Sedimentological constraints on Middle Miocene extensional tectonism of the southern Las Vegas Range, southern Nevada

Jack Edward Deibert
University of Nevada, Las Vegas

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Sedimentological constraints on Middle Miocene extensional tectonism of the southern Las Vegas Range, southern Nevada

Deibert, Jack Edward, M.S.
University of Nevada, Las Vegas, 1989
SEDIMENTOLOGICAL CONSTRAINTS ON MIDDLE MIocene
EXTENSIONAL TECTONISM OF THE SOUTHERN
LAS VEGAS RANGE, SOUTHERN NEVADA

by

Jack Edward Deibert

A thesis submitted in partial fulfillment
of the requirements for the degree of

Master of Science

in

Geology

Department of Geoscience
University of Nevada, Las Vegas
August, 1989
The thesis of Jack Edward Deibert for the degree of Master of Science in Geology is approved.

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August, 1989
ABSTRACT

An integrated depositional-tectonic model is proposed for the reconstruction of a Miocene lacustrine basin in the southern Las Vegas Range, Nevada, that was fragmented during a period of extensional tectonism. The model suggests that the sediments were deposited in an isolated basin during initiation of normal faulting in the Sheep Range extensional allochthon 12 to 16 Ma. After 12 Ma, deposition in this basin was terminated by high-angle block faulting followed by oroflexural rotation along the Las Vegas Valley shear zone.

The Middle Miocene sequence is exposed as isolated outcrops within grabens between the Sheep Range, Fossil Ridge, and the Las Vegas Range. Collectively, the recognition and interpretation of sedimentary facies, analysis of the provenance of terrigenous material, correlation of volcanic ash units, and mapping of structural features indicate that the isolated exposures represent only fragments of a once laterally continuous sequence. This sequence is informally defined as the Gass Peak formation and does not correlate lithostratigraphically with the Horse Spring Formation nor with any nearby Tertiary deposits. The lower, Fossil Ridge member is 75 m thick and consists of bioclastic limestone with minor interbeds of conglomerate. The upper, 260-m-thick Castle Rock member consists of limestone, dolomite, siltstone, sandstone, conglomerate, and
megobreccia with minor tuffaceous and gypsiferous beds. Both of these members represent deposition in a shallow, tectonically active, permanent lake into which episodic, high-intensity stream discharge deposited coarse clastic sediment 12 to 16 Ma. Post 12 Ma, north-striking, high-angle normal faults displaced the Gass Peak formation on both sides of what is presently Fossil Ridge. Subsequently, oroflexural bending, which was the result of drag along the Las Vegas Valley shear zone, reactivated the north-striking faults as strike-slip faults and rotated them into their present northeast-striking orientation.

The age and unique depositional basin of the Gass Peak formation support published models that the Sheep Range extensional allochthon was an extensional terrane separate from the Lake Mead area. Additionally, the proposed model suggests that oroflexural deformation along the Las Vegas Valley shear zone is younger than 12 Ma and is younger than normal faulting.
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INTRODUCTION

Tertiary sedimentary units exposed at the southern terminus of the Las Vegas Range, Nevada (Fig. 1) provide essential data in resolving the timing and style of structural deformation in the Sheep Range extensional allochthon and associated Las Vegas Valley shear zone. These Tertiary rocks were deformed by normal faults within the Sheep Range extensional allochthon and rotated about a vertical axis by movement along the Las Vegas Valley shear zone (Ebanks, 1965; Guth, 1980, 1981; Wernicke and others, 1982, 1984; Nelson and Jones, 1987). Prior to this study, the timing and relationship between these two structural features was poorly understood. The purpose of this investigation was to define the stratigraphy, age, and correlation of the Tertiary sediments and to determine their depositional and structural relationship to the deformation in the Sheep Range extensional allochthon and movements along the Las Vegas Valley shear zone.

An integrated depositional-tectonic model is proposed for the reconstruction of a Middle Miocene lacustrine basin in the southern Las Vegas Range, Nevada (Fig. 1), that was fragmented during a period of extensional tectonism. This model requires that multiple basins formed in southern
Figure 1. Regional location map of the study area. Heavy lines with ball symbols represent major normal faults with the ball on the hangingwall. Heavy lines with triangles represent major thrust faults with triangles on the hangingwall. The area of the Sheep Range extensional allochthon is marked by diagonal lines.
Nevada during the early stages of extension; each basin had an unique depositional history; and the resulting sedimentary sequences were dismembered by continued extensional movements. Furthermore, the model suggests that normal faulting in the Sheep Range extensional allochthon predates some oroflexural deformation adjacent to the Las Vegas Valley shear zone. The model is based on considerations of regional geology, local stratigraphy and structure; radiometric dating and correlation of ash units; and analysis of lithofacies and their depositional environments.

The investigation was accomplished by mapping, measuring stratigraphic sequences, and sampling rocks throughout the study area. Petrographic studies were conducted on pertinent samples for lithology, fossil content, and mineral identification. Chemical analyses of selected volcanic ash units were performed to assist in stratigraphic correlations. Orientations of sedimentary structures were recorded for use in basin analysis and interpretation of depositional processes. The study area (Fig. 1) lies within the U.S.G.S. Gass Peak 7.5' quadrangle and the southeast portion of the Corn Creek Springs 7.5' quadrangle. Unpublished maps by Maldonado and Schmidt (submitted) and Ebanks (1965) served as base maps for locating and studying the Tertiary outcrops.
REGIONAL GEOLOGY

In southern Nevada, Upper Proterozoic and Paleozoic rocks were thrust eastward during the Mesozoic Sevier orogeny and subsequently were deformed by normal and strike-slip faults during Tertiary extensional tectonism. These deformed Upper Proterozoic and Paleozoic rocks are the source for the Tertiary conglomerates throughout southern Nevada: thus, understanding their structure and distribution is critical to resolving provenance, transport direction, and Tertiary deformation. Consequently, published and unpublished maps, and summaries of the geology in surrounding ranges (Longwell and others, 1965; Guth, 1980, 1981, 1986) and the study area (Ebanks, 1965) were integrated into this more detailed investigation.

Upper Proterozoic and Paleozoic rocks of the region consist of a thick passive margin sequence of sandstone, carbonate rocks and minor siltstone and shale (Fig. 2; Longwell and others, 1965; Guth 1980, 1986). In the eastern part of the study area, a complete stratigraphic sequence of Upper Proterozoic Stirling Quartzite through Cambrian Bonanza King Formation is exposed in a homocline that is disrupted only by minor small-scale folds and faults (Fig. 3). The most prominent of these minor structures is a small
Figure 2. Generalized stratigraphic column for the Sheep and Las Vegas ranges.
Figure 3. Generalized geologic map of the study area. Faults are shown as heavy lines. Ball symbols on faults are placed on the hangingwall of normal faults. Triangle symbols are placed on the hangingwall of thrust faults. Thin lines represent contacts between stratigraphic units. Stratigraphic symbols are located on Figure 2 except for the following: O-D = Ordovician through Devonian rocks undivided; Tgpf = Tertiary Fossil Ridge member of the Gass Peak formation (shown in stippled pattern); Tgpc = Tertiary Castle Rock member of the Gass Peak formation (shown in diagonal line pattern); Qal = Quaternary alluvium. Arrows with "M" represent locations of measured sections. Arrow with "B" represents location of interbedded breccia. Arrows with "C" are locations of interpolated contacts between the members of the Gass Peak formation. Arrows with double barbs are small tectonic fold axes. Orthogonal arrows locate a monocline with the larger arrow marking the trend and the smaller arrow marking the limb. Horizontal bedding is shown as a circled cross. Strike and dip of strata is shown as elongated "T" symbols with the number corresponding to the dip angle. Overturned beds are shown with hooked strike and dip symbol. Geology by the author after Ebanks (1965) and Maldonado and Schmidt (submitted). Location of cross-section for Figure 16 is marked as A and A'.
monocline that occurs at the east end of Fossil Ridge (Fig. 3). The axis of the monocline trends north with the limb dipping west. Erosion of this structure has exposed the contact between the Cambrian Nopah Formation and Bonanza King Formation, which trends almost north-south where Fossil Ridge joins the Las Vegas Range. This contact is perpendicular to the regional strike, and its peculiar orientation probably was the reason previous workers (Ebanks, 1965; Longwell and others, 1965) mapped it as a fault. The contact produces an apparent offset if one considers only the regional strike and dip. However, considering the monocline structure and the conformable stratigraphic sequence exposed in a gully along the contact, this contact clearly is not a fault. The Nopah Formation crops out only along Fossil Ridge with part of the overlying Ordovician Pogonip Group exposed along the western terminus of the ridge (Fig. 3). The Mississippian through Permian Bird Spring Formation is exposed along the southern portions of the study area, occurring directly south of, and in fault contact with, the Stirling Quartzite.

During the Mesozoic Sevier orogeny, the Stirling Quartzite and overlying formations were thrust eastward over the Bird Spring Formation along the Gass Peak thrust (Fig. 1 and Fig. 3). The stratigraphic displacement along this thrust is 5900 m, and horizontal displacement is possibly up to 60 km (Guth, 1980). The trace of the thrust trends north
along the Las Vegas Range except near its southernmost exposure, which lies within the study area (Fig. 1). There the thrust trace apparently has been rotated 90° clockwise about a vertical axis resulting in an eastward trend. This rotation has been attributed to oroflexural bending along the Las Vegas Valley shear zone (Burchfiel, 1965; Ebanks, 1965). Paleomagnetic studies of the Cambrian formations in the upper plate of the Gass Peak thrust have confirmed that this easterly trend resulted from rotation about a vertical axis and not from apparent rotation due to ramp geometry or erosional level (Nelson and Jones, 1987). The plane of the thrust fault within the study area (Fig. 3) dips 40° to the north. Strata of the lower plate rocks near the thrust are overturned and dip 40° to 65° northward (Ebanks, 1965; Maldonado and Schmidt, submitted, unpublished mapping). Rocks of the upper plate dip 55° northward at the thrust contact and dip 30° northward some 5 km north of the thrust trace (Fig. 3).

After Sevier thrusting and before Tertiary extension, a period of regional erosion denuded the upper plate of the Gass Peak thrust exposing the Mississippian Indian Springs Formation as the youngest stratigraphic unit (Guth, 1981). No Mesozoic rocks are exposed in the Las Vegas and Sheep ranges (Longwell and others, 1965; Guth, 1980, 1986).

