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LATE CENOZOIC GEOLOGY AND TECTONICS OF STEWART AND MONTE CRISTO VALLEYS, WEST - CENTRAL NEVADA

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

by

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ABSTRACT

Discontinuous right-normal-slip faults comprise the N30°W trending, 45 kilometer long Stewart - Monte Cristo fault zone (SMCFZ). Initiation of the SMCFZ postdates the 15.5 to 11.0 m.y. Esmeralda Formation. Right-normal-slip $\geq$1 meter occurred on the southern segment of the SMCFZ during the 1932 Cedar Mountain earthquake (Ms = 7.2-7.3). Geomorphic evidence supports at least three and possibly five or six surface faulting events on the southern segment during the latest Pleistocene and Holocene. Gentle folds in the Esmeralda Formation east of and sub-parallel to the SMCFZ are coeval with and genetically related to faulting.

Structural development of the SMCFZ is similar to other right-lateral wrench faults and is consistent with laboratory wrench fault models. The SMCFZ is the youngest and southeasternmost fault of a system of major late Cenozoic, left-stepping, en echelon right-slip faults in the central Walker Lane.

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INTRODUCTION

Location and Extent of Study Area

This investigation covers an approximate area of 510 square kilometers (200 square miles) within Stewart and Monte Cristo valleys in eastern Mineral County and northeastern Esmeralda County, west-central Nevada (Figure 1). The area is accessible on well-graded dirt roads from State Route 361 at Finger Rock Wash, 18.5 km (11.6 miles) south of Gabbs, and from U.S. Highway 95 at Mina. Less accessible routes are possible through the Monte Cristo Range and Cedar Mountains. The area is open land under the jurisdiction of the Bureau of Land Management.

Regional Tectonic and Geologic Setting

Stewart and Monte Cristo valleys are located near the eastern margin of the central portion of the Walker Lane as defined by Locke, Billingsley, and Mayo (1940, p. 522-523). This northwest trending belt of irregular topography, situated between the Sierra Nevada and the north-northeast trending basins and ranges of the Great Basin, was first noted by Spurr (1901, p. 258) and was delineated and considered a major structural feature by Gianella and Callaghan (1934a, p. 18-22). Gianella and Callaghan considered the Walker Lane to be a major zone of right-slip in the western Great Basin and later studies have supported this interpretation (Ferguson and Muller, 1949; Nielsen, 1965; Shawe, 1965; Albers, 1967; Stewart and others, 1968; Bonham and Slemmons,

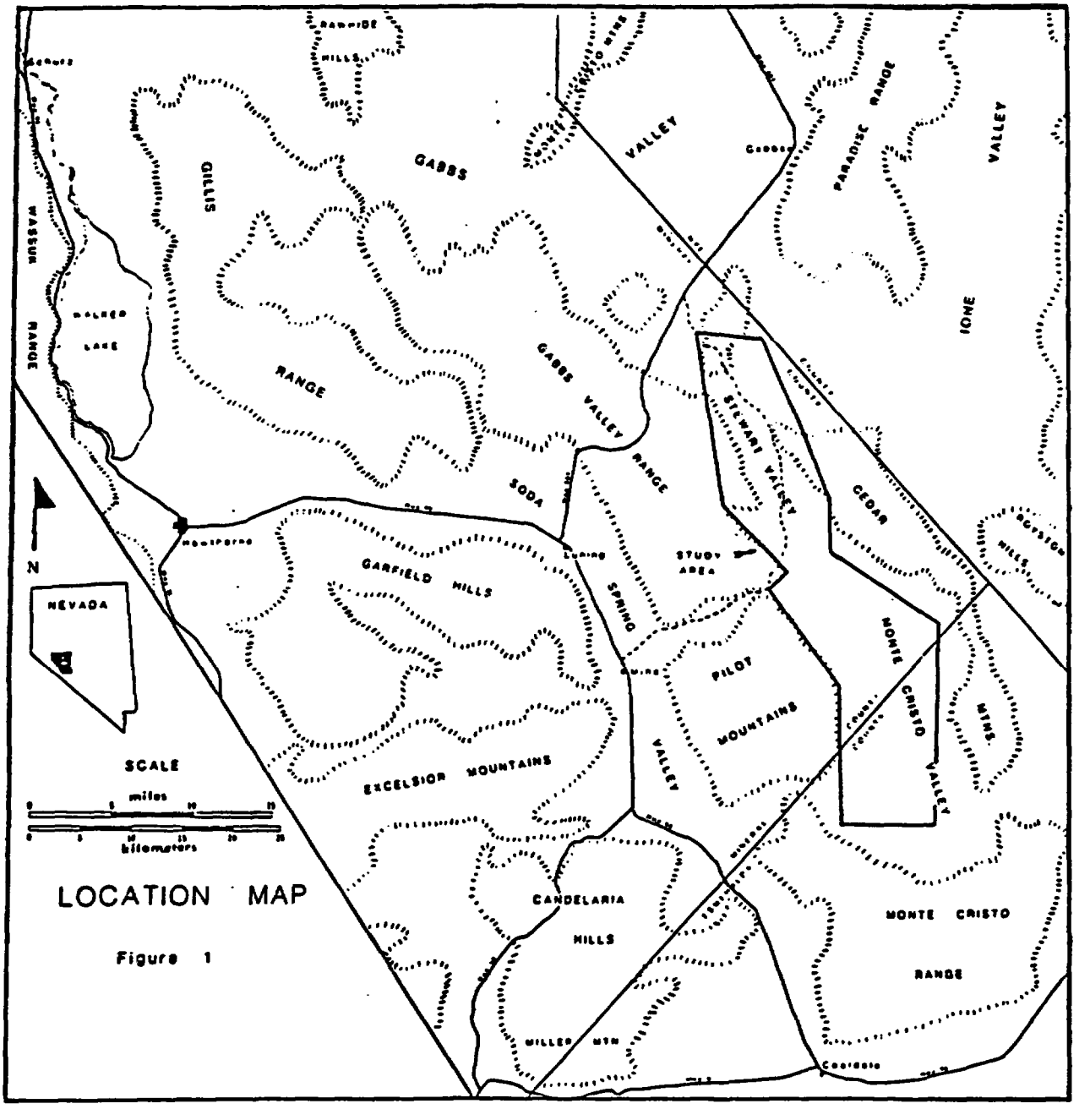

Figure 1 - Location map of Stewart and Monte Cristo valleys.

1968; Hardyman and others, 1975; Hardyman, 1978; Bell and Slemmons, 1979; Slemmons and others, 1979; Ekren and others, 1980; Hamel, 1983; Molinari, 1983). Several major east-to-northeast trending lineaments and left-slip fault zones are conjugate to the dominate northwest fault trends within the Walker Lane (Callaghan and Gianella, 1935; Shawe, 1965; Slemmons, 1967; Bingler, 1971; Albers and Stewart, 1972; Gilbert and Reynolds, 1973; Rogers, 1975; Ekren and others, 1976; Speed and Cogbill, 1979; Sanders and Slemmons, 1979; Moore, 1981; Rowan and Wetlaufer, 1981). Figure 2 shows the spacial relationship of the study area to the major faults and lineaments of the Walker Lane.

The Stewart - Monte Cristo valleys area was the site of the Ms = 7.2 to 7.3, December 20, 1932, Cedar Mountain earthquake (Gianella and Callaghan, 1934a, b; Byerly, 1935; Wilson, 1936; Slemmons and others, 1965; Coffman and von Hake, 1973) and continues to have seismicity (Gumper and Scholz, 1971; Ryall and Priestley, 1975).

The geology of the surrounding mountain ranges includes Mesozoic sedimentary, volcanic, and plutonic rocks and Oligocene through Pliocene volcanic rocks. The Gabbs Valley Range and the hills north of Stewart Valley are primarily Oligocene and Miocene silicic ashflow tuffs, Miocene intermediate lavas and volcanic breccias, and lesser amounts of Triassic and Jurassic sedimentary and volcanic rocks and Mesozoic quartz monzonite (Vitaliano and Callaghan, 1963; Ekren and others, 1980, Stewart and others, 1982). The northern Cedar

