the SFZ typically fall within three groups:
Group 1 - north to northeast-trending faults with east-facing scarps;
Group 2 - northwest-trending faults with northeast-facing scarps; and
Group 3 - northeast-trending faults with northwest-facing scarps.

Group 1 faults are related to present Basin and Range
deformation, in response to generally east-west extension. Group 2 faults have similar orientations to northwest-trending zones of right-lateral strike slip in central Oregon and central Nevada that are believed to have resulted from diminished extension in this portion of the Basin and Range province (Lawrence, 1965). Group 1 and 2 faults displace Group 3 faults.

A north-trending segment of the SFZ, named the Alvord fault in this study, is expressed at the surface as a scarp which displaces Quaternary sediments for a length of 17 km along the Central Steens range front. Evidence of recent faulting includes: 1) a sharp scarp crest; 2) a well-developed basal graben which displaces coarse, moderately consolidated alluvial fan deposits of unit Qaf₁; and 3) displacement of a 1280 m elevation shoreline of pluvial Lake Alvord correlated to the 16,000 to 12,000 yr B.P. high stand of Lake Lahontan.

In order to provide an estimate of the timing of the most recent faulting event, topographic profiles were surveyed perpendicular to the fault for analysis of the eroded slope. This analysis was based on the assumptions that: 1) the scarp was formed during a single rupture event; and 2) \( k = 1.1 \text{ m}^2/1000 \text{ yr} \) provides a good first estimate of the mass diffusivity constant for the study area.

Resultant "t" values were all less than 2000 yr. Therefore, the Alvord fault has been subjected to tectonic movement in late Holocene time as evidenced by: 1) displacement by the fault of the 16,000 to 12,000 yr B.P. pluvial Lake Alvord shoreline; 2) preservation of a
basal graben along a portion of the fault; and 3) a youthful scarp which when analyzed provides "t" values of less than 2000 years.

Based on evidence provided by the relative positions of pluvial and interpluvial deposits and faults within the SFZ, the following sequence of late Quaternary depositional and tectonic activity is proposed:

1) occupation of the Alvord Valley by pluvial Lake Alvord to an upper elevation of 1310 m at approximately 24,000 to 16,000 yr B.P.;

2) lake spillover at Big Sand Gap during the pluviation and subsequent incision of the gap to an elevation of 1280 m, followed by stabilization of the lake level at this elevation;

3) progradation of alluvial fans (Qaf₁) in response to a lowered lake level;

4) a period of maximum pluviation occurring from 16,000 to 12,000 yr B.P. throughout the Great Basin province, but the level of Lake Alvord was restricted by Big Sand Gap. Climatically-induced lake level reduction began at about 12,500 yr B.P.;

5) progradation of alluvial fans (Qaf₂) toward a base level of 1219 m (present playa surface) in the central portion of the valley. Fans at Pike Creek, Little Alvord Creek, and Big Alvord Creek overlap gravel bar units Qbb₂;
6) Late Holocene surface rupture event along at least 17 km of the Alvord fault.

The faults within the SFZ may be related to a zone of Holocene seismicity which trends roughly north-northeast to south-southwest along the northwest margin of the Black Rock-Carson Sink zone of extension.
REFERENCES

Benson, L. V., 1978, Fluctuation in the level of pluvial Lake Lahontan during the last 40,000 years: Quaternary Research, v. 9, p. 300-318.


APPENDIX A

Scarp profiles of the Alvord fault. Profiles without a basal graben were used in degradation analysis; those with grabens were excluded from the analysis. All scarps were formed during a single surface rupture event.
Profile No. 1

2a = 1.5m
θs = 19°
θf = 2°
xₘ₈₄ = 2.6m
Kₜ = 1.69
ₜ = 1500 yr

Profile No. 2

2a = 2.5m
θs = 26°
θf = 5°
xₘ₈₄ = 2.4m
Kₜ = 1.44
ₜ = 1310 yr
Profile No. 3

- $x_{84} = 1.4m$
- $\theta_s = 21^\circ$
- $\theta_f = 3^\circ$
- $2a = 1.4m$
- $kt = 0.49$
- $t = 450 \text{ yr}$

Profile No. 7

- Graben
Profile No. 10

$2a = 2.5\text{ m}$
$
\theta_s = 21^\circ$
$
\theta_f = 8^\circ$

$x_{84} = 2.85$

$a$

$t = 1850 \text{ yr}$

Profile No. 11

$2a = 2.2\text{ m}$
$
\theta_s = 20^\circ$
$
\theta_f = 7^\circ$

$x_{84} = 2.8$

$a$

$t = 1780 \text{ yr}$
Profile No. 13

$2a = 1.78\text{m}$
$\theta_s = 17^\circ$
$\theta_f = 7^\circ$
$x_{84} = 2.83\text{m}$
$K_t = 1.99$
$t = 1800\text{ yr}$
Profile No. 16

Profile No. 17

$2a = 1.7m$

$\theta_s = 18^\circ$

$\theta_f = 5^\circ$

$x_{94} = 2.58$

$K_t = 1.66$

$t = 1510 \text{ yr}$