Normal and strike-slip faulting deformed rocks within the region during Tertiary extension. The major faults
responsible for deformation of the upper plate of the Gass Peak thrust were the Mormon Pass fault and the Las Vegas Valley shear zone (Fig. 1). The Mormon Pass fault has been interpreted as the major west-dipping detachment fault of the Sheep Range extensional allochthon and may sole out along the Gass Peak thrust (Guth, 1980, 1981; Wernicke and others, 1984). It is recognized by the juxtaposition of Ordovician and Middle Cambrian rocks with 1600 m of section missing and approximately 20° of eastward rotation of strata in the Sheep Range (Wernicke and others, 1984). The fault trends north along the west flank of the central Las Vegas Range, but, at its southern most exposure, it trends northeast and is parallel to the northern flank of Fossil Ridge.

The Las Vegas Valley shear zone is a proposed northwest trending, right lateral strike-slip fault whose trace is buried beneath the alluvium in the Las Vegas Valley (Longwell, 1960). The presence of the fault is implied by localized clock-wise bend in the trend of ranges adjacent to the valley and up to 65 km offset of correlative thrusts and Paleozoic facies across the valley (Ross and Longwell, 1964; Burchfiel, 1965; Stewart and others, 1968). The shear zone was interpreted as a transform boundary between the Sheep Range extensional allochthon occurring north of the shear zone and the unextended Spring Mountains region to the south (Guth, 1981; Wernicke and others, 1982, 1984). Previously,
the timing of movement along the shear zone was restricted to between 15 Ma and 10.7 Ma based on the apparent rotation of 15 Ma strata and undeformed 10.7 Ma volcanics crossing the fault (Fleck, 1970). The 10.7 Ma younger bound on the age of the shear zone is equivocal because the shear zone may not extend into the area where these volcanic rocks are exposed (Bohannon, 1979). As discussed subsequently in the "Tertiary Structure" section, the apparent rotation of radiometrically dated volcanic sediments in the study area indicate that the rotation is no older than 12 Ma.

Synchronous with extensional deformation during the Tertiary, a wide variety of non-marine sediments were deposited throughout southern Nevada. They are exemplified by the widespread Miocene Horse Spring Formation of the Lake Mead region (Bohannon, 1984) but also include the temporally correlative rocks of this study (Guth, 1980, 1981; Guth and others, 1988).

Quaternary alluvial-fan sediments unconformably overlie the Tertiary deposits. Modern drainages dissect the alluvial-fan sediments, exposing the Quaternary and Tertiary strata in the Las Vegas Range.

Problems of vertical rotation

Problems arise in the analysis of the Tertiary basin and associated fluvial transport directions because rocks in the study area may have undergone a clock-wise rotation.
about a vertical axis as a result of movement along the Las Vegas Valley shear zone. The timing of this movement is poorly constrained. Therefore, all directions given in the "Lithofacies and Depositional Processes," and in the "Depositional Model" sections of this paper are corrected for rotations only about a horizontal axis. This correction produces maps that may be re-oriented by the reader to test any relevant structural hypotheses. Rotation about a vertical axis are addressed in the "Summary of Tertiary Geologic History" and "Discussion of Regional Implications" sections of this paper.

TERTIARY STRATIGRAPHY

Gass Peak formation

A unique Middle Miocene lithostratigraphic unit in the southern Las Vegas Range is described and informally defined here as the Gass Peak formation (Fig. 4). The recognition of this local stratigraphy allows the scattered Miocene outcrops in the study area to be reconstructed into a pre-deformation configuration for local basin analysis and for comparison with temporally equivalent units throughout southern Nevada.

Figure 4. Generalized stratigraphic column for the Gass Peak formation. The letter "A" is the stratigraphic occurrence of tuff sample T7. The letter "B" is the stratigraphic occurrence of the correlative ash unit, tuff sample T13.
Conglomerate: clast-supported, poorly sorted, sub-rounded clasts of quartz arenite. Interbedded with white limestone, dolomite and volcanic ash.

Megabreccia: monolithic, clast-supported breccia with clasts of quartz arenite and carbonate 1 mm to 3 m in diameter.

Interbedded limestone, sandstone, and siltstone: white to tan, thin to medium bedded, gypsiferous. Commonly ripple cross-laminated and soft-sediment folded.

Conglomerate: clast-supported, poorly sorted, sub-rounded clasts of carbonate and quartz arenite. Interbedded with tan to white limestone.

Bioticlastic limestone: buff to cream colored, thin to medium bedded, commonly containing bioclasts of ostracods, gastropods and plant fragments. Minor volcanic ash.
The Gass Peak formation is a faulted sequence with a minimum thickness of 335 m. Its lower boundary is faulted, and its upper boundary is an unconformity. It consists of interbedded limestone, siltstone, sandstone, conglomerate, megabreccia, and minor amounts of volcanic ash. The name of the formation is derived from Gass Peak, a prominent mountain located just southeast of the exposures of the formation (Fig. 3). The Gass Peak formation crops out over a 20 km² region between the north slope of Fossil Ridge and the southern slopes of Gass Peak and Castle Rock (Fig. 3) where exposures occur in deep drainages below the overlying Quaternary alluvium.

The rocks of the Gass Peak formation were originally assigned to the Horse Spring Formation (Ebanks, 1965; Longwell and others, 1965). One of the intents of this paper is to suggest discontinuing the use of this lithostratigraphic assignment and to informally define a new lithostratigraphic unit.

The general stratigraphy of the Gass Peak formation has been assembled from numerous isolated, fault-bounded outcrops (Fig. 3) using stratigraphic correlations based on similarities in lithology, fossil content, geochemically fingerprinted ash units and simple structural models to interpret hidden relationships. Parts of the vertical stratigraphy (Fig. 4) are speculative but reasonable, given the limited outcrops and the resolution of biostratigraphic
and geochronologic data. Thus, the stratotype for the Gass Peak formation is designated as a composite-stratotype.

Neither the true thickness of the formation nor its basal contact can be determined with certainty. All of the contacts between the Gass Peak formation and the underlying Upper Proterozoic and Paleozoic rocks are faults. In addition, the contact between the Fossil Ridge member and the Castle Rock member is not exposed, but they seemingly are conformable units (Fig. 4).

Fossil Ridge member

The Fossil Ridge member of the Gass Peak formation (Tgpf), which is exposed along the south flank of Fossil Ridge (Fig. 3), is composed primarily of limestone with minor units of conglomerate, volcanic ash, sandstone and siltstone. The stratigraphic section for the Fossil Ridge member was measured at the largest outcrop of the Gass Peak formation south of Fossil Ridge (Fig. 3). This 75-m-thick section is defined as the type section for the Fossil Ridge member. The stratigraphy is not clearly defined due to folding and covered slopes and the thickness is estimated. The contacts between the Fossil Ridge member and the underlying Cambrian Nopah and Bonanza King formations are interpreted as faults contrary to Ebanks (1965, p. 16) who briefly described a basal conglomerate in his description of these rocks. Upon detailed examination of this local
stratigraphy, I conclude that at least 57 m of Gass Peak formation lie stratigraphically below this conglomerate. The basal conglomerate described by Ebanks (1965), as well as other beds, are juxtaposed against the Cambrian Bonanza King Formation, but this juxtaposition appears to be a fault. This interpretation is based on the brecciated and recrystallized nature of rocks along the contact and the variable angular discordance between the Gass Peak formation and the Bonanza King Formation. Based on this evidence I believe that no basal conglomerate is exposed, and all the contacts with the underlying Paleozoic rocks are faults. The base of the Fossil Ridge member is placed at the stratigraphically lowest limestone bed exposed, approximately 9 m below a distinctive volcanic ash bed (A in Fig. 4) which is described subsequently.

The 75-m-thick limestone sequence is buff to cream colored, fine-grained, and thin- to medium-bedded with rare light brown chert nodules (Fig. 5a). The limestone is generally a bioclastic lime mudstone to wackestone. The bioclasts are dominantly ostracods with minor gastropods and plant fragments (Fig. 5b). The ostracod species (Appendix A) have been identified by Swain (1965, personal comm. to Ebanks).

Interbedded pebble to boulder conglomerates occur approximately 27 m above the base of the section (Fig. 4 and 5c) and occur throughout the remainder of the section, but
Figure 5. Features of the Gass Peak formation. (A) Thin- to medium-bedded bioclastic limestone with a interbedded ash unit. Hammer lies on the ash unit and is 34 cm long. (B) Photomicrograph of bioclastic limestone exhibiting ostracod bioclasts. Scale bar 0.5 mm. (C) Clast-supported conglomerate interbedded within bioclastic limestone. Hammer is 34 cm in length. (D) Thin-bedded limestone, sandstone, and siltstone. Hammer is 34 cm in length. (E) Limestone with gypsum voids. The coin is approximately 2 cm in diameter. (F) Normally graded conglomerate and volcanic ash. Hammer head is 18 cm in length.
they are most abundant in the upper 10 m. The conglomerates are clast-supported and poorly sorted with bedding ranging from 0.2 to 1.2 m in thickness. Clasts are mostly sub-rounded, dark-colored carbonates and less commonly fine-grained white quartz arenites.

A volcanic ash unit occurs 9 m above the base of the limestone section (Fig. 4 and 5a). It is 0.25 m thick and is a biotite-bearing vitric rhyolite ash tuff (see Table 1 for geochemical analysis, sample T7).

The contact between the Fossil Ridge member and the Castle Rock member is nowhere exposed, but its position is implied from the extrapolation of consistent bedding attitudes at two localities (Fig. 3).

**Castle Rock member**

The Castle Rock member of the Gass Peak formation (Tgpc), which is exposed along the northern flanks of Gass Peak and Fossil Ridge (Fig. 3), consists of 260 m of siltstone, sandstone, limestone, volcanic ash, conglomerate, and megabreccia. This member was measured at the large outcrop of Gass Peak formation along the northern flank of Castle Rock (Fig. 3) and interfingering relationships within the member were observed at other outcrops. The measured section locality serves as the type section for the Castle Rock member. The basal and upper contacts are defined at the lowest beds exposed at the locality and the highest
Table 1. Chemical analysis of selected tuff samples.
Whole rock major element chemistry was determined by Inductively Coupled Plasma techniques at Chemex Labs, Inc. Sparks, Nevada. Average error for each major element value is approximately two percent. Trace elements were analyzed by Instrumental Neutron Activation Analysis at the Phoenix Memorial Laboratory, University of Michigan. Average percentage errors for each trace element are given in the table. Entries labeled ND indicate that element was not detected.
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stratigraphic bed outcropping beneath the Quaternary alluvium along the measured section.