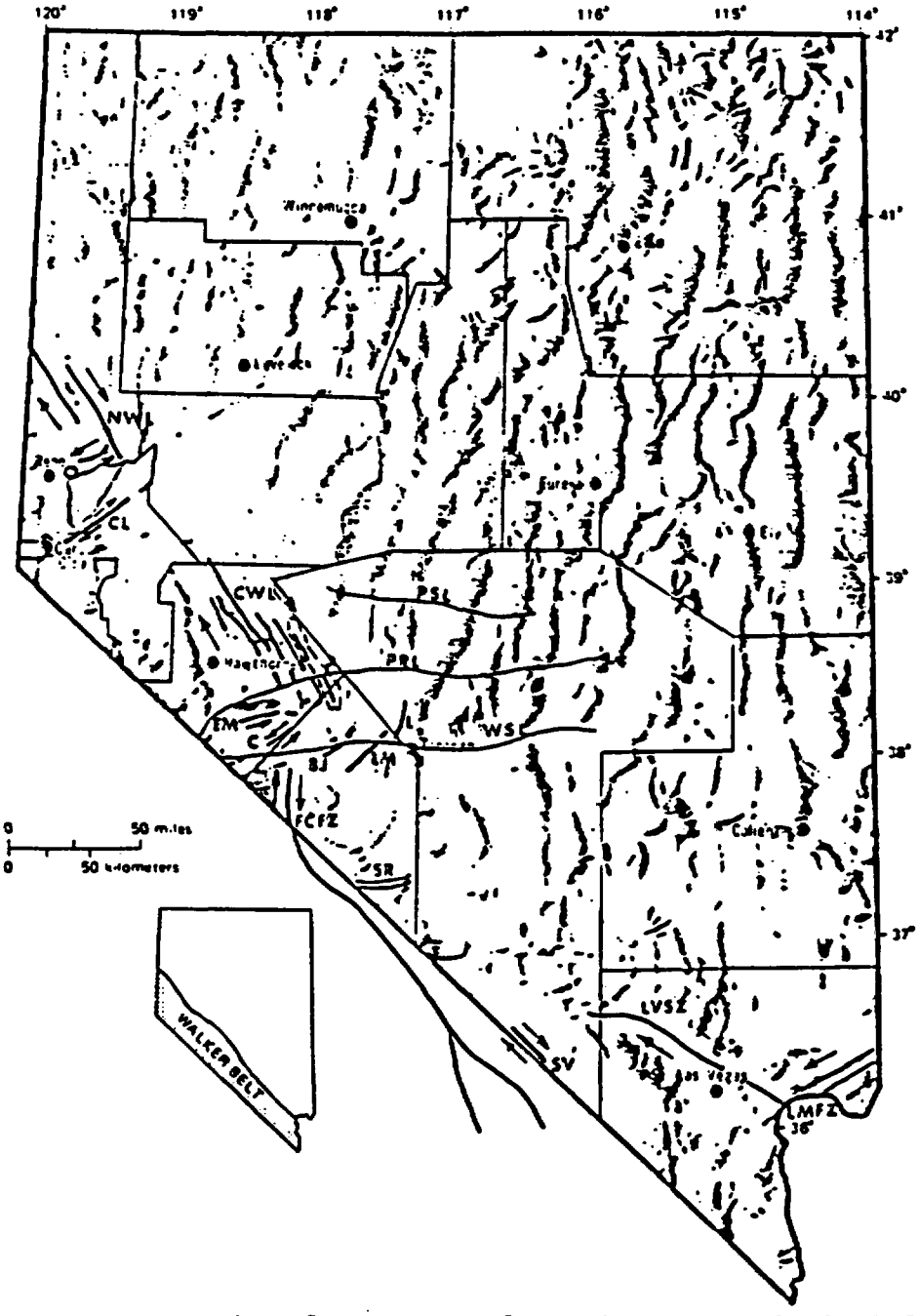

Figure 2 - Map showing location of study area (dashed lines) in respect to major lineaments and late Cenozoic fault zones in the Walker Lane. NWL-north Walker Lane, O-Olinghouse fault, Cl-Carson lineament, CWL-central Walker Lane, EM-Excelsior Mountains fault, C-Candelaria fault zone, BJ-Blair Junction fault zone, LM-Lone Mountain fault, L-Liberty fault, FCFZ-Furnace Creek-Death Valley fault zone, SR-Slate Ridge faults, SV-Stewart Valley fault, LVSZ-Las Vegas shear zone, LMFZ-Lake Mead fault zone, PSL-Pritchard's Station lineament, PRL-Pancake Range lineament, WSL-Warm Springs lineament.

Mountains are predominantly Triassic sedimentary rocks and Cretaceous(?) granitic rocks with Tertiary intermediate lavas and breccias and silicic tuff at the north and south margins (Ross, 1961). The southern Cedar Mountains consist of rhyolitic tuffs and tuff breccias of probable Miocene age, sediments of Esmeralda Formation, and a lesser amount of Triassic and Jurassic sedimentary rocks (Albers and Stewart, 1972). The Pilot Mountains consist of Triassic and Jurassic sedimentary and volcanic rocks and Miocene andesitic lavas, rhyolitic tuff, and sediments around the flanks of the range and as isolated patches within the range (Nielsen, 1964; Oldow, 1978). Rocks in the Monte Cristo Range are primarily rhyolitic tuffs, flows, and intrusive bodies and andesite of Miocene age. Esmeralda Formation, Pliocene(?) basalt, and isolated blocks of pre-Tertiary sedimentary rocks also outcrop within the range (Albers and Stewart, 1972).

Purpose and Scope

Gianella and Callaghan (1934a, b) reported 17.5 kilometers (11 miles) of relatively continuous surface faulting in Monte Cristo Valley and distributed minor surface faulting east, north, and northwest of Stewart Valley associated with the 1932 Cedar Mountain earthquake. The discontinuous fault pattern and relatively small surface displacement reported is anomalous when compared to other historical earthquakes of similar magnitude. Gianella and Callaghan were hampered in their investigation by snow and a lack of aerial photographs, they thus were not able to adequately field check all of the

area for surface faulting.

The objectives of this thesis research were:

1) to map in detail the geology of Stewart and Monte Cristo valleys with an emphasis on late Cenozoic (post-20 m.y.b.p.) structural geology,

2) detect any additional surface faults caused by the 1932 earthquake and obtain geologic and geomorphic data to help explain the anomalous fault pattern,

3) evaluate the style, magnitude, and recurrence of surface faulting during the late Quaternary, and

4) determine the late Cenozoic geologic history of the area and its space-time relation to known Walker Lane and basin-range style structures in adjoining areas.

In addition to mapping approximately 510 square kilometers (200 square miles) in Stewart and Monte Cristo valleys, aerial photographs covering the surrounding area (at least 8 kilometers outside of the mapped area) were studied to detect any additional surface faults associated with the 1932 earthquake and any major late Cenozoic faults not delineated on any previously published maps or reports.

Methodology

Initial work consisted of literature research and review of poor to average quality, 1:16,000 scale black and white Bureau of Land Management aerial photographs on file at the Nevada Bureau of Mines and Geology. In addition, excellent quality 1:80,000 scale black and white aerial photographs, which included coverage of the surrounding mountain ranges,

were purchased from the EROS Data Center in South Dakota. Analysis of these photographs allowed a more regional view of lineaments and other geologic features observed on the 1:16,000 scale photographs. This preliminary phase was completed in the fall of 1981 and spring, 1982 and the thesis scope and geographic boundaries were defined. Low-sun angle black and white aerial photographs of the entire study area were shot at a scale of 1:18,000 during the summer of 1982. Low-sun angle photographs highlight surface faults while still maintaining high resolution and contrast where topographic relief is relatively low. Preliminary interpretation of the low-sun angle photography was completed and field work was initiated during the fall of 1982. Forty-one days of field work were completed between October, 1982 and June 1983.

The 1:18,000 scale aerial photographs were used as a base for field mapping of structures and contacts between Tertiary rock units. Mapping of the Quaternary units was primarily done by aerial photo interpretation but was complimented by field observations, comparison with unpublished soils data (provided by the Soil Conservation Service, Fallon, Nevada), and field checking of questionable and selected critical areas.

Map data was transferred to 1:24,000 scale U.S.G.S. topographic maps (Plates 1, 2 and 4) using a Kail radial Planimetric Plotter. Two 1:250,000 scale color NASA Skylab photographs, on file in the Mackay School of Mines

photogeology lab, were used in conjunction with previously published maps and reports to develop an interpretation of regional late Cenozoic tectonics. Much of the mapping along the western (west of longitude 118°) and southern edges of the area was reconnaissance in nature and slightly modifies previously published and unpublished work (Nielsen, 1964; Ekren and Byers, 1978; Oldow, 1981, 1983 written communication). The reader is referred to these reports for details of the Mesozoic structural geology in the southeastern Gabbs Valley Range and eastern Pilot Mountains.

Previous Work

The first work concerning the geology of the Stewart Valley area was the excellent report by J.P. Buwalda (1914). He describes in good detail the stratigraphy of the mid-to-late Miocene sedimentary rocks located in Stewart and Ione valleys. Structural features, general geology and physiographic history of the valleys and surrounding mountain ranges are also discussed.

Gianella and Callaghan (1934a, b) studied in detail the surface faulting effects associated with the 1932 Cedar Mountain earthquake. Their description of the geology of the area is general in nature and relies heavily upon previously published reports on mining districts within the surrounding mountains.

The first published geologic maps of Stewart and Monte

Cristo valleys are those by Ferguson and Muller (1949) and Ferguson, Muller, and Cathcart (1953). These reports are primarily concerned with the Mesozoic structure and stratigraphy of the surrounding mountain ranges. Mapping of the Cenozoic rocks was reconnaissance in nature and discussions of the stratigraphy and structure are brief and general. Their work was only slightly modified for the geologic maps in Mineral County (Ross, 1961) and Esmeralda County (Albers and Stewart, 1972) geologic reports.

Detailed geologic maps of the Pilot Mountains (Nielsen, 1964; Oldow, 1978) and the Gabbs Valley Range (Ekren and Byers, 1978) have been published and additional work is in progress in the Monte Cristo Range (J. Stewart, 1983, oral communication) and Pilot Mountains (J. Oldow, 1983, written communication).

All additional published reports on the thesis area are concerned with the fossils and, to a lesser extent, the stratigraphy of the mid-to-late Miocene sedimentary rocks in Stewart Valley (Merriam, 1916; Stirton, 1932; Firby, 1964; Mawby, 1965).

Physiography[1]

Stewart and Monte Cristo valleys are two separate drainage basins which form a single structural basin bounded by the Gabbs Valley Range and Pilot Mountains on the west, Monte Cristo Range on the south, and Cedar Mountains on the east. A series of low hills which extend to the southwest from the southern end of the Paradise Range are the northern boundary of the area (Figure 1). The overall trend of the valleys and boundary ranges is northwest-southeast.

Stewart Valley is the northernmost of the two valleys and is an elongate semi-bolson. An axial stream channel, Finger Rock Wash, heads at the drainage divide between the two valleys, extends along the axis of Stewart Valley, and drains to the northwest into Gabbs Valley. The overall drainage pattern is dendritic and the tributary channels have downcut as much as 50 meters (165 feet) into the piedmont slope to form a subdued "badlands" style of topography. The piedmont slope consists of remnants of several erosional surfaces cut into the underlying rock (rock-pediment remnants) and capped with pedisediment, and inset fans within the tributary channels. Elevation of the valley floor ranges from 1890 meters (6200 feet) at the

[1]Throughout the geologic literature various terms have been used for particular landforms and some terms have been given variable and somewhat conflicting definitions. This report uses the landform classification of Peterson (1981) except where otherwise noted. Although some terms may not have been previously used or widely accepted within the geologic literature, it is felt that the consistent order of this comprehensive classification justifies its use.

drainage divide at the south end of the valley to 1524 meters (5000 feet) at the north end of the valley near Finger Rock. Summit peaks in the Gabbs Valley Range and Cedar Mountains rise to elevations of 2518 meters (8310 feet) and 2466 meters (8089 feet) respectively. Although the piedmont angle between the mountain ranges and piedmont slope is sharp, the relief and steepness of the range fronts are considerably less than a typical fault-bounded range front within the Great Basin region of the Basin and Range Province.