The lower 185 m of the Castle Rock member consists of thin- to medium-bedded, tan to white, unfossiliferous siltstone, sandstone and lime mudstone (Fig. 4 and 5d). All of these rock types are parallel to ripple laminated. Near the top of this 185-m-thick sequence, the rocks are gypsiferous. The gypsum is coarse-grained selenite with crystals 1 to 200 mm in length (Fig. 5e). Commonly the crystals have been removed by dissolution, leaving just the selenite crystal-shaped void.

The upper 75 m of the Castle Rock member consists of a maroon to white interbedded sequence of pebble to boulder conglomerate, limestone, dolomite, and volcanic ash (Fig. 5f). The conglomerate beds range from 0.1 to 2 m in thickness and occur from 185 to 240 m in the section. Typically the conglomerates are clast-supported and internally the beds are massive to normally graded (Fig. 5f). Clasts are sub-rounded to sub-angular, brown to maroon quartz arenites. Within the conglomerate, rare amounts of silicified fossil wood fragments are present. The brown to tan wood fragments, which are composed of microcrystalline quartz and massive opal, range from 100 to 200 mm in diameter and 200 to 1000 mm in length (Fig. 6a). The wood tentatively has been identified as *Palmoxylon* (palm wood) (D.I. Axelrod, 1987, personal comm.).
Figure 6. Features of the Gass Peak formation. (A) Silica replaced palm wood (*Palmoxlon*). Scale bar 3 cm. (B) Palm fossil *Sabalites* preserved in volcanic ash. Scale bar 3 cm. (C) Interbedded megabreccia in the Castle Rock member. Hammer is 34 cm in length. (D) Close-up of megabreccia. Scale bar 4 cm. (E) Megabreccia displaying ghost stratigraphy. Scale bar 10 cm. (F) Matrix supported megabreccia clasts in a limestone unit of the Castle Rock member.
The interbedded sequences of dolomite, limestone, and volcanic ash between the conglomerate beds, range from 0.1 to 1 m in thickness. The lime mudstone and dolomite are white, massive, and unfossiliferous.

The volcanic ash units, where they are interbedded with conglomerate, contain abundant terrigenous clastic and plant material, suggesting that the ash was reworked or redeposited. The plant material consist of broken and whole palm leaves (Fig. 6b) and is identified as *Sabalites* using the fossil palm classification of Read and Hickey (1972). Where the volcanic ash is not interbedded with conglomerates, the ash contains very little terrigenous sediment, and thus it suggests little reworking.

Five of the ash units were analyzed for bulk mineralogy, heavy mineral suites, major and trace elements, and radiometric dating for use as stratigraphic marker horizons (Tables 1 and 2, Fig. 7). These data indicate a strong similarity between samples T13 and T14. They contain a unique accessory mineral assemblage of sphene, zircon, allanite, magnetite, hornblende, and often apatite. In addition, one or two unique light gray ash units composed entirely of glass shards occur approximately 4 to 10 m above each of the ash units from which T13 and T14 were collected. Thus, I argue that these two almost identical ash horizons (T13 and T14) are not only correlative but were once laterally continuous with one another across the area.
Table 2. Summary of tuff components of selected samples.

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Figure 7. Location map for volcanic ash samples and correlative ash unit. The correlative ash unit is marked by a thin line with an arrow indicating the dip of the bed. Locations of the tuff samples are marked by heavy arrows.
Consequently, these horizons will be referred to as the correlative ash unit (B in Fig. 4). Furthermore, the rare earth elemental analysis (Fig. 8) also allow this ash unit to be differentiated from or correlated with other small isolated outcrops including a small exposure of the Gass Peak formation on the north slope of Fossil Ridge (Fig. 7). Figure 7 shows the outcrops and structural attitude of the correlative ash unit in the Gass Peak formation.

Megabreccias are juxtaposed to the Castle Rock member throughout the study area. The nature of the basal contacts of the megabreccia with the Castle Rock member is usually not exposed. Most of the rocks of the Gass Peak formation erode easily, leaving isolated exposures of the more resistant megabreccia. Only one outcrop displays the interbedded nature of the megabreccia with the formation (Fig. 3 and 6c). At this outcrop, the megabreccia is clearly interbedded with the Castle Rock member approximately 2 m below the correlative ash unit. The megabreccia deposits are tabular with thicknesses ranging from 1 to 20 m and are laterally continuous for up to 75 m. The megabreccia deposits are typically massively bedded, clast supported and monolithic with angular clasts 1 to 3000 mm in diameter. Clasts are either limestone derived from the Bird Spring Formation, or they are quartz arenites derived from the Stirling Quartzite and Wood Canyon Formation (Fig. 6d). In some of the megabreccia outcrops,
Figure 8. Rare earth element graph of two volcanic ash units. The graph illustrates the similarity of rare earth element concentrations in tuff samples located to the north and south of Fossil Ridge. This similarity suggests that these two tuff samples are from the same ash horizon, the correlative ash unit.
a crude ghost stratigraphy of the original formation can be observed in areas of up to 10 m in diameter (Fig. 6e). Along strike and adjacent to many of the megabreccia units, matrix-supported breccias are interbedded with limestone. Typically the matrix is carbonate mudstone, and clasts are subangular, range from 1 to 300 mm in diameter, and are composed of the adjacent megabreccia lithology (Fig. 6f).

**Quaternary deposits**

Quaternary alluvial-fan gravels overlie with angular unconformity the top of the Gass Peak formation. In the northwest portion of the study area, Maldonado and Schmidt (submitted) assigned a Tertiary age to poorly cemented gravels suggesting that they conformably overlie the Gass Peak formation because they have dips of 20°-30°, and the gravels are juxtaposed with the megabreccia deposits (Fig. 3). After detailed examination, I conclude that the contact between these gravels and the megabreccia is not exposed, and thus their depositional relationship is equivocal. The lithology, rounding, sorting, size, and cementation of the gravels is identical to that of the Quaternary alluvial-fan gravels. The dips of the gravels could be primary depositional dips and not tectonic. Thus, I argue for a Quaternary rather than Tertiary age assignment for these gravels.
Age of the Gass Peak formation

The age of the Gass Peak formation is Middle Miocene based on radiometric dating of volcanic ash units that range from 16 Ma to 12.1 Ma. These radiometric dates, which are based on previous sampling and sampling from this study, are the best tool for the chronological correlation of the Gass Peak formation with other sedimentary and volcanic deposits in southern Nevada.

Initially, Ebanks (1965) collected both ostracod and volcanic ash samples from the Tertiary rocks in the study area to determine the relative and radiometric age of the rocks. The ostracod samples were examined by Swain (1965, written comm. to Ebanks) (Appendix A) and determined to be of Oligocene to Miocene age. The volcanic ash sample yielded dates of 15.2 Ma and 15.9 Ma from K-Ar dating of biotite grains by J.F. Sutter (1968, personal comm. to Ebanks) (Appendix B). However, the geographic and stratigraphic location of these samples are not known. Based on the fossil type and age of the ash, it is probable that both of these samples are from the Fossil Ridge member.

Biotite separates from ash units in both the Fossil Ridge and Castle Rock members were collected during this study and were dated using conventional K-Ar analysis by Robert Fleck at the U.S.G.S. (1988, written comm.). The ash sample dated from the Fossil Ridge member (sample T7, Fig. 7), which occurs 9 m stratigraphically above the base of the
measured section, yielded an age of 16.0 Ma. The ash dated from the Castle Rock member (sample T13) is from the correlative ash unit (Fig. 7), and it yielded an age of 12.1 Ma. A biotite separate of this ash was K-Ar dated previously by Fleck (1988, personal comm.), and its age was determined to be 13.6 Ma ±0.3 Ma. This apparent discrepancy is probably due to the fluvial mixing of ashes of different ages prior to deposition, which is consistent with the abundant terrigenous and plant material within the ash unit. Therefore, the 12.1 Ma date indicates that the ash deposit must be no older than 12.1 Ma. Separate grain analysis, which is in progress, is necessary to more accurately determine the age of the volcanic material.

Regional correlation

This study confirms that the Gass Peak formation is generally temporally correlative with the Miocene Horse Spring Formation of the Lake Mead region and with some of the Tertiary basinal deposits of the Sheep Range extensional allochthon as previously suggested by Ebanks (1965), Longwell and others (1965), and Guth and others (1988). Nevertheless, from the refined stratigraphy of the Gass Peak formation given in this paper, it can be shown that these sediments are not, in their entirety or in part, lithologically correlative.
Radiometric ages suggest that the Fossil Ridge and Castle Rock members of the Gass Peak formation are, at least in part, temporally equivalent to the Thumb and Lovell Wash Members of the Horse Spring Formation. The 75-m-thick Fossil Ridge member containing a 16 Ma ash bed is in part equivalent to the 850-m- to 1300-m-thick Thumb Member of the Horse Spring Formation (Bohannon, 1984), the age of which is bracketed between 17.2 Ma and 13.5 Ma based on radiometrically dated tuff units. The Thumb Member is chiefly terrigenous, consisting of sandstone, conglomerate, siltstone, breccia, minor unfossiliferous limestone, and it contains thick beds of pure gypsum at many localities (Bohannon, 1984). The Fossil Ridge member of the Gass Peak formation, however, is dominantly fossiliferous limestone with only minor conglomerate and no gypsum. Thus, the Fossil Ridge member of the Gass Peak formation and the time equivalent Thumb Member of the Horse Spring Formation are not lithologically similar and are not the same lithostratigraphic unit.

The 260-m-thick Castle Rock member of the Gass Peak formation, which is at least in part younger than 12.1 Ma, may be temporally equivalent to the Lovell Wash Member of the Horse Spring Formation (Bohannon, 1984). Radiometric dating of tuff units bracket the age of the Lovell Wash Member between 13.0 Ma and 11.9 Ma. The 250-m- to 450-m-thick Lovell Wash Member is chiefly white limestone and
dolomite, gray claystone, gray tuff, tuffaceous sandstone, and minor conglomerate (Bohannon, 1984). The thickness of the Lovell Wash Member and the Castle Rock member are similar and the lithologies are, in a general sense, comparable. However, gypsiferous beds and megabreccia deposits present in the Castle Rock member are absent in the Lovell Wash Member as are fossil palm wood and palm leaves. I strongly suggest that the Castle Rock member of the Gass Peak formation and the time equivalent Lovell Wash Member of the Horse Spring Formation are not the same lithostratigraphic unit.

Radiometric ages also suggest that the Gass Peak formation may be temporally equivalent to nearby Tertiary basinal sediments of the Sheep Range extensional allochthon, the Black Hills basin and the Wamp Spring basin. The Black Hills basin sediments are divided into three informal members: the lower, Basin Canyon member; the middle, Quijinup Canyon member; and the upper, Dead Horse Trail member (Guth and others, 1988). The radiometric age range of each formation is not known, but the whole sequence is radiometrically dated between 16 Ma and 12 Ma (Guth and others, 1988). The 300-m-thick Basin Canyon member consists of fine-grained, tuffaceous, claystone and siltstone with minor gypsiferous and calcareous beds. This member is not lithologically correlative with the lower, Fossil Ridge member of the Gass Peak formation because the Fossil Ridge
member does not contain gysiferous units and is dominantly limestone. The 230-m-thick Quijinump Canyon member is composed of megabreccia deposits consisting of resedimented Ordovician and Silurian formations interbedded within the Basin Canyon member. One of these megabreccia deposits is radiometrically dated at 14-15 Ma. Based on this age and the composition of the megabreccia clast, this member does not correlate either lithostratigraphically or temporally with the Castle Rock member. The 1,200-m-thick Dead Horse Trail member consists of poorly sorted, vaguely bedded, pebble to cobble conglomerate with the clasts derived from the Ordovician Pogonip Group through Silurian Laketown Dolomite. The Gass Peak formation does not contain any member with this extreme thickness of conglomerate nor with conglomerates dominated by these clast types and, therefore, does not lithostratigraphically correlate with the Gass Peak formation.

Tertiary sediments of the Wamp Spring area are 365 m to 475 m thick and are composed of poorly sorted, pebble to boulder conglomerate with minor beds of volcanic ash (Haslet and others, 1981; Mansholt, 1983). Most of the conglomerate clasts were derived from the Bird Spring Formation with lesser amounts derived from the Wood Canyon Formation and the Stirling Quartzite. Radiometric ages of ash units from these sedimentary rocks range between 16 Ma and 13.8 Ma (Guth and others, 1988). These dates suggest a temporal
equivalence with the Gass Peak formation, but the Gass Peak formation does not contain thick units of conglomerate nor does it contain conglomerate of the same composition. Therefore, a lithostratigraphic correlation of the Tertiary sediments in Wamp Springs and those of the Gass Peak formation should not be made.

**LITHOFACIES AND DEPOSITIONAL PROCESSES**

Five lithofacies (Fig. 9) indicative of marginal lacustrine deposition occur within the Gass Peak formation: the bioclastic limestone lithofacies; the carbonate-clast-dominated conglomerate lithofacies; the thin-bedded limestone, sandstone, and siltstone lithofacies; the quartz arenite-clast-dominated conglomerate lithofacies; and the megabreccia lithofacies. Understanding the depositional processes of these lithofacies and the environment in which they were active is important in developing a depositional model and basin history. The recognition of lithofacies and their interpretation is based on fossil content; mineralogy and lithology; sedimentary structures; clast size, type and imbrication; and vertical and horizontal sequences.

**Bioclastic limestone lithofacies**

The bioclastic limestone lithofacies occurs throughout the Fossil Ridge member (Fig. 9 and 10), and both its fossils and lithology suggest deposition in the littoral
Figure 9. Lithofacies map of the Gass Peak formation. Bioclastic limestone lithofacies is marked in a brick pattern. Carbonate-clast-dominated conglomerate lithofacies is shown as a stippled pattern. Thin-bedded limestone, sandstone, siltstone lithofacies is marked by horizontal lines. Quartz arenite-clast-dominated conglomerate is marked with vertical lines. Megabreccia lithofacies is marked in a solid pattern.
Figure 10. Stratigraphic occurrence of the lithofacies of the Gass Peak formation.
zone of a permanent fresh-water lake. It is composed of thin- to medium-bedded, white- to cream-colored, bioclastic lime wackestone to mudstones (Fig. 5a).

The bioclasts consist of whole and broken ostracods, gastropods, and plant fragments. The ostracod genera *Cypricercus* and *Heterocrvpris*, identified by Swain (Appendix A), are indicative of a eutrophic, littoral lacustrine environment having alkaline pH and positive Eh conditions (Swain and others, 1970). The plant fragments, resembling genus *Typha*, are oriented not only parallel to bedding but locally are perpendicular to it; the latter represents an in situ growth position and is most characteristic of the littoral part of lacustrine environments where rooted aquatic vegetation occurs (Dean and Fouch, 1983).

The limestone itself is locally rippled and is composed of a minimum of 80 percent calcite as determined by the insoluble residue method (Ireland, 1971), and the remaining 20 percent is terrigenous sediment and secondary silica. These percentages are consistent with the high percentage of calcite (>60 percent) commonly recorded from littoral deposits of carbonate-rich lakes (Dean, 1981) where carbonate precipitation is stimulated by animals with carbonate shells and plants, as well as by algal and macrophyte photosynthesis. As would be expected in such a setting, the limestone locally contains cryptomicrobial
stromatolites displaying small-scale laminated, concave-downward structures (Fig. 11a). These structures represent carbonate precipitation in a shallow littoral environment nearly devoid of terrigenous input (Dean and Fouch, 1983). Symmetrical ripple laminations with ripple indexes of 6 to 7 and linear crest up to 600 mm in length are rare in the bioclastic limestone lithofacies. Nevertheless, this type of wave-formed ripple lamination usually is restricted to water depths of a few meters, but during storms can form at depths up to 10 to 15 m (Allen, 1982). Therefore, they are consistent with deposition in the shallow, coastal regions of lakes (Reineck and Singh, 1973).

During deposition of this lithofacies, the lake water was permanent and fresh. In addition, the salinity of the lake was fairly constant because evaporite minerals are absent. The abundance of fresh-water organisms (plants, ostracods, gastropods) and the lack of mud-cracks in the bioclastic limestone lithofacies also suggest the sediments were not subjected to desiccation.

**Carbonate-clast-dominated conglomerate lithofacies**

The carbonate-clast-dominated conglomerate lithofacies occurs in the upper part of the Fossil Ridge member (Fig. 9 and 10) and represent distal, coarse-grained, subaqueous debris flows of lacustrine deltas. It is composed of massive to poorly graded, clast-supported conglomerate with
Figure 11. Lithofacies of the Gass Peak formation.

(A) Cryptomicrobial stromatolites displaying small-scale laminations. Scale bar divisions are in inches. (B) Soft sediment folds displaying disharmonic folding. Knife is 10 cm in length. (C) Soft-sediment faults. Knife is 10 cm in length. (D) Channelized and loaded conglomerate and limestone contact. Hammer is 34 cm in length. (E) Linear orientation of palm fragments along a bedding surface. Scale bar 10 cm. (F) Conchoidally fractured dolomite. Hammer is 34 cm in length.
abundant carbonate clasts (Table 3), and it is interbedded and interfingers with the bioclastic limestone lithofacies (Fig. 5c and 10). The upper and lower contacts are sharp, planar and nonerosional, but small-scale ball-and-pillow structures occur locally along their base. These structures suggest rapid sedimentation on and loading of soft, liquid-saturated sediments (Reineck and Singh, 1973; Allen, 1982). The conformable nature of the conglomerate beds with the bioclastic limestone lithofacies, in addition to the lack of dissolution or desiccation features in the limestones, suggest that the conglomerate was deposited in a subaqueous lacustrine environment. This interpretation also is substantiated by the character of the conglomerates themselves; although, these are less definitive. The conglomerate bedding thickness ranges from 0.2 to 1.2 m with lateral continuity up to 30 m. Interbedded limestone units range in thickness from 0.2 to 2 m producing conglomeratic sequence up to 10 m in thickness. The conglomerate is poorly lithified, poorly sorted and clast-supported. Its clasts are sub-rounded and range in size from 3 to 300 mm (Fig. 5c). Internally, some of the conglomerate beds display crude horizontal and cross-stratification and poor clast imbrication. The clast imbrication data is scattered (Fig. 9 and 12a), but statistically (Krause and Geijer, 1987) it shows a dominant flow direction from the west. The matrix of the conglomerate is composed of sand- to silt-
Table 3. Conglomerate clast counts from the Gass Peak formation.

Conglomerate clast counts from the Fossil Ridge member

<table>
<thead>
<tr>
<th>station</th>
<th>quartz arenite</th>
<th>carbonate</th>
<th>chert</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>63</td>
<td>275</td>
<td>73</td>
</tr>
<tr>
<td>2</td>
<td>37</td>
<td>265</td>
<td>71</td>
</tr>
<tr>
<td>3</td>
<td>38</td>
<td>333</td>
<td>24</td>
</tr>
<tr>
<td>4</td>
<td>24</td>
<td>372</td>
<td>25</td>
</tr>
<tr>
<td>total</td>
<td>162</td>
<td>1245</td>
<td>193</td>
</tr>
<tr>
<td>total %</td>
<td>10%</td>
<td>78%</td>
<td>12%</td>
</tr>
</tbody>
</table>

Conglomerate clast counts from the Castle Rock member

<table>
<thead>
<tr>
<th>station</th>
<th>quartz arenite</th>
<th>carbonate</th>
<th>chert</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>375</td>
<td>15</td>
<td>10</td>
</tr>
<tr>
<td>2</td>
<td>365</td>
<td>32</td>
<td>3</td>
</tr>
<tr>
<td>3</td>
<td>367</td>
<td>30</td>
<td>3</td>
</tr>
<tr>
<td>4</td>
<td>382</td>
<td>8</td>
<td>10</td>
</tr>
<tr>
<td>total</td>
<td>1489</td>
<td>85</td>
<td>26</td>
</tr>
<tr>
<td>total %</td>
<td>93%</td>
<td>5%</td>
<td>2%</td>
</tr>
</tbody>
</table>
Figure 12. Composite rose diagram for conglomerate clast imbrications for the Gass Peak formation. Both rose diagrams are constructed from lower hemisphere steronet plots of poles to A-B planes of clasts. Petal interval is 10 degrees with the petal area proportional to the number of clasts in each petal. (A) Composite rose diagram of clast imbrications for the Fossil Ridge member. Grand vector mean is 95 degrees ± 77 degrees. Number of clasts is 161. (B) Composite rose diagram for the Castle Rock member. No valid grand mean vector can be calculated from plot. Number of clasts is 94.
sized particles of the same composition as the clasts. Small siltstone and sandstone lenses up to 0.4 m thick are interbedded with the conglomerate sequences, and individual lenses display either massive bedding, parallel lamination, or planar cross-stratification.