Monte Cristo Valley is a bolson with many landforms typical of bolsons throughout the Basin and Range Province although its overall shape is more equidimensional. A playa, Kibby Flat, is located in the central portion of the basin at an elevation of 1605 meters (5268 feet). Axial-stream channels drain the northwest and south-central portions of the basin. The rest of the basin is drained by numerous wash channels that extend to the playa. The piedmont slopes primarily consist of alluvial fans and a dissected fan piedmont whose component landforms include erosional fan remnants and inset fans, and a fan skirt. The fan piedmont and alluvial fans are the dominant landforms in the western, eastern, and northwestern portions of the basin whereas a dissected fan piedmont and fan skirt are dominant in the northern and southern portions. Ballenas occur in the northwestern portion of the basin. There is a small alluvial flat surrounding the playa. Summit peaks reach elevations of 2712 meters (8952 feet) in the Pilot Mountains, 2423 meters (7996 feet) in the Monte Cristo Range,

and exceed 2121 meters (7000 feet) in the southern Cedar Mountains.

In addition to the formal geographic names used within this report, several informal names are also used. These are: Poleline Spring Wash, the drainage located on the eastern flank of the Gabbs Valley Range in which Poleline Spring occurs and whose mouth is 4.1 kilometers (2.6 miles) due west of Simon Well; Stewart Valley Road, the main road that parallels Finger Rock Wash in Stewart Valley; Mina Road, the road that enters Monte Cristo Valley from the west and merges with the Stewart Valley Road; Monte Cristo Valley Road, the road located on the west side of Monte Cristo Valley that extends from the intersection of the Mina and Stewart Valley Roads to Dunham Mill site in the south; "Y"-intersection, intersection of the Monte Cristo Valley Road, Stewart Valley Road, and Mina Road in northwest Monte Cristo Valley.

MESOZOIC STRATIGRAPHY

Sedimentary Rocks

Permian and Triassic sedimentary rocks outcrop along the western margin of the study area in the southeastern portion of the Gabbs Valley Range and eastern Pilot Mountains. Descriptions of these units are based solely on previously published reports.

Mina Formation (Pm): The late Permian Mina Formation outcrops throughout most of the southern portion of the Pilot Mountains (Plate 2). The rocks were previously considered to be part of the Excelsior Formation (Ferguson and others, 1953) but were redefined by Speed (1977). Fine and coarse-grained volcanogenic turbidite with lesser amounts of chert, pelitic mudstone, and quartz sandstone comprise the formation. Speed (1977) divided the formation into three members: a coarse turbidite member up to 500 m (1640 feet) thick, a pelitic member greater than 70 m (230 feet) thick, and chert-clastic member up to 500 m (1640 feet) thick.

Luning Formation (TR l): The upper Triassic Luning Formation outcrops in the southern Gabbs Valley Range and northern Pilot Mountains (Plates 1 and 2). The formation has three members and is 2500 m (8365 feet) thick at the type section in the northeastern Pilot Mountains (Oldow, 1981). The lower member is 900 m (2950 feet) of clastic rocks and medium to fine-grained crystalline limestone and calcarenite with a relative abundance of 60% and

40%, respectively. The clastic rocks are calcareous and non-calcareous mudstone, sandy mudstone, and minor amounts of fine-grained quartz wacke. The middle member is 900 m (2950 feet) of unfossiliferous, interbedded conglomerate, sandstone, sandy mudstone, and limestone. The upper member is 750 m (2460 feet) of interbedded fine to medium-grained limestone and dolomite and sandy mudstone with a relative abundance of 70-80% and 20-30%, respectively. In the southern Gabbs Valley Range only the upper member (Ꞧ lu) is present (Ekren and Byers, 1978). Here the rocks are massive-weathering, thick bedded dolomite and dolomitic limestone and thick to very thick bedded, medium to coarse-grained limestone. Structural complexities prevent a definite determination of the thickness but it is a minimum of 800 m (2625 feet) (Ekren and Byers, 1978).

Plutonic Rocks

A small plug of porphyritic granodiorite (Kgr) is intruded into the Luning Formation on the east flank of the Pilot Mountains (Plate 2). Granitic rocks on the west flank of the Pilot Mountains are Cretaceous in age (Stewart and others, 1982) and this plug is probably of a similar age. A modal analysis of a thin-section showed a phenocryst composition of 60% plagioclase, 24% k-feldspar, 11% quartz, 4% hornblende, and 1% biotite (Ross, 1961).

TERTIARY STRATIGRAPHY

Volcanic Rocks

Volcanic rocks ranging in age from late Oligocene to Pliocene(?) occur on the flanks of all the ranges surrounding the Stewart and Monte Cristo valleys. All of the age dates[2] and most of the lithologic descriptions are from previously published reports.

Lavas of Giroux Valley (Tlg): The lavas of Giroux Valley which outcrop in the southeast Gabbs Valley Range (Plate 1) are dark gray olivine-pyroxene basalt. Phenocrysts 2 mm and less in size constitute 10-40% and the groundmass texture varies from basaltic to pilotaxitic. The phenocrysts are plagioclase, clinopyroxene, and olivine (Ekren and Byers, 1978). Lavas and a tuff-breccia with a more silicic composition that outcrop in the southern Gillis Range are considered to be correlative with the lavas of Giroux Valley and were K-Ar dated at 26.7±0.9 m.y. on biotite and 25.7±0.6 m.y. on sanidine (Ekren and others, 1980).

Guild Mine Member of Mickey Pass Tuff (Tbmg): The Guild Mine Member of the Mickey Pass Tuff overlies lavas of Giroux Valley in a sliver between two strands of the Bettles Well fault zone in the Gabbs Valley Range (Plate 1). The

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[2]K-Ar age dates with an asterisk(*) are the recalculated values from Moore (1981) who used updated decay constants of Steiger and Jaeger (1977). Age dates from Morton and others (1977) were originally calculated with the updated decay constants.

Guild Mine Member is a compound cooling unit of pumice-rich densely welded tuff with a composition grading from rhyodacite at the base to rhyolite at the top. Phenocryst volume ranges between 25-40% throughout the unit and lithic fragments are common in the basal portion. K-Ar dates of between 26.3±0.9 m.y. and 28.0±1.0 m.y. were obtained from biotite and sanidine in samples from the Gabbs Valley and Gillis Ranges (Ekren and Byers, 1978; Ekren and others, 1980) and Singatse Range (Proffett and Proffett, 1976).

Singatse Tuff (Tbs): The Singatse Tuff occurs as a thin outcrop underlain by lavas of Giroux Valley and overlain by a detached block of lavas of Mount Ferguson near the Bettles Well fault in the southeast Gabbs Valley Range (Plate 1). The Singatse Tuff is densely welded, rich in lithic fragments, and is rhyodacite in the basal portion and quartz latite throughout the rest of the unit. The compositional change occurs abruptly at a partial or complete cooling break (Ekren and Byers, 1978; Ekren and others, 1980). K-Ar age dates of 27.4±0.9 m.y. and 27.2±1.1 m.y. were obtained on biotite in samples from the north Pilot Mountains and the Singatse Range respectively (Marvin and others, 1977; Proffett and Proffett, 1976).

Tuff of Southern Pilot Mountains (Tvt): This bluish-gray densely welded tuff and tuff breccia outcrops in the low hills on the southeast edge of the Pilot Mountains (Plate 2). Phenocrysts are mainly hornblende and plagioclase with lesser amounts of pyroxene. The phenocrysts are abundant and are

generally less than 2 mm long. This tuff is most likely related to the tuffs in the Candeleria region (Speed and Cogbill, 1979b) to the west or may be a distal portion of the Singatse Tuff. The tuff does not appear to be related to the Castle Peak Tuff (Moore, 1981) which outcrops to the south in the Monte Cristo Range since it is dissimilar to the lithologic description of the Castle Peak Tuff and has a distinctly different morphology and color when viewed on aerial photographs.

Lava of Intermediate Composition (Tlr): Occurs as two small masses intruded into the lavas of Giroux Valley near the Bettles Well fault zone in the southeast Gabbs Valley Range (Plate 1). Ekren and Byers (1978) did not give a lithologic description and the outcrops were only verified by aerial photo interpretation in this study.

Andesite Lavas (Ta): Andesitic lavas outcrop on the east flank of the Pilot Mountains and northwest flank of the Cedar Mountains (Plates 1 and 2). The lavas in the northeast and east Pilot Mountains primarily consist of a basal hypersthene augite andesite breccia that interfingers with and is overlain by hornblende andesite breccia. Locally olivine basalt and thin units of welded dacite and rhyodacite tuff and tuff breccia occur within the andesite breccias and a few dikes and plugs of hornblende dacite and hornblende andesite intrude the breccias. Several of these intrusives occur along Tertiary faults (Nielsen, 1964). None of these units were differentiated for the purposes of this study.

A hornblende andesite intrusive in the low hills east of the range near the Good Hope Mine was K-Ar dated on hornblende at 24.7±0.7 m.y. (Morton and others, 1977).

The andesitic lavas that outcrop on the southeast edge of the Pilot Mountains are most likely related to the Gilbert Andesite which occurs throughout the Monte Cristo Range (Albers and Stewart, 1972). K-Ar dates on biotite yielded ages of 15.5±0.6* m.y. and 15.5±0.5* m.y. (Albers and Stewart, 1972; Silberman and others, 1975) for the Gilbert Andesite.