The sheet-like nature of the conglomerate beds with nonerosive bases, minor grading, planar cross-stratified sandstone lenses, and poor internal stratification of the conglomerate are similar to subaqueous gravelly mass-flow conglomerates described by Nemec and Steel (1984) and Nemec and others (1984). These types of conglomerates typically are deposited by cohesionless debris flows. The carbonate clast-dominated conglomerate lithofacies lacks large forsets that are typical of coarse-grained Gilbert-type deltas that develop in deep water lakes (Galloway and Hobday, 1983). This lack of forsets suggests shallow water at the site of deposition as does the associated bioclastic limestone lithofacies. Even though deposited in a marginal lacustrine environment, the lack of terrigenous material in the bioclastic limestone lithofacies suggests that the terrigenous sediment was not reworked after deposition and requires that the streams or transporting agent of the clastic material be flashy or episodical. The nearly instantaneous delivery of abundant coarse clastic material probably produced mass-flows that transported the conglomerate into the lake and, perhaps in part because of
the gentle slope of the lake bottom, it came to rest in moderately shallow water. Within the lacustrine system, wave energy was not sufficient to rework or significantly transport the coarse-clastic material. This situation is common in lacustrine environments, and consequently the margins of many lakes are marked by coarse clastics (Picard and High, 1981).

The carbonate-clast-dominated conglomerate lithofacies resemble deposits of lacustrine fan deltas as described by Nemec and Steel (1984) and Nemec and others (1984). Regretfully, neither the subaerial component of the Gass Peak formation deltas nor a fluvial system has been recognized in the study area, and therefore, it cannot be determined whether the conglomerate represents fan-delta or braid delta deposits (McPherson and others, 1987). Nevertheless, the coarse, immature nature of the clasts suggest the streams transporting the material to the lake were young streams or braided streams on the distal part of alluvial fans. The transport distance may not have been great since carbonate clasts may become rounded with only 7 km of transport (Pettijohn, 1975). The head water may have eroded Ordovician to Devonian rocks to produce the clast composition in the conglomerate (Fig. 2). Although abundant non-diagnostic carbonate clasts are present, the fossiliferous limestone clasts are chiefly composed of crinoid debris that may have been derived from either the...
Mississippian Joana Limestone or the Bird Spring Formation (Fig. 2). The Joana Limestone is more likely, however, because the Bird Spring Formation contains abundant fusulinids in addition to crinoid fragments, and no fusulinids were found in any of the conglomerate clasts. Thus, a local Ordovician through Mississippian provenance is suggested for the conglomerates. The less common white, sub-rounded, fine-grained quartz arenites are similar to units in the Ordovician Eureka Quartzite and in the Devonian Devils Gate Limestone and Oxyoke Canyon Sandstone (Fig. 2).

**Thin-bedded limestone, sandstone, and siltstone lithofacies**

The thin-bedded limestone, sandstone and siltstone lithofacies occurs in the lower parts of the Castle Rock member (Fig. 9 and 10) stratigraphically below the quartz arenite-clast-dominated conglomerate lithofacies and is interbedded with the megabreccia lithofacies (Fig. 10). It is composed of interbedded unfossiliferous lime mudstone, sandstone and siltstone with minor gypsiferous intervals (Fig. 5d). Bedding is thin and commonly displays small-scale, soft-sediment folds and faults. The lithologies and sedimentary structures of this lithofacies suggest deposition in a shallow marginal lacustrine environment. The lack of fossils and the presence of gypsum suggest that salinity of the lake waters were periodically elevated.
The lime mudstones are white to cream in color and contain no fossils or allochemical particles. Sedimentary structures within the limestone include ball-and-pillow structures (up to 1 m thick and up to 2 m in length) and small-scale soft-sedimentary folds and faults (Fig. 11b). The soft-sedimentary folds have amplitudes up to 300 mm and display a disharmonic style of deformation. Their fold axes trend in a northwest and southeast direction (Fig. 11c). The associated small-scale faults display normal displacements of up to 400 mm and strike in a northwest direction (Fig. 11c). The folds and faults are interpreted as developing shortly after deposition while the sediment was very ductile because of the style of deformation and the undeformed nature of the overlying sediments (Allen, 1982). The soft-sediment folds and faults seemingly are the result of strata moving down a gentle slope with the movement being initiated by either seismic shock or sediment loading. These mechanisms also may have produced the ball-and-pillow structures (Howard and Lohrengel, 1969).

The thin-bedded sandstones and siltstones are tan, very fine- to coarse-grained, and poorly sorted with sub-angular to sub-rounded grains. They display ball-and-pillow structures and small-scale soft-sediment folds and faults similar to those in the limestone. They have symmetrical ripple laminations with ripple indexes of 8 to 10 and linear crests up to 450 mm in length. These wave ripples are
similar to those found in the bioclastic limestone facies and are common in the margins of lacustrine environments (Reineck and Singh, 1973).

Gypsum occurs as medium to coarse sand-sized grains in the siltstone and sandstone sections and as 1 to 200 mm selenite crystals in the limestone units. Many of the selenite crystals have been removed by dissolution, leaving only selenite crystal-shaped voids (Fig. 5e). Selenite crystals of this type typically form from the capillary evaporation of brines in saline lacustrine mud flats (Eugster, 1980). If this were the process in the Castle Rock member, then these sediments were subaerially exposed sometime shortly after deposition. This does not appear to be the case, because other evidence of subaerial exposure, including mudcracks, flat pebble conglomerates, and paired evaporitic and clastic varves typical of playa lake complexes, are not present (Lowenstein and Hardie, 1985). Thus, the gypsum may have precipitated in a subaqueous environment, but elevated salinities are required for its precipitation and preservation.

Collectively, the sedimentary structures in this lithofacies establish a northwest trending lake margin and a northeast sloping lake bottom (Fig. 13). Fold axes and the strike of fault planes of soft-sediment deformational features typically are oriented parallel to the basin margin (Potter and Pettijohn, 1977). Using these criteria, the
Figure 13. Stereonet diagram for sedimentary structures. Stereonet is a lower hemisphere plot. Note that the sedimentary structures suggest a northwest trending basin margin with the basin sloping toward the northeast.
TREND AND PLUNGE OF
- FOLD AXES OF SOFT SEDIMENT FOLDS
- TREND OF RIPPLE MARK CRESTS
- DIP DIRECTION OF SOFT SEDIMENT FAULTS
- LINEATION OF PLANT FRAGMENTS
fold and fault orientations from the southern margin of the study area (Fig. 13) indicate a general paleoslope direction to the northeast. Symmetrical ripple laminations in lacustrine environments form roughly parallel to the lake margin (Picard, 1967). Ripple laminations from the south margin of study area (Fig. 13) imply a northwest lake margin. These sedimentary features are restricted to the southern outcrops of the Castle Rock member.

**Quartz arenite-clast-dominated conglomerate lithofacies**

The quartz arenite-clast-dominated conglomerate lithofacies occurs in the upper parts of the Castle Rock member (Fig. 9 and 10) and represents distal, coarse-grained, subaqueous debris-flow deposits of lacustrine deltas. It is composed of clast supported conglomerate with interbeds of volcanic ash, limestone, dolomite and siltstone. The conglomerate is composed of sub-rounded to sub-angular clasts that range in size from 3 to 500 mm, with a mode increasing upward from 50 to 90 mm. The clasts are dominantly maroon to white, medium- to coarse-grained, quartz arenite (Table 3). Quartz arenite of this type occurs in the Stirling Quartzite and the Wood Canyon Formation (Fig. 2). If the clasts were first cycle sediments, they would require on the order of 300 km of transport (Pettijohn, 1975). If the clasts were second cycle sediments, it could be possible that they were derived
from a proximal source such as the synorogenic Cretaceous conglomerate of Brownstone Basin. Fossil palm wood is present in some conglomerate units (Fig. 6a), suggesting the conglomerate material was derived, in part, from a subaerial source.

The maroon to white conglomerate beds range in thickness from 0.1 to 2 m. The conglomerate beds can be traced along strike for up to 40 m before they become covered. Internally the conglomerate beds are massive to normally graded (Fig. 5f). The lower and upper contacts are sharp and planar with minor channelized bases and loading structures (Fig. 11d). These conglomerates are similar to those in the carbonate-clast-dominated conglomerate lithofacies except that these commonly display normal grading. The relatively nonerosive and sheetlike nature of the conglomerate suggest a laminar mass-flow origin (Nemec and Steel, 1984; Nemec and others, 1984), and probably they were transported as cohesionless debris flows. The normally graded beds suggest that some of the debris flows became more turbulent liquified flows and dropped sediment by direct settling in a subaqueous environment (Lowe, 1982). Similarly, the change from debris flow to a liquified flow is most common in subaqueous settings where the added water content of the flow decreases its strength and viscosity causing increased turbulence in the flow. This mechanism of deposition also is consistent with the poor imbrication
(Fig. 9 and 12b) because in such flows turbulent traction transport of clasts is insignificant.

Interbedded white volcanic ash has a median grain size of 0.05 mm and a composition as described in the stratigraphy part of this paper (Table 1 and 2). The beds of volcanic ash range in thickness from 0.2 to 3 m and are internally planar or ripple laminated. The ripple laminations are symmetrical, and they have ripple indexes of 8 to 10 and linear crest up to 500 mm in length. These wave ripples indicate lake water depths of only a few meters (Allen, 1982). Rare ball-and-pillow structures occur in some ash beds and range in size from 0.2 to 0.3 m in thickness and width. Whole to broken palm leaves of *Sabalites* are preserved at some ash locations suggesting the ash was transported, at least in part, down a fluvial system. Most of the palm leaves have a preferred linear orientation along bedding planes orthogonal to the trend of the ripple crests (Fig. 11e). The plant orientations suggest that the ash was deposited rapidly from down-slope currents and later reworked by wave actions.

Tertiary volcanic centers are not known locally, so the volcanic ash must have been derived from a distant volcanic source. Median grain size (0.05 mm) of the ash suggests that the maximum distance to the source was approximately 100 km (Fisher and Schmincke, 1984, p.155). Using this criterion and the age of the tuffs, possible source areas of
similar age would include the Lake Mead, Caliente/Kane Springs Wash, and the Nevada Test Site areas (Longwell and others, 1965; Tschanz and Pampeyan, 1970; Cornwall, 1972). The Lake Mead area volcanics are an unlikely source because they lack large-volume pyroclastic deposits. No attempt has been made to correlate the ash beds of the Gass Peak formation with specific volcanic units in the Caliente/Kane Springs Wash area or the Nevada Test Site area.

The tan siltstone interbeds of this lithofacies range in thickness from 0.2 to 9 m and are internally massive, planar laminated, or ripple laminated. The ripple laminations are identical to those in the volcanic ash interbeds, and the siltstones also contain oriented palm leaves. Thus, like the ash, these beds were derived from a fluvial source and deposited by downslope currents and later reworked by wave actions.

The white limestone and dolomite interbeds range from 1 to 21 m in thickness (Fig. 11f). They are massive lime mudstones and dolomitic mudstones with no fossils or allochems. Because of their fine-grained texture and homogeneity, the dolomite beds fracture conchoidally. The carbonates, based on their laterally continuous nature and association with other nonmarine lithofacies, are interpreted as forming in a lacustrine setting. As in the carbonate-clast-dominated lithofacies, no evidence for subaerial exposure of the carbonates are present. The light
color, conchoidal fracture, and overall fine-grained nature of the dolomite suggests the dolomite may have formed by penecontemporaneous replacement of pre-existing limestone units (Wolfbauer and Surdam, 1974; Weber, 1964). The lack of lacustrine flora and fauna and the presence of dolomite imply the lake waters were saline at the time of deposition (Eugster, 1980).

The quartz arenite-clast-dominated lithofacies is interpreted as representing distal, coarse-grained, subaqueous debris flows from deltas along the margin of a lake whose salinity was variable. The deltaic sedimentation would be similar to that described in the carbonate clast-dominated lithofacies, but the source area for the clasts was different and these flows were more turbulent and less viscous.

**Megabreccia lithofacies**

The megabreccia lithofacies is composed of monolithic, well-cemented breccia that occurs as tabular, massively-beded deposits that range from 1 to 20 m in thickness. They are laterally continuous for up to 75 m and are interbedded with the thin-beded limestone, sandstone, and siltstone lithofacies (Fig. 9 and 10). This unusual lithofacies seemingly represents rock-fall avalanche debris derived from evolving fault scarps and was initially transported subaerially and then subaqueously before it came
to rest in the littoral lake environment. Ebanks (1965) previously interpreted the megabreccias as fault emplaced slideblocks. However, the breccia does not cross-cut the Gass Peak formation but instead is interbedded and compositionally grades into lithofacies interpreted as lacustrine deposits.

The monolithic megabreccias are clast supported and are either composed of quartz arenite clasts derived from the Stirling Quartzite and Wood Canyon Formation or carbonate clasts derived from the Bird Spring Formation with very little mixing of these two lithologies. The matrix is composed of fine-grained, silt- to sand-sized fragments of the same lithology as the larger clasts. Clasts in the breccia show rotation relative to each other and are 1 to 3000 mm in diameter and fit together with a three-dimensional jigsaw puzzle effect. Within some breccia units, a crude ghost stratigraphy of the original formation can be observed in patches up to 10 m in diameter (Fig. 6e).

Along strike with many of the breccia deposits, beds belonging to the thin-bedded limestone, sandstone, siltstone lithofacies interbedded with rudstone beds composed of clasts of the same lithology as those of the adjacent megabreccia. These rudstones are matrix-supported with sub-angular clasts ranging in diameter from 1 to 300 mm (Fig. 6f). Regretfully, the contact between the rudstone and the megabreccia is not exposed, but the gradational change along
strike suggest nearly synchronous deposition with the megabreccia being forcefully implaced into the lime-rich mud.

The monolithologic composition, geometry, and texture of the megabreccia units are very similar to some rock-fall-avalanche and rock-fall-landslide deposits. Such deposits are common in marginal lacustrine and playa deposits adjacent to local, active normal fault scarps (Moores, 1968; Moores and others, 1968; Krieger, 1977). Rock-fall avalanches or rock-fall landslides are free-falling, newly detached segments of bedrock that lack a basal slip plane. Their clasts, which can be of any size, are derived from a cliff, steep slope, cave or arch (Sharpe, 1960). If the rock-fall is large or the fall great, the rock-fall will be completely fractured on impact and will move outward from the base of the slope in a high velocity laminar flow (Shreve, 1966). The flow probably is the result of buoyancy given to the mass by air trapped and compressed within and beneath the rock-fall (Varnes, 1958). Rock-falls of this fashion can travel up to 15 km horizontally from their point of origin on moderate to horizontal slopes (Krieger, 1977). Deposits from these rock-falls are tabular and lobate in form; are composed of clast-supported, poorly sorted monolithic breccia showing rotation of angular clasts relative to one another; preserve parts of relic
stratigraphy; and have a matrix composed of smaller fragments of the same composition as the larger clast.

Thus, it is envisioned that the monolithologic breccias were rock-falls that derived debris from high relief source areas and carried it into a marginal lacustrine environment. The monolithologic nature of the megabreccia indicates that the source area for each deposit was composed entirely of a single lithology, and consequently there were separate source areas that produced the Stirling Quartzite monolithologic breccia and the Bird Spring Formation monolithologic breccia. Although the sources of debris were isolated from one another, they were in close proximity because the deposits accumulated in the same depositional basin after a short distance of transport. As the avalanche entered the lake, it must have disrupted in situ deposits of the thin-bedded limestone, sandstone, and siltstone lithofacies producing short-lived debris flows composed of the matrix-supported rudstones. These rudstones thus represent the distal or marginal subaqueous deposits of the megabreccia.

DEPOSITIONAL MODEL

A two stage lacustrine model is proposed for the Gass Peak formation that represents deposition in a shallow, tectonically active, permanent lake into which flashy discharging streams debouched coarse terrigenous sediment
(Fig. 14). This depositional model is based on a synthesis of the reconstructed stratigraphy, lithofacies and the interpretation of depositional processes, as well as paleocurrent and paleoslope data.

**Early-stage of deposition**

The first preserved record of basin accumulation is the thick section of shallow, marginal lacustrine limestone in the lower part of the Fossil Ridge member. These rocks indicate the existence of a fresh-water, carbonate rich, permanent lake, with abundant aquatic flora and fauna (Fig. 14). The rate of basin subsidence was approximately equal to the rate of deposition because the lithofacies and depositional environments are constant. The lake must have had outflowing streams in order to maintain the fresh-water chemistry throughout the deposition of the Fossil Ridge member. Deep-water deposits are not recognized in any part of the Gass Peak formation implying either the lake was small or the lake was never deep enough to produce profundal sediments.

Later during this stage, clast-supported subaqueous debris flows from small deltas deposited coarse clastic sediment episodically into the carbonate-dominated lacustrine environment. These subaqueous flows transported clasts southwesterly. The clasts were derived from local
Figure 14. Schematic diagram for the early-stage and late-stage of deposition. Diagram is in map view with location of basin margins and highlands generalized with present day flow directions. FR = present location of Fossil Ridge. GP = present location of Gass Peak. O-M = source area for Ordovician through Mississippian clasts. ZCs = Stirling Quartzite. €wc = Wood Canyon Formation. MPPbs = Birds Spring Formation.
A

SOURCE AREA
BASIN MARGIN
FLOW DIRECTION

KNOWN OUTCROP AREA
OF THE FOSSIL RIDGE MEMBER
LOCATION OF OUTLET IS UNKNOWN

FRESH-WATER LAKE WITH ABUNDANT AQUATIC FAUNA AND FLORA

△ GP

B

KNOWN OUTCROP AREA
OF THE CASTLE ROCK MEMBER

SOURCE OF ZCS AND MPPBS MEGABRECCIA UNKNOWN, BUT PROXIMAL △ GP

SALINE LAKE
highlands composed of Ordovician through Mississippian rocks.

**Late-stage of deposition**

Lithofacies of the Castle Rock member portray a shallow, somewhat saline lake into which coarse-grained clastic material in the form of conglomerate and megabreccia were carried from southwestern highlands and perhaps active fault scarps (Fig. 14). The end of this stage of deposition requires a more tectonically active extensional basin than does the early stage. Between the early and late stages, the basin must have become hydrologically closed. Saline minerals like gypsum and dolomite, in addition to the lack of fresh-water fossils, in the Castle Rock member indicate that the lake waters changed from fresh to saline. This hydrologic restriction of the basin could have been due to climatic changes or tectonism. Strong evidence for active extensional faulting during this time is present both regionally (Guth, 1980, 1981) and locally in the Gass Peak formation. I suggest that tectonism alone could account for the hydrological restriction by modifying the basin and highland configuration.

The initial sediments preserved from this stage were deposited near the lake margins and consisted of fine-grained terrigenous and carbonate material that was reworked by wave action and down-slope movements. The down-slope
movements were probably the results of sediment loading of liquid saturated sediments or seismic shock. Also during this time, the lake waters became saline enough to precipitate gypsum in the soft limestone, sandstone, and siltstone.