The black or dark, reddish brown hornblende bearing andesitic lavas that outcrop on the northwest flank of the Cedar Mountains are in fault contact with or unconformably overlain by the Miocene Esmeralda Formation. Approximately 2 km (1.25 miles) northwest of Stewart Spring there is an angular unconformity of 13° between the flow foliation in a lone outcrop of the lavas and the overlying diatomites and diatomaceous shales of the Esmeralda Formation which are the oldest Esmeralda sediments in northern Stewart Valley (Mawby, 1965, p. 16). These sediments were considered to be stratigraphically similar to the 15-15.5 m.y. old sediments on the west side of Stewart Valley thus these andesitic lavas are most likely older than 15.5 m.y.

<u>Lavas of Mount Ferguson Undivided</u> (<u>Tlfu</u>): The lavas of Mount Ferguson consist of numerous lava flows and flow-breccias of intermediate composition that form almost the entire flank of the Gabbs Valley Range in the study area (Plate 1). The source area for these rocks is centered in the Mount Ferguson

and Gabbs Mountain areas in the Gabbs Valley Range and southern Gabbs Valley, respectively. The lavas range in composition from porphyritic trachyandesite to porphyritic quartz latite. K-Ar age dates obtained from various flows range from 15.0±0.5 m.y. (plagioclase) to 22.5±0.6 m.y. (hornblende) (Morton and others, 1977; Ekren and Byers, 1978; Ekren and others, 1980).

The dominant rock type in the study area is a thick sequence of volcanic debris flow breccia (Tlfb) locally overlain by lavas (Tlf). The breccias consist of pebbles, cobbles, and boulders suspended in a tuffaceous matrix (Figure 3). The breccias have a relatively uniform appearance throughout the sequence although the size and concentration of coarse material varies in individual flows. There are three major lithotypes of clasts in the well-exposed section along Poleline Spring Wash. The predominate clast closely resembles the matrix and is a light gray, phenocryst-rich, prophyritic lava. The main phenocrysts, plagioclase and hornblende, are up to 4 mm long and biotite is a lesser constituent. A second lithotype is a dark gray, glassy, flow-banded porphyritic lava with phenocrysts of plagioclase, hornblende and biotite. The number and average size of the phenocrysts is less than the previously described clasts. These two lithotypes are similar to the hornblende quartz latite and silicic quartz latite described by Ekren and Byers (1978) that outcrop on Gabbs Mountain northwest of the study area. The other major lithotype is a reddish-brown to purplish-brown porphyritic lava with plagioclase phenocrysts up to 3 mm long and

Figure 3 - Outcrop of volcanic debris flow breccias of the lavas of Mount Ferguson on the east flank of the Gabbs Valley Range. Note hammer below center boulder for scale.

a trace of hornblende. These lithotypes are generally present throughout the breccias but there is some variability in clast lithology to the north. Locally underlying the breccias in Poleline Spring Canyon are several meters of buff white volcanic breccia. This breccia consists of 20-30% clasts, generally less than 10 cm in diameter, in a tuffaceous matrix. There are a variety of clasts all of which are volcanic and the dominant lithotype is a white non-welded tuff with biotite, sanidine, and quartz phenocrysts. South of Poleline Spring Wash, K-Ar dates of 15.1±0.5 m.y. on plagioclase and 15.4±0.5 m.y. on hornblendes were obtained for the breccias (Morton and others, 1977).

Volcanic Breccias of Cedar Mountains (Tcmb): Volcanic debris flow breccias of at least 50 meters (165 feet) thick outcrop in the hills just north of Stewart Spring (Plate 1). The matrix and dominant clast lithotype are almost identical in appearance and mineralogy to the Mount Ferguson breccias (Tlfb). The Cedar Mountains breccias are slightly more crystal rich with biotite being almost as abundant as hornblende. Other clast lithotypes represent older volcanic rocks outcropping in the Cedar Mountains to the east. These breccias overlie the oldest Esmeralda sediments and are overlain by younger Esmeralda sediments near Stewart Spring and thus are roughly equivalent in age to the Mount Ferguson breccias.

Older Volcanic Instrusive(?) (Til): Several small plugs of brownish purple lava with plagioclase and lesser amounts of hornblende phenocrysts protrude through the Esmeralda

Formation in west-central Stewart Valley (Plate 1). Whether these rocks intruded the Esmeralda Formation or are paleotopographic highs is somewhat amgibuous. These lavas have a similar appearance and phenocryst mineralogy to rocks of the lavas of Mount Ferguson and may predate the Esmeralda Formation. Approximately 3.2 km (2 miles) west of Stewart Spring two low hills are formed by what appears to be opalized and silicified Esmeralda Formation. This alteration has completely destroyed all the primary bedding of the sediments. A small plug of the lava outcrops near the southern base of the easternmost hill suggesting that the altered sediments are a result of the intrusion of lava. It is possible that emplacement of these volcanic rocks both predate and postdate deposition of Barstovian age Esmeralda sedimentary rocks.

Younger Volcanic Intrusive (Ti2): Light to dark gray lava that weathers to dark or rust brown intrudes the Mount Ferguson breccias (Tlfb) and Esmeralda Formation on the east flank of the Gabbs Valley Range (Plate 1). The lavas contain small phenocrysts (2 mm or less) of hornblende and minor plagioclase. Phenocryst size decreases at the margins and the lavas are locally columnar jointed. Several small domes intrude the Esmeralda Formation south of Poleline Spring Wash and there are probably additional small intrusives in the breccias but the similar color and phenocryst mineralogy of the breccia matrix and intrusives made it difficult to distinguish between the two units.

Latite Flows and Intrusives (Tl): Dark gray to black late Miocene latite outcrops as flows on Table Mountain in the southern Gabbs Valley Range and intrusive volcanic necks and flows in northern Stewart Valley (Plates 1 and 2). The latites are porphyritic and aphyric with hornblende microphenocrysts or phenocrysts ± clinopyroxene or aphyric with clinopyroxene as the only microphenocryst (Ekren and Byers, 1978). Nielsen (1964) noted a volcanic neck at the head of Dunlap Canyon in the Pilot Mountains that is the probable source for the Table Mountain latite flows. The volcanic necks near Finger Rock intrude the Esmeralda Formation.

Basalt (Tb): Dark gray and black aphyric basaltic lava with microphenocrysts of olivine and clinopyroxene (Ekren and Byers, 1978) conformably(?) overlies the Esmeralda sediments in southwest Stewart Valley and occurs in a detached block overlying Esmeralda sediments along the Bettles Well fault (Plate 1). Ekren and Byers (1978) considered this basalt to be late Miocene. Remnants of basalt flows also outcrop in southern Monte Cristo Valley (Plate 2). The nearest source area for these basalts is to the west in the Candelaria Hills (Ferguson and others, 1953) where K-Ar dates on whole rock samples range from 2.9±0.1* m.y. to 4.0±0.4 m.y.* (Silberman and others, 1975; Marvin and others, 1977), suggesting a Pliocene age for the basalts in Monte Cristo Valley.

Undifferentiated Tertiary Volcanic Rocks (Tv): Tertiary volcanic rocks that were not mapped in detail occur in several locations around the perimeter of the mapped area. In the northern Cedar Mountains these rocks include late Oligocene and early to mid-Miocene silicic ash-flow tuffs and andesitic lavas and breccias. Knopf (1921) described the Tertiary volcanic rocks at the Simon and Omco Mines which are located just east of the mapped area. These units most likely make up the bulk of the Tertiary volcanic rocks in the northern Cedar Mountains. The Simon Mine is in the central portion of the northern Cedar Mountains, just east of Wildrose Spring (see Plate 1). The Tertiary section from oldest to youngest, is porphyritic hornblende andesite, keratophyre (divitrified and vapor-phase altered alkali trachyte), pyroxene andesite, vitric dacite tuff, and quartz latite tuff. The Omco (Olympic) Mine is located on the north flank of the Cedar Mountains where the pre-Esmeralda, Tertiary section is two white rhyolite tuffs separated by a quartz trachyte (may be altered quartz latite) and air-fall(?) tuff. The basal rhyolite is highly altered and the younger rhyolite is rich in lithic fragments. Ekren and others (1980) reported the presence of the Guild Mine Member of the Mickey Pass Tuff and the Singatse Tuff in the northern Cedar Mountains. These tuffs probably correspond with the older and younger rhyolite tuffs, respectively. The middle unit (quartz tracyte) may be the Weed Heights Member of the Mickey Pass Tuff or a basal cooling unit of Singatse Tuff. Additional units present in

the northern Cedar Mountains are the porphyritic andesitic volcanic breccia (Tcmb) and andesitic lavas.

In the southeastern Gabbs Valley Range these rocks are predominantly intermediate composition lavas and volcanic breccias of the lavas of Mount Ferguson but may locally include older rocks such as the lavas of Giroux Valley, Guild Mine Member of the Mickey Pass Tuff, and Singatse Tuff. Locally the Luning Formation (TR lu) may be exposed underlying the volcanic rocks along the Bettles Well fault.

The low hills north of Stewart Valley are underlain by lavas of Mount Ferguson but locally older ash-flow tuff and tuff breccia is exposed. These tuffs are most likely the same as those which outcrop near the Omco Mine.

Rhyolite ash-flow tuff and tuff breccias, and Tertiary basalts outcrop in the northwestern Monte Cristo Range in the southwestern corner of the study area. This area is presently being mapped for a U.S. Geological Survey project (J. Stewart, oral communication, 1983).

Sedimentary Rocks

Esmeralda Formation (Te): Mid-to-late Miocene sedimentary rocks of lacustrine, fluvial, and alluvial orgin completely underlie Stewart Valley and northern Monte Cristo Valley. To the south they can be mapped almost continuously to the northwest edge of Kibby Flat in the upthrown block along a fault. In addition, similar sedimentary rocks outcrop fairly extensively to the southeas of the playa and in an isolated patch along the southwest edge of the mapped area

(Plates 1 and 2).