Near the end of this stage, braided streams carried coarse-grained clastics, palm wood and leaves, and volcanic ash northeastwardly into the lake margin creating deltas. Subaqueous debris flows from these deltas carried material down gentle slopes in a northeastward direction, and the coarse clastics were interbedded with the finer-grained lacustrine sediments. The sub-rounded clasts composed almost entirely of the Wood Canyon Formation and the Stirling Quartzite were either transported long distances to the deltas or were recycled from an earlier conglomeratic deposit. Megabreccia deposits transported by rock-fall landslides or avalanches also were carried episodically into the shallow margins of this lake. These deposits were derived exclusively from a proximal source of Stirling Quartzite, Wood Canyon Formation and Bird Spring Formation. The abundance of coarse debris during this time strongly implies an increase in relief between the highlands and the lake. Furthermore, it suggests significant erosion and deposition after active tectonic uplift of the source (Moores, 1968).
The presence of *Sabalites* palm trees implies the local Middle Miocene climate was arid subtropical with most of the precipitation (380 - 510 mm) occurring chiefly in the summer (D.I. Axelrod, 1987, written comm.). Such a rainfall could account for the permanent nature of the lake as well as the flashy nature of the discharging streams.

The termination of deposition in this lacustrine basin is not known because of the unconformable nature of the upper contact of the Gass Peak formation with the overlying Quaternary sediments. This unconformity is a moderate to high-angle angular unconformity implying that tectonism affected the Gass Peak formation after its deposition.

**TERTIARY STRUCTURE**

Oroflexural bending along the Las Vegas Valley shear zone and high-angle normal faulting have deformed the Gass Peak formation. The structural relationships, together with the lithofacies analysis, suggest that Tertiary lacustrine basins were initiated by broad downwarping or by early movement along major extensional faults and were later dismembered by smaller high-angle normal faults. Evidence for the faulting and folding of the Gass Peak formation are fault scarps, fault breccia, fault drag folds, and displaced lithofacies.
Normal faults

Four major normal faults (faults A-D in Figure 3) offset and deform the Gass Peak formation. All of these faults are interpreted as high-angle normal faults based on the displacement and drag of the Gass Peak formation along them and on restored cross-sections. Subsequent to normal movement, minor sinistral strike-slip movement occurred along fault B based on slickensides and fold orientations. This structural interpretation differs from the previous ones proposed by Ebanks (1965) and Longwell and others (1965).

Fault A is interpreted as a high-angle normal fault dipping northward based on apparent block motions and rotation of strata into the fault. Fault A is marked by a prominent fault scarp along the north flank of Gass Peak with a general west to east trend and a dip of 55° northward. This fault juxtaposed the Gass Peak formation in the hanging wall with the Stirling Quartzite or Wood Canyon Formation in the footwall, suggesting down-to-the-north displacement. The Gass Peak formation is not exposed south of this fault due to removal by erosion. Along the fault trace, the attitude of the Gass Peak formation varies from horizontal to a 30° dip southward into the fault indicating that the strata rotated into the fault.

Both faults B and C also are interpreted as high-angle normal faults even though fault B had minor sinistral
strike-slip movement subsequently. Faults B and C trend west along the south and north sides of Fossil Ridge, respectively. A minimum of 270 m of vertical displacement had to have occurred along both of these faults to account for the displacement of the thin-bedded limestone, sandstone, and siltstone lithofacies on either side of Fossil Ridge (Fig. 9). The 270 m of displacement is calculated on the present elevation difference between the top of Fossil Ridge and the outcrops of the Gass Peak formation. Fossil Ridge could not have been exposed during the time of deposition because the lithofacies indicate continuity across the area, no debris was derived from the ridge, and fluvial systems carried sediment across its present location. The fault trace lies along the contact between the Gass Peak formation and the Paleozoic bedrock and is probably continuous along the length of Fossil Ridge. The Gass Peak formation along fault B dips between 10° and 70° to the south, and along fault C it dips 70°-80° northward representing normal fault drag. One fault surface is observable along the southern flank of Fossil Ridge. The fault surface strikes N70°E and dips 83° S, with subhorizontal slickensides that trend S60°W and plunge 5° to 15°. The subhorizontal orientation of slickensides suggests that the last movement on fault B was dominantly strike-slip. Other evidence for strike-slip movement includes small folds and reverse faults immediately south of fault B.
Figure 15. Stereonet of tectonic folds and faults. Stereonet is a lower hemisphere plot. The orientation of folds and faults suggest both were produced from northeast-southwest shortening along fault B.
• TREND AND PLUNGE OF TECTONIC FOLD AXES

TECTONIC FAULT PLANES PLOTTED AS GREAT CIRCLES
These fold axes and faults trend approximately N55° W, forming a 35° angle with the trace of fault B (Fig. 15). The relative orientation of these fold axes to the fault trace strongly suggest drag along a sinistral strike-slip fault (Wilcox and others, 1973).

Using the interpretation of faults A and B as high-angle normal faults, a reconstructed north to south cross-section (Fig. 16) demonstrates that 30 percent extension has occurred between the upper contact of the Bonanza King Formation on Fossil Ridge and the Gass Peak thrust. Dips and stratigraphic thicknesses used in the reconstructions are based on the unfaulted stratigraphic sequence that occurs just northeast of the study area. Faults A and B apparently lack large, horizontal, north-to-south displacements or listric rotation of strata because Fossil Ridge and the Gass Peak thrust maintain the same relative structural and stratigraphic position compared to the unfaulted rocks just northeast of the study area.

Fault D is an implied normal fault striking northwest between the Gass Peak thrust and Fossil Ridge with down-to-the-west displacement. Although not exposed, the fault is required because the Gass Peak formation does not crop out east of this zone and an unfaulted Paleozoic sequence stands 250 m above this zone to the east. In addition, near this implied fault, the Gass Peak formation dips 20° southwest and decreases to horizontal away from the zone. Ebanks
Figure 16. Geologic N-S cross-sections. (A) Reconstructed cross-section before Tertiary extension. Stratigraphic and structural data is taken from the relatively unfaulted region just east of the study area. (B) Present day cross-section of the study area. Location of cross-section is marked on Figure 3. All unit symbols are listed in Figure 3.
(1965) mapped this fault but suggested that it extended northward across the east end of Fossil Ridge. Such a continuation of this fault is unlikely because a normal stratigraphic sequence is exposed along that end of Fossil Ridge, and there is no evidence of a parallel-to-bedding fault. The relationship of the junction of fault D with faults A and B is not apparent. It is possible that the ends of fault D curve and join faults A and B creating a scoop-shaped geometry.

The continuation of the Mormon Pass fault into the area north of Fossil Ridge and through Yucca Gap is uncertain because little stratigraphic separation or stratal rotation are present between Fossil Ridge and the south end of the Sheep Range (Fig. 3). Thus, the Mormon Pass fault must have produced little displacement at Yucca Gap, or alternatively, there must be a series of down-to-the-south normal faults north of Yucca Gap that juxtapose strata in such a way that little movement is apparent. Field work was not conducted to resolve these hypotheses, but the Mormon Pass fault can be traced to the east end of Fossil Ridge parallel to and associated with fault C. Possibly fault A is a splay or a synthetic fault to the Mormon Pass fault and must have little displacement near Yucca Gap.

A regional pattern of small-scale, north-trending, down-to-the-west normal faults exist along Fossil Ridge and in the southern Las Vegas Range. These faults may be
related to the oroflexing of the region as proposed by
Nelson and Jones (1987). A series of these down-to-the-west
normal faults deform the Gass Peak formation between Fossil
Ridge and Gass Peak. The amount of displacement of these
faults is small, and the lack of marker units in the Miocene
rocks and the alluvium preclude precise determination.
Nevertheless, some of these faults have minimum vertical
displacements of 5 m.

Timing of normal faults and oroflexural movement

Only the relative timing of movement along the four
major normal faults and the regional oroflexural bending of
the Las Vegas Range can be established by using regional
relationships and field data previously discussed in this
paper. Three of the major normal faults cutting the
formation trend east: an orientation perpendicular to the
regional normal faults of the Sheep Range extensional
allochthon (Guth, 1980, 1981). These relationships suggest
that the regionally oldest normal faults formed along north-
south trends and later were reactivated as sinistral strike-
slip faults and rotated into an east-west trend by
oroflexural bending along the Las Vegas Valley shear zone
(Ekren and others, 1968). This regional interpretation is
consistent with the slickensides and possible drag folds
along fault B that imply left-lateral movement after normal
faulting. Although this evidence is not strong, it does
suggest that the majority of oroflexural rotation of the southern Las Vegas Range took place after deposition of the Gass Peak formation and after the major movement along normal faults. Further supporting evidence may come from paleomagnetic studies on the Gass Peak formation which are currently in progress.

**SUMMARY OF TERTIARY GEOLOGIC HISTORY**

It is concluded from this study that Middle Miocene lacustrine sediments must have been deposited unconformably over the eroded hanging wall of the Gass Peak thrust, 16 Ma to 12 Ma, and were later (post 12 Ma) dismembered by extensional and strike-slip faulting. This geologic history suggests isolated lacustrine depocenters formed in southern Nevada during the onset of Tertiary extension and the resulting sedimentary sequences were dismembered by continued extensional tectonism. Furthermore, this study shows that extensional faulting in the Sheep Range extensional allochthon predated the oroflexural deformation along the Las Vegas Valley shear zone. These conclusions are based on the synthesis of the regional geology, Miocene stratigraphy, lithofacies, deposition model, and structural geology.

Regional mapping has indicated that Upper Proterozoic and Paleozoic rocks were thrust eastward along the Gass Peak thrust during the Mesozoic Sevier orogeny. During this
thrusting, the Wood Canyon Formation and Stirling Quartzite of the lower plate of the Gass Peak thrust and the Wheeler Pass thrust were subaerially exposed. Quartz arenite clasts were derived from these formations and deposited in Cretaceous synorogenic conglomerates of thrusts further to the east (Axen, 1984). After thrusting, the area of the Sheep and Las Vegas Ranges underwent extensive erosion (Guth, 1980, 1981) leaving the Mississippian Indian Springs Formation as the highest stratigraphic level in the hanging wall of the Gass Peak thrust and the Upper Proterozoic Stirling Quartzite exposed at its leading edge (Fig. 17a; Guth, 1980, 1981). The Middle Miocene Gass Peak formation was deposited unconformably over these rocks involved in thrusting and the present day location of the Gass Peak thrust.