All previous reports have assigned these rocks to the Esmeralda Formation as originally defined by Turner (1900) in central Esmeralda County. Turner envisioned a large lake covering much of west-central Nevada in which fresh water sedimentary rocks were deposited during the Miocene and early Pliocene. Subsequent work (Robinson and others, 1968; Albers and Stewart, 1972) has shown that many of these rocks in Esmeralda County are of differing ages and have been deposited in basins separate from that of the type section in the Weepah Hills, Esmeralda County (Robinson and others, 1968); thus, the name Esmeralda Formation should only be used for sedimentary rocks directly related to the type section.

The sedimentary rocks in Stewart and Ione Valleys are in part coeval and the two basins were connected around the north end of Cedar Mountain (Buwalda, 1914; Mawby, 1965). Buwalda (1914, p. 343) traced beds from Ione Valley southward, "through a number of exposures along the east side of Cedar Mountain, to the Monte Cristo Range" and suggested a connection to the type section of Turner (1900). Albers and Stewart (1972, p. 38) noted a lithologic similarity and suggested a possible connection between the sections in the Monte Cristo Range and Coaldale area and the type section. This interpretation is supported by Moore (1981). Since the sedimentary rocks outcropping north of Kibby Flat are less than 8 kilometers (5 miles) from those of southern Monte Cristo Valley

and the Monte Cristo Range it is probable there was original depositional continuity between these areas. Due to the probable continuity with the type section, concordant age dates (Table 1), and similar fossil assemblages (Merriam, 1916; Stirton, 1932, Firby, 1964, Mawby, 1965; Moore, 1981), the name Esmeralda Formation will be used for the Miocene sedimentary rocks mapped in this study.

The Esmeralda Formation in Stewart and Monte Cristo valleys consists of sandstones, shales, siltstones, claystone, limestone, chert, tufa, diatomite, conglomerate and air-fall tuff. Mawby (1965) described four broad units within the formation in northern Stewart Valley and indicated a total thickness of at least 445 meters (1470 feet). Due to a lack of distinctive marker beds within the units, facies changes, and structural complexities, no attempt was made to map these units throughout the area. Based on observations made during this study, the thickness determined by Mawby should be considered a minimum.

Fine to coarse-grained sandstones of primarily lacustral origin are the dominant lithology within the Esmeralda Formation and are commonly rich in tuffaceous material. Coarser units, many of which are crossbedded, are more abundant along the eastern edge of Stewart Valley and in the upper portion of the section. These units are suggestive of alluvial and fluvial-deltaic deposition at or near the lake margin.

Siltstone, claystone, and paper shales are predominate in the central portion of Stewart Valley with minor interbeds

Table 1

K-Ar Age Dates for the Esmeralda Formation in Mineral and Esmeralda Counties, Nevada

Rock Unit (General Locality)	Reference and Material Dated	Age in m.y.
1. Esmeralda Formation (Cedar Mountain area)	Evernden and others (1964), biotite	10.9[a]
2. Esmeralda Formation, tuff (Fish Lake Valley)	Evernden and others (1964), biotite	11.4[a]
3. Esmeralda Formation, tuff (North Fish Lake Valley)	Evernden and others (1964), biotite	11.7[a]
4. Esmeralda Formation, tuff (Cedar Mountains)	Evernden and others (1964), sanidine	11.8[a]
5. Esmeralda Formation, tuff interbedded with coal, type locality (North base Silver Peak Range)	Evernden and James (1964), biotite	13.0[a]
6. Esmeralda Formation, tuff (Fish Lake Valley)	Robinson and others (1968), biotite	13.4[a]
7. Esmeralda Formation (Gabbs Valley)	Ekren and Byers (1978), Apache tears	15.2±0.4
8. Intermediate debris-flow breccia interfingering with Esmeralda Formation (Stewart Valley)	Morton and others (1977), plagioclase hornblende	15.1±0.5 15.4±0.5

[a]Age dates recalculated by Moore (1981) with updated decay constants from Steiger and Jaeger (1977).

of diatomite and resistant, yellowish-brown, dirty limestone. These sediments are mostly calcareous but resistant beds of siliceous shale and chert are not uncommon. They generally are white, buff, or light grayish brown in color, are poorly resistant, and are best exposed in channel cuts or steep sideslopes (Figure 4).

Several beds of air-fall tuff occur at various horizons in Stewart Valley. These beds are usually thickly laminated to thinly bedded and contain sedimentary structures such as ripple marks, crossbedding, slump folds, and water escape features which indicate the tuffs were waterlain or reworked. The tuffs range in composition from fine-grained pumice and crystal tuff to volcanic ash. Two distinct tuff beds in the upper portion of the formation outcrop fairly extensively in north-central Stewart Valley and have been correlated by Mawby (1965) with tuffs in Ione Valley dated by Evernden and others (1964). The lowermost of these tuffs is a bluish-gray ash K-Ar dated on biotite at 10.9 m.y.*. The upper tuff is approximately 24 meters (80 feet) up-section and is a white, pumice rich, biotite tuff K-Ar dated on sanidine at 11.8 m.y.*. Similar tuffs outcrop in southeastern Stewart Valley but a petrographic and geochemical analysis are needed for a definite correlation. These tuffs were mapped as marker beds (Plate 1) since they provide the only consistent markers for structural and age relations.

On the west side of Stewart Valley, south of Poleline Spring Wash, lacustrine and fluvial sediments of the Esmeralda

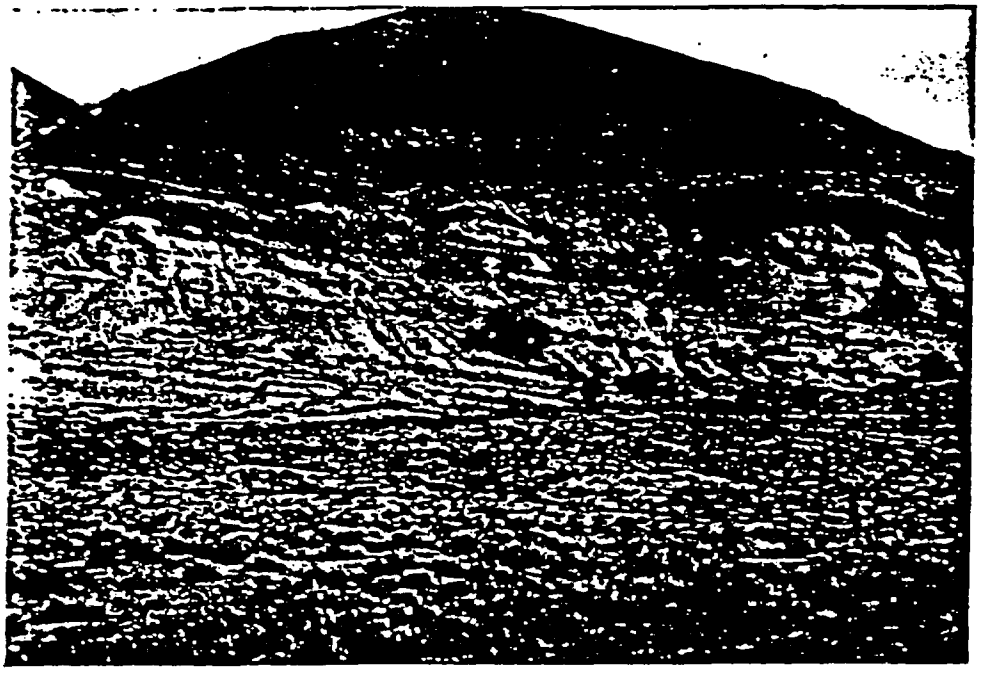

Figure 4 - Gently dipping diatomaceous shales of the Esmeralda Formation exposed in sideslope of a low hill in eastern Stewart Valley. Tertiary volcanic rocks form hill on west flank of Cedar Mountains in the background.

Formation underlie, abut against in a depositional contact, and overlie the 15.1-15.4±0.5 m.y.* volcanic breccias (Tlfb).

These age dates bracket the Esmeralda Formation in Stewart Valley and northern Monte Cristo Valley between approximately 15.5 m.y. and 10.5 m.y. (mid-to-late Miocene) which is consistent with mammalian and molluscan fossil assemblages which indicate an early Barstovian to early Clarendonian age for the sedimentary rocks (Firby, 1964; Mawby, 1965; Merriam, 1916, Stirton, 1932). Similar sediments in southern Gabbs Valley yielded a K-Ar date of 15.2±0.4 m.y. on "Apache tears" collected from basal gravel beds (Ekren and Byers, 1978). Late Clarendonian or early Hemphillian age fossils were collected in Esmeralda-like sediments outcropping between the Pilot Mountains and Gabbs Valley Range near Bettles Well (Nielsen, 1964) and a 7.5±0.2 m.y.* age date was obtained on hornblende in a tuffaceous interbed (Marvin and others, 1977). No direct connection between these sediments and the Esmeralda Formation in Monte Cristo Valley was detected. These younger sediments were apparently deposited in a separate, isolated basin and should not be included in the Esmeralda Formation or if sediments of this age were deposited in Stewart and Monte Cristo valleys, they have since been eroded away or buried by Quaternary alluvium.

<u>Alluvium and Colluvium (Tac)</u>: Alluvial fan remnants of pebble to boulder size subrounded clasts in a tuffaceous matrix that outcrop in a wedge-shaped block between two splays of the Bettles Well fault in the southeast Gabbs

QUATERNARY STRATIGRAPHY

Quaternary Stratigraphic Nomenclature

The Quaternary stratigraphic nomenclature is taken from Dohrenwend (1982) and landform nomenclature follows that of Peterson (1981), unless noted otherwise. The initial mapping of the Quaternary units was by photogeologic interpretation of 1:18,000 black and white low-sun-angle aerial photographs. The units were identified by mapping the geomorphic surface formed on the alluvium and pedisediment. Peterson (1981, p. 42-43) defines a geomorphic surface as "...a part or several parts of the land surface that has been formed during a particular time period." It "...may comprise a single landform, several landforms, or parts of yet others," and the material the surface is formed on must be of sufficient thickness so that it contains, or could contain, a pedogenic soil. The geomorphic surface may or may not have formed coincident with the landform on which it occurs and the age of the soil formed on the surface is a minimum age for the underlying material. In general, the higher the stable geomorphic surface relative to surrounding surfaces, the older the surface is. Exceptions are surfaces on fan collars and fan aprons (Peterson, 1981), segmented alluvial fans that result from high uplift rates on range front faults (Bull, 1964), and eolian deposits. High rates of late Quaternary faulting can also cause a younger surface to be stratigraphically higher than an older surface. A stable geomorphic

surface is "...one where there is a pedogenic soil in the surficial material that forms the surface" (Peterson, 1981, p. 44). Stable surfaces form when climatic changes or tectonics induce changes in an erosional-depositional system which cause a variation in the loci and mode of deposition. Conversely, an active surface is one where "...recent or ongoing erosion or deposition...has destroyed or buried a pre-existing pedogenic soil" (Peterson, 1981, p. 44).

Correlation and relative age of the surfaces was based on the stratigraphic position of the surface, accordancy of summits, on-surface drainage pattern, and amount of relief between the surface summit and the floor of on-surface drainage channels. In general, an older surface has more developed and fewer on-surface drainages than a younger surface. An unpublished soil survey of the area provided by the Soil Conservation Service, Fallon, Nevada, was used to help correlate surfaces between different portions of the two basins. The approximate age ranges for soils on the surfaces (Table 2) are based on ages determined for similar soils in the Reno-Carson City area of northwest Nevada. These ages are considered to be minimum ages since the study area is more arid than the Reno-Carson City and was not occupied by pluvial lakes during the late Pleistocene. If these inferred ages are accurate for the study area, then relatively high rates of downcutting in Stewart Valley and uplift of the southeastern Pilot Mountains are suggested by the young ages of the pediments and alluvial fans in these areas, respectively.

Table 2

Reno-Carson City Area, Nevada (Bell, 1981)			Stewart and Monte Cristo Valleys, Nevada (this study)		
Soil Name	Soil Order or Great Group	Approximate Age	Quaternary Stratigraphic Units	Soil Great Group	Inferred Age
Little or no soil horizon	Entisol (A-C)	<2,000 - 3,000	Qf3c Qf3b	Torriorthent	late Holocene
			Qf3a	Torriorthent	mid-to-late Holocene
Toyeh	Camborthid Haploxeroll	5,000 - 12,000	Qf2b,Qp2b	Camborthid Calciorthid	early Holocene
Churchill	Haplargid	35,000	Qf2b	Haplargid	late Pleistocene
Cocoon	Haplargid Durargid Paleargid Durixeroll	100,000	Qf2a Qp2a	Durargid Haplargid Natrargid	late Pleistocene
Humboldt Valley	Durargid Paleargid Durixeroll	200,000	Qf1b Qp1b Qp1a	Duarargid Natrargid Natrargid	mid-to-late Pleistocene

The Quaternary units in Stewart and Monte Cristo valleys are primarily classified as alluvial fan (Qf) or pediment (Qp) deposits. Additional depositional units include playa (Qpa) and sand dunes (Qsd). The number subscript (1, 2, 3) classifies the unit as old, intermediate and young deposits, respectively. The letter subscript (a, b, c) indicates sub-units with "a" being the oldest and "c", the youngest. Fan and pediment deposits with the same subscripts (i.e., Qf2a and Qp2a) are considered to have roughly equivalent ages.

Alluvial fan deposits form landforms with a constructional geomorphic surface. Landforms comprised of fan deposits include erosional and non-buried fan remnants, fan aprons, inset fans, fanlettes, fan skirts, stream terraces, alluvial flats, and active drainage channels. Some of the old and intermediate-age erosional fan remnants have broadly rounded crests with little or no remaining constructional surface. These landforms are termed ballenas after Peterson (1981). Figure 5 shows stratigraphic relationship betweeen several different fan deposits.

Pediment deposits (pedisediment) are relatively thin veneers of alluvium overlying an erosion surface cut into the underlying rock or older alluvium (Figure 6). The geomorphic surface formed by the pedisediment is somewhat younger than the erosion surface cut into the underlying material. Pediments are most commonly cut into the easily eroded sediments of the Esmeralda Formation within Stewart

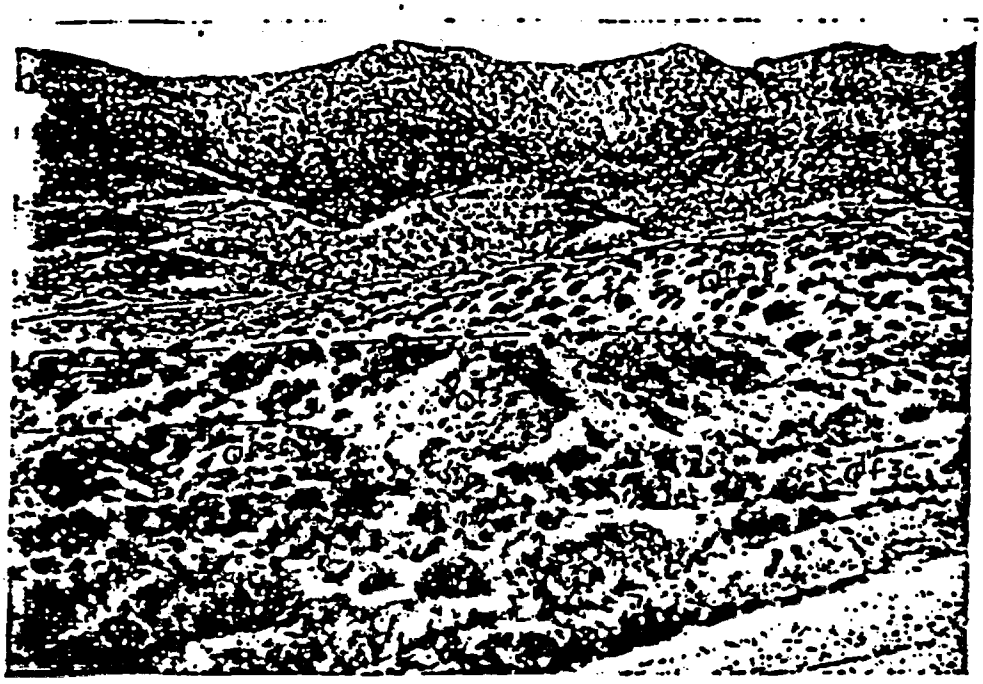

Figure 5 - Photo showing the relative position of several fan surfaces in Monte Cristo Valley near the Good Hope Mine. See text for unit symbol descriptions. View to the west, Pilot Mountains in the background.

Figure 6 - Late Pleistocene pedisediment overlying gently dipping shales and sands of the Esmeralda Formation in northern Stewart Valley.

Valley and in northern and southern Monte Cristo Valley. Pediments of this type are also referred to as pediment-terraces (Royse and Barsch, 1981).

Post-pediment dissection has been great enough that these landforms are actually pediment remnants, if flat summits are preserved, or ridgeline remnants with rounded crests. Ridgeline remnants usually have less developed (younger) soils than summits of pediment remnants. Many of the small map delineations identified as pediment deposits may be ballenas or erosional fan remnants but due to their isolated locations and small area, it is difficult to determine if the surfaces are constructional in origin.

Locally in Stewart Valley erosional surfaces cut across the Esmeralda Formation which have little or no overlying pedisediment and consequently, were not mapped. Some of these surfaces are correlative with surfaces underlying mapped pediment deposits but others are intermediate surfaces. Figure 7 shows several pediment remnants in Stewart Valley. Pediment deposits in Monte Cristo Valley are most commonly preserved as caps on ridgeline remnants.

Alluvial Fan Deposits (Qf)

The Quaternary fan deposits are composed of poorly sorted deposits of boulder, cobbles, pebbles, sand, and silt deposited by braided stream channels or debris-flows. Clast lithologies are primarily Tertiary volcanics but also include Mesozoic limestones and quartzites where these rocks outcrop

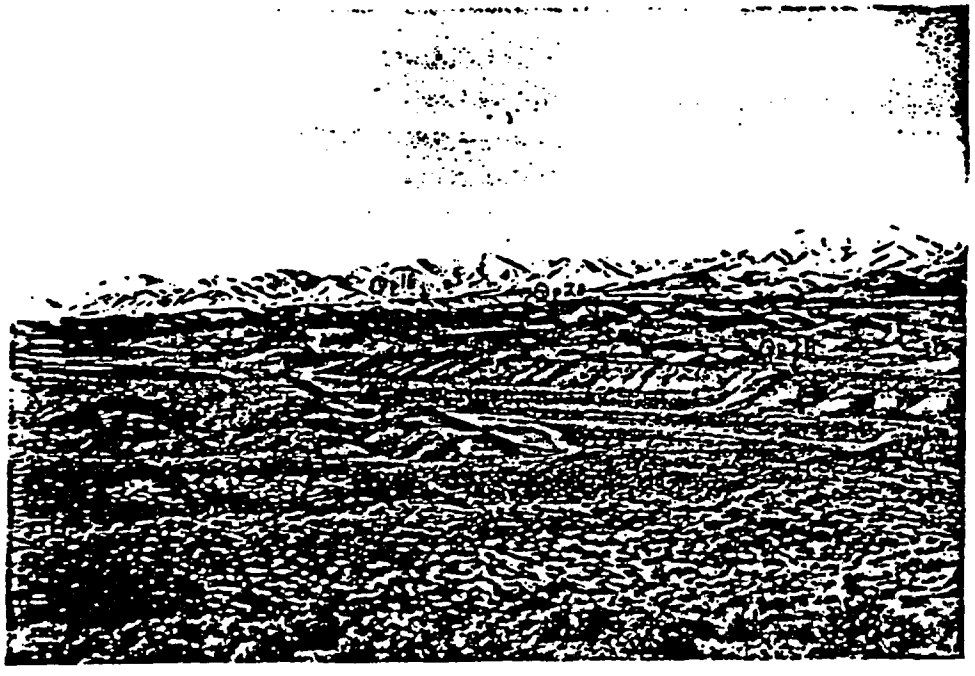

Figure 7 - Photo showing pediment remnants in Stewart Valley. View to the west looking towards snow covered Gabbs Valley Range. See text for description of symbols.

in the source area. Sorting tends to increase and clast size decreases with increasing distance from the source area.