Based on the structural interpretation of the study area, the Gass Peak formation is believed to have undergone a significant clock-wise rotation about a vertical axis after its deposition. The amount of rotation is interpreted to have been between 60° and 80° based on paleomagnetic data (Nelson and Jones, 1987) and the rotation of faults. Therefore, in the reconstructed history of the basin, I have chosen to rotate paleodirection indicators 70° counterclockwise from their present orientation.
Figure 17. Schematic block diagram for the deposition and subsequent deformation of the Gass Peak formation. Arrow with "M" represents a generalized Middle Miocene north direction for the Gass Peak formation. W and E represent present day west and east directions. O-D = Ordovician through Devonian rocks undivided. O = Ordovician rocks undivided. C = Cambrian rocks undivided. ZC = Upper Proterozoic rocks undivided. MIPP = Mississippian through Permian rocks undivided. Tgp = Gass Peak formation which is also represented by the stippled pattern.
A

PRE-16 Ma

B

16-12 Ma

C

POST 12 Ma

D

POST 12 Ma
The lower deposits of the Gass Peak formation were deposited in a shallow, fresh-water lake 16 Ma (Fig. 17b). The formation of the lake basin is interpreted to have formed from broad regional downwarping during the onset of extension in the Sheep Range extensional allochthon. This depocenter and its associated stratigraphic sequence were separate from the depocenter for the temporally equivalent Horse Spring Formation of the Lake Mead area as well as separate from other temporally equivalent units in the Sheep Range extensional allochthon. Bioclastic lime mud was deposited near the margin of the Gass Peak lake. The regional extent of this lake must have covered an area larger than the present outcrops of the Fossil Ridge member (6 km²). Gravels were carried into the lake from northeasterly flowing streams whose sources were the proximal highlands to the southwest. The paleocurrent directions and clast types of the conglomerates are consistent with the source area being the upper plate of the Gass Peak thrust.

Continued extensional tectonism caused the initial basin to become hydrologically restricted and the lake waters to become saline. During this stage, the Castle Rock member of the Gass Peak formation was deposited (Fig. 17b). Limestone, sandstone, siltstone and gypsum were deposited along the margins of a shallow, saline lake approximately 12 Ma. The minimum regional extent of this lake stage is
constrained by the present outcrops of the Castle Rock member (20 km²). The true extent of the lake at this stage probably was larger in extent because only marginal lacustrine deposits are recognized. Coarse clastic material both from rock-fall avalanches derived off of local high-relief areas and from braided-stream systems was deposited episodically along the margin of the lake forming megabreccia and conglomerates. The lithologies of the coarse clastic material suggest that the clasts were derived from the east or south.

The near monolithic nature of the sub-rounded Wood Canyon Formation and Stirling Quartzite quartz arenite clasts of the conglomerates suggests that the clasts are either a first cycle sediment derived from along the strike of the Gass Peak thrust where the formations presently crop out, or more likely, they are a second or multicycle sediment derived from an erosional surface conglomerate on the upper plate of the Gass Peak thrust or from an earlier conglomeratic deposit such as the synorogenic Cretaceous conglomerates. These ideas are supported by the fact that quartz arenite clasts need on the order of 300 km of transportation to become rounded (Pettijohn, 1975), and the present day source of the quartz arenites is only a few 100 m away from these conglomerates. Both the recycled and first cycle models have problems in explaining the near monolithic nature of the conglomerate because local
carbonate units, whose presents is indicated by the Bird Spring Formation megabreccia clasts, do not occur as clasts in the quartz arenite conglomerate. However, the recycled origin is favored because rounded Wood Canyon and Stirling Quartzite clasts are present in Cretaceous synorogenic conglomerates exposed to the south and to the east of the study area (Axen, 1984). Erosion of these deposits could produce a near monolithic source of quartz arenite with little transportation.

Megabreccia commonly forms along the margin of active fault margins or areas of great relief (Shreve, 1968). It is not clear which of these two origins produced the megabreccia deposits of the Castle Rock member. The megabreccia may have been derived from the active footwall of fault A (Fig. 3), but this fault at present bounds the marginal lacustrine deposits and to have sufficient vertical relief during the time of megabreccia deposition it certainly would have altered the depositional character of the lake. No other large vertical displacement faults occur within 15 km of the deposit nor displace the correct lithologies to produce the megabreccia. Another fault possibility could have been a southerly fault that cut the Wheeler Pass thrust before the thrust moved eastward along the Las Vegas Valley shear zone. This fault would lie presently beneath the Las Vegas Valley, and its existence is speculated only because it could have produced the
megabreccia deposits. The megabreccia may have been derived from highlands to the east where both the hanging wall and footwall of the Gass Peak thrust were preserved. A 6-m-wide zone of highly brecciated rock occurs along both sides of the west-dipping Gass Peak thrust (Ebanks, 1965, p. 23). This brecciated zone, if exposed further east at the time of deposition, could have served as a slip horizon from which a rock-fall avalanche could have been initiated with moderate relief.

Continued extension faulted and fragmented the Gass Peak formation on both sides of what is now Fossil Ridge and the area north (present day coordinates) of Gass Peak (Fig. 17c). Major displacement along the Mormon Pass fault in this area is interpreted to have occurred during this stage, post 12 Ma. Later, oroflexural bending, which was the result of drag along the right-lateral Las Vegas Valley shear zone, reactivated the north-striking, dip-slip faults as sinistral strike-slip faults and rotated them into their present day NE striking orientation (Fig. 17d). Quaternary alluvial fan sediments were unconformably deposited over the Gass Peak sediments and modern day erosion has exposed the present outcrops of both the fan sediments and the underlying Gass Peak formation.
DISCUSSION OF REGIONAL IMPLICATIONS

The geologic history proposed in this paper for the Miocene Gass Peak formation of the southern Las Vegas Range has broad implications for the timing and nature of Tertiary extension and strike-slip faulting in southern Nevada. New data on the timing of normal faulting within this region of the Sheep Range extensional allochthon indicates initiation approximately 16 Ma and continuation until after 12 Ma. Normal faulting produced by this extension is temporally equivalent with that in the Lake Mead region (Bohannon, 1984; Weber and Smith, 1987) and nearby basins in the Sheep Range extensional allochthon (Guth and others, 1988), but the resultant depositional basins within these areas were isolated. Consequently, the stratigraphy of the Gass Peak formation is unique and stands in marked contrast to that of the temporally equivalent Horse Spring Formation in the Lake Mead area and to the other nearby Tertiary basins. The unique depositional basin of the Gass Peak formation supports previously published models that the Sheep Range extensional allochthon was a separate extensional terrane from that of the Lake Mead area (Wernicke and others, 1984). Additionally, the age of the Gass Peak basin is younger than those to the west in the Sheep Range extensional allochthon supporting the model that extension of the allochthon
progressed from west to the east (Guth, 1980, 1981; Guth and others, 1988).

The majority of normal faulting within the Sheep Range allochthon occurred prior to significant oroflexural rotation along the Las Vegas Valley shear zone. Thus, these structural events appear to represent a continuum. The relative relationship would predict that the paleomagnetic signature of the Gass Peak formation would show an approximate 90° clock-wise rotation about a vertical axis. If, when tested, this prediction is correct, then oroflexural deformation along the Las Vegas Valley shear zone did not start until after 12 Ma. Furthermore, any model relating extension north of the Las Vegas Valley shear zone and right-lateral displacement along the zone would have to account for some oroflexural deformation that post-dated major normal faulting.

The nature of the shear zone and its relationship to extension could be refined by the following lines of investigation. (1) Radiometric dating and paleomagnetic analysis of volcanic ash units in older Quaternary ? or Tertiary ? gravels in the southern Las Vegas Range. (2) Sedimentation and tectonic studies of other Tertiary sediments along the shear zone within the Sheep Range extensional allochthon. (3) Field work to locate the other "half" of the Gass Peak basin, if it still exists, along the southern part of the Las Vegas Valley shear zone.
REFERENCES


Appendix A. A copy of the letter written by F.M. Swain to W.J. Jr. Ebanks, 1965 discussing the identification and age of ostracod samples from the Horse Spring Formation of the southern Las Vegas Range.
Mr. William J. Ebanks, Jr.
Department of Geology
Rice University
Houston 1, Texas

Dear Mr. Ebanks:

The collection of ostracodes you sent me from the Horse Spring Formation, Oasis Peak Quadrangle, Nevada contains the following:

Heterocypris sp. aff. H. incongruens (Ramdohr)
Cypricerous sp. aff. C. affinis (Fischer)
Amplocypris? sp. aff. A. sinuosa Zalanyi
Dogriinella sp. aff. D. taeniata Schneider

These all occur in the Lower Humboldt beds of central and northeastern Nevada and are possibly Miocene in age (Swain, 1964, Inter-mountain Assoc. Petrol. Geol., 13th Am. Field Conf. Guidebook, p. 176, pl. 2). A recent suggestion of an Oligocene age of part of the Lower Humboldt has been made. I would tentatively suggest an Oligo-miocene age for your collection. Typical Eocene species are not present in the sample.

With your permission I should like to keep this collection for further work.

Sincerely yours,

[Signature]

Professor

July 12, 1965
Appendix B. A copy of the letter written by J.F. Sutter to W.J. Jr. Ebanks, 1968 discussing the radiometric dating of biotite separates from the Horse Spring Formation of the southern Las Vegas Range.
April 26, 1968

Jim,

The following is the data from that volcanic tuff that has been so long overdue. I separated out the biotite and by looking at it under a binocular scope it appears to be about 80% pure with the other material being composite grains of biotite and plagioclase and maybe a few glass shards.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Mineral</th>
<th>Spike No.</th>
<th>%K</th>
<th>%Ar40*</th>
<th>Moles of Ar40*/gm K</th>
<th>Age in my x 10^-7</th>
</tr>
</thead>
<tbody>
<tr>
<td>EBANKS</td>
<td>(biotite)</td>
<td>774</td>
<td>4.60</td>
<td>48.6</td>
<td>2.719</td>
<td>15.25</td>
</tr>
<tr>
<td>EBANKS</td>
<td>(biotite)</td>
<td>780</td>
<td>4.60</td>
<td>32.7</td>
<td>2.815</td>
<td>15.25</td>
</tr>
</tbody>
</table>

Incidentally, the potassium was determined by both the Tetraphenyl Boron gravimetric procedure and the Atomic Absorption Flame Photometric procedure and these values agreed within 1% of each other.

Better late than never I guess. Hope this data fits in with some sort of interpretation.

As always,