Old Alluvial Fan Deposits (Qflb): Older alluvial fan deposits occur as erosional fan remnants and ballenas in northwest Monte Cristo Valley. Relief formed by subparallel on-surface drainages is up to 9 m (30 feet). Contacts between Qflb deposits and younger deposits occur along steep erosional sideslopes or may be gradational at distal margins. Qflb surface soils are Durargids that are probably as old as mid-Pleistocene but may be as young as the late Pleistocene Humboldt Valley Soil (Table 2).

Intermediate Alluvial Fan Deposits (Qf2a, Qf2b): Intermediate fan deposits occur as erosional fan remnants and nonburied fan remnants in southernmost and northernmost Stewart Valley and throughout Monte Cristo Valley. Qf2a surfaces are moderately dissected by subparallel on-surface drainages with up to 6 m (20 feet) of relief. Contacts between intermediate fan deposits and younger deposits may be a steep erosional sideslope or gradational. Qf2a surface soils are Durargids and Haplargids that are probably equivalent in age to the late Pleistocene Cocoon Soil (Table 2). Qf2b surfaces are moderately to slightly dissected by distributary channels and some minor on-surface drainage occurs. Relief due to dissection is greatest on the east flank of the Pilot Mountains south of the Good Hope Mine. Contacts with young fan deposits may be an erosional sideslope but more commonly is gradational. Qp2b surface soils are Haplargids,

Calciorthids, and Camborthids and are approximately age equivalent to the late Pleistocene Churchill Soil and latest Pleistocene-early Holocene Toyeh Soil (Table 2).

Young Alluvial Fan Deposits (Qf3a, Qf3b, Qf3c): Young alluvial fan deposits occur throughout the study area as erosional fan remnants (Qf3a); inset fans, alluvial flats, fanlettes, stream terraces, and erosional fan remnants (Qf3b); and active drainage channels and fan aprons (Qf3c). The Qf3a and Qf3b surfaces are slightly dissected to undissected with a maximum relief between these surfaces and Qf3c channels of about 2 m (7 feet). Drainages on Qf3a surfaces are generally distributary, radiating from the fan apex while those on the Qf3b surfaces are braided. Qf3b and Qf3c surfaces have little or no pedogenic soil formed and are latest Holocene to Recent (<3,000 years). Qf3a surfaces generally have a thin, slightly vesicular A horizon and a weakly differentiated Bk horizon. These surfaces are considered to be mid-to-late Holocene in age.

Pediment Deposits (Qp)

Quaternary pediment deposits are usually less than 15 m (50 feet) thick and are texturally, lithologically, and morphologically similar to age equivalent fan deposits. Many of these may be fan deposits but erosion has destroyed the original constructional surface making them difficult to distinguish from pediment deposits. Pediment deposits capping ridgeline remnants on the upthrown side of the faults in

northwest Monte Cristo Valley were identified by their relative stratigraphic position with respect to known surfaces and typically have less developed (younger) soils than correlative fan deposits or pediment remnants in other portions of the study area due to post-faulting erosion.

Old Pediment Deposits (Qpla, Qplb): Qpla pediment deposits occur as an isolated remnant capping hill 5912 (elevation in feet) in central Stewart Valley and as one triangular-shaped pediment remnant on the western edge of the valley. The surfaces appear to be stratigraphically higher than Qplb surfaces and are probably mid-Pleistocene in age. The Qplb pediment deposits are best preserved in central Stewart Valley where they occur as linear pediment remnants as much as 50 m (165 feet) above the active drainage channels at their base. Qplb deposits also occur as ridgeline remnants throughout the study area. Surface soils on Qplb pediment remnants are Natrargids and are considered to be mid(?)-to-late Pleistocene in age (Table 2)). The summits of ridgeline remnants have soils as young as latest Pleistocene-early Holocene (Calciorthids).

Intermediate Pediment Deposits (Qp2a, Qp2b): Qp2a pediment deposits are the most extensive Quaternary deposits in the study area. These deposits occur as pediment remnants inset below the Qplb pediment remnants in Stewart Valley, as caps on numerous ridgeline remnants in northern Monte Cristo Valley, and as caps on isolated ridgeline remnants throughout the study area. Most pediment remnant surface soils are Natrargids

considered to be late Pleistocene in age. Qp2a pediment remnants in southernmost Stewart Valley and ridgeline remnants throughout the area generally have latest Pleistocene to early Holocene soils (Camborthids, Calciorthids) on the crest. Qp2b deposits occur as pediment remnants and caps on ridgeline remnants inset from Qp2a pediment deposits in Stewart Valley and northern Monte Cristo Valley. Surface soils on Qp2b pediment deposits are latest Pleistocene to early Holocene in age (Camborthids, Calciorthids).

Playa Deposits (Qpa)

Playa deposits of Kibby Flat in central Monte Cristo Valley are light brown to buff clay, silt and fine sand with some sand and gravel at the playa margins. Lacustrine sediments may interfinger with alluvium at the playa margin but lack of shoreline landforms suggest the playa was not a pluvial lake during the late Pleistocene.

Sand Dunes (Qsd)

Small mounds at the southwest margin of Kibby Flat playa are stabilized by vegetation and surrounded by late Holocene to recent alluvium (Qf3b, Qf3c). These features were not field checked and may be remnants of a lacustrine bar but lack of additional shoreline landforms suggests these are of eolian origin.

LATE CENOZOIC STRUCTURAL GEOLOGY

Introduction

Although previous reports have noted various aspects of the late Cenozoic (post-20 m.y.b.p.) structural geology of the Stewart and Monte Cristo valleys there has been no attempt to map and interpret the structural geology within the valleys. Buwalda (1914, p. 348-349) noted several open folds in the Esmeralda Formation in Stewart Valley with flanks whose "dips seldom exceed ten to fifteen degrees." Gianella and Callaghan (1934a, b) mapped in detail most of the surface faults formed during the 1932 Cedar Mountain earthquake. They noted that most of the surface faulting occurred at the base of pre-existing breaks in slope but did not describe these older scarps. Right-lateral shift was inferred based on the rupture pattern and measured displacements. Mawby (1965, p. 23) noted numerous minor normal faults in northern Stewart Valley, a right-lateral strike-slip fault on the west side of Stewart Valley, and dips of usually less than 20° within the Esmeralda Formation.

The detailed mapping of this study (Plates 1, 2) has delineated a N30°W trending zone of discontinuous faults that controls the late Cenozoic tectonism in Stewart and Monte Cristo valleys (Molinari, 1983). The zone extends for at least 45 kilometers (28 miles) along the west edge and south-central portion of Stewart Valley into Monte Cristo Valley, where it apparently terminates south of Kibby Flat playa

Numerous folds within the Esmeralda Formation east of the fault zone are interpreted to be genetically related to the fault zone. Deformation is post mid-Miocene and has continued to the present.

The southern segment of the fault zone was reactivated in 1932 and previously has been informally referred to as the "Cedar Mountain fault" in reference to the earthquake. It is part of a longer fault zone located west of the Cedar Mountains in Stewart and Monte Cristo valleys which is herein named the Stewart-Monte Cristo fault zone (SMCFZ).

The SMCFZ is defined as the discontinuous series of northwest trending right-normal-slip faults which extend along the west margin of Stewart Valley and into Monte Cristo Valley to the southwest corner of Kibby Flat. The fault zone is subparallel to, and situated east of, the Bettles Well fault and faults along the east flank of the Pilot Mountains, and west of the fault along the west flank of the Cedar Mountains.

Other faults within the area include: 1) north-to-northeast trending, high-angle normal and oblique-slip faults; 2) range front faults along the east flank of the Pilot Mountains, southeast flank of the Gabbs Valley Range, and west flank of the northern Cedar Mountains; and 3) the Bettles Well fault (Nielsen, 1965, p. 1305)[3] including a probable

[3]Nielsen named this fault the Battles Well fault after a well located in the pass between the Pilot Mountains and the Gabbs Valley Range. The most recent 1:24,000 U.S.G.S. topographic map shows this location as Bettles Well and this spelling was confirmed with local residents of Mina, Nevada (J.S. Oldow, personal communication, 1983).

southern extension in the southwestern corner of the area.

For the purpose of discussion the area is divided into three (north, central, south) sections (Figure 8) based on the style and magnitude of deformation. None of the geologic structure in the north and central sections has been previously mapped or discussed in detail. In the southern section the fault scarps, formed in 1932 and described by Gianella and Callaghan (1934a, b), are more accurately located. In addition, this study delineated three additional major fault scarps formed in 1932 (faults D, E, H, Plate 6) and several late Quaternary faults not reactivated in 1932. It presents evidence for the style, magnitude, and recurrence of surface faulting along the south segment of the SMCFZ during the late Quaternary, and refines the previously mapped late Cenozoic structural geology on the margins of the study area.

North Section

The north section extends from the north end of the study area to the major N75°E trending anticline in south-central Stewart Valley (Plate 5, location A). Deformation in this section is expressed as several northwest trending faults in lavas and breccias of the lavas of Mount Ferguson (Tlf and Tlfb) and several minor faults and numerous folds in the Esmeralda Formation (Te).

The north segment of the SMCFZ extends for approximately 13 kilometers (8.1 miles) along the east flank of

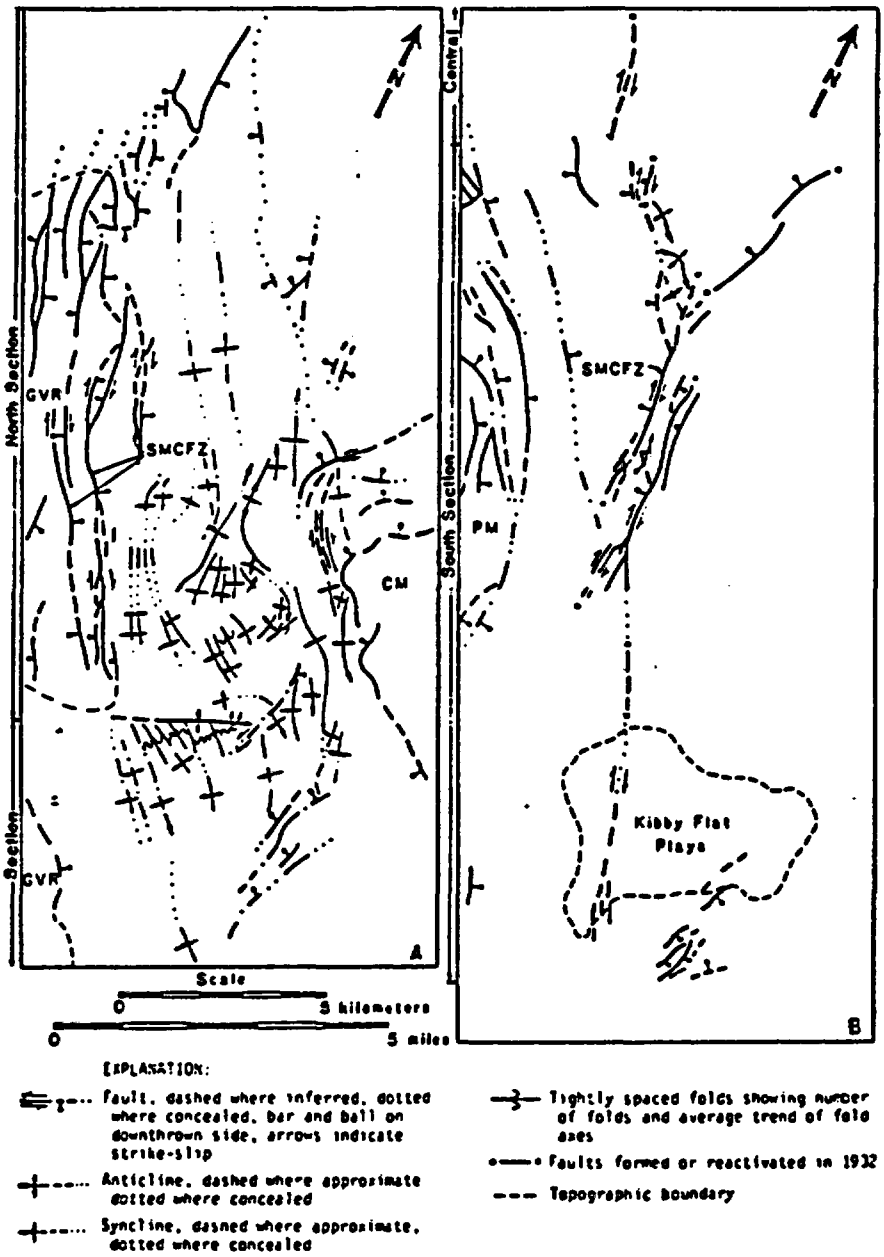

Figure 8 - Map showing north, central, and south sections discussed in text in relation to structural geology of the study area. SMCFZ - Stewart-Monte Cristo fault zone, GVR - Gabbs Valley Range, CM - Cedar Mountains, PM - Pilot Mountains.

Figure 9 - Fault scarp in volcanic debris-flow breccias at the base of the east flank of the Gabbs Valley Range.

Figure 11 - Linear hillside valley along fault trace on east flank of the Gabbs Valley Range. Poleline Spring Wash at bottom, view to the north. "U" on upthrown block, arrows show direction of lateral displacement.

Figure 12 - Angular unconformity of 8° occurs between the volcanic debris flow breccias (Tlfb) and the overlying Esmeralda Formation (Te) on the east flank of Gabbs Valley Range. View to the north. The breccias dip 21°E and the Esmeralda Formation dips 13°E.

Figure 13 - Well-exposed anticline in the Esmeralda Formation in Stewart Valley. Limbs of the fold dip 43°E and 18°W. View is to the south.

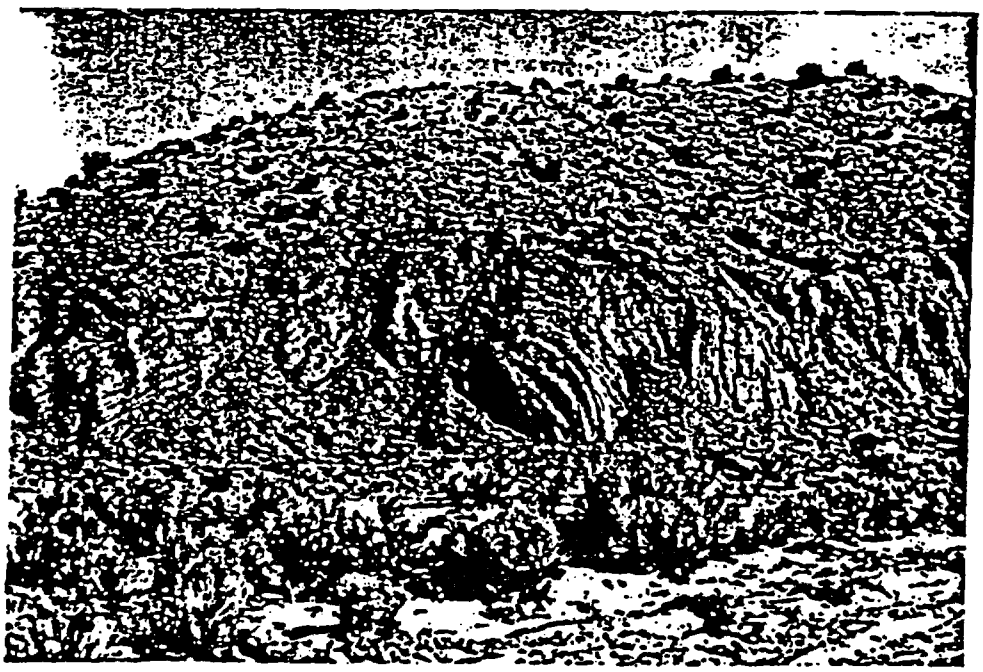

Figure 14 - Core of N75°E trending anticline in the Esmeralda Formation exposed along Finger Rock Wash in Stewart Valley. View to the east. Northern limb dips 33° north and southern limb is slightly overturned and dips 87°W.

(Figure 14). This fold trends approximately normal to the predominant northwest fold trend. East of Finger Rock Wash, this fold either dies out or is truncated by a fault with an arcuate trend (Plate 5, location E). The fault truncates another east plunging anticline which may be the eastern extension of the N75°E anticline; indicating 0.7 km (0.44 miles) of left-lateral slip since the formation of the fold. Structural and stratigraphic relationships along this fault also suggest a vertical component of slip but no definite determination of the amount was possible. The fault could only be traced for approximately 1.4 km (0.9 miles). Late Pleistocene (Qp2) pediment gravels that overlie the fault do not appear faulted but Gianella and Callaghan (1934a, b, figure 2) show some 1932 surface faulting in this area which may have occurred on this fault.

No angular unconformities were noted within the Esmeralda Formation in Stewart Valley, indicating tilting and folding was initiated after deposition of the formation (post-m.y.b.p.). Mid(?)-to-late Pleistocene (Qpl and 2) pediment gravels overlie the folded Esmeralda Formation in an angular unconformity and do not appear to be tilted or folded. At locations where the pediments are well preserved, the average rate of tilting in the Esmeralda Formation estimated for the last 5 to 10 million years indicates a maximum of 2° of tilting after the deposition of the pedisediment, thus, active folding cannot be ruled out.

A northwest trending fault is inferred to exist along Finger Rock Wash in northern Stewart Valley (Plate 5,

location F). Evidence for this fault is the repetition of tuff marker beds in the Esmeralda Formation and the local variation in the strike and dip of bedding on opposite sides of the fault (Plate 1). A branch fault splays to the north and is exposed in Esmeralda beds near the intersection of Stewart Valley Road and Omco Wash. The branch fault has a northerly strike and dips 62° to the west. The similar attitudes of bedding west and east of the two faults suggests the faults postdate the eastward tilting of bedding although rotation of the faults cannot be definitely dismissed. Total normal dip separation across the two faults is approximately 105 m (345 feet) (Plate 3, x-section A-A') and neither fault shows evidence of late Holocene (post-Qf3b) displacement. No evidence for a component of strike-slip on the faults was obtained.

There are certainly additional faults in the north section of the area but due to lack of exposure, poor stratigraphic control, short length, or minor displacement, these faults were not detected or mapped if seen in only one outcrop.

Central Section

The central section extends for approximately 8 km (5 miles) south from the N75°E trending anticline to the "Y"-intersection in northwest Monte Cristo Valley. Bedding in the Esmeralda Formation is folded, faulted, or only gently tilted at various locations in this section and there are

no conspicuous faults in Quaternary deposits. In the northern portion of this section, numerous northwest trending, tightly spaced, gentle to open folds are formed on the south limb of the N75°E anticline (Figure 15). Dips of the limbs range from 9° to 84° and the average wavelength of the folds is less than 30 m (98 feet). None of the folds can be traced for more than 2 km (1.25 miles) to the southwest. The strike of bedding is generally northeast-southwest on the east side of southernmost Stewart Valley, and northwest-southeast on the west side with an overall gentle easterly dip. This is reflected by a Z-shaped bend in the outcrop pattern of tuff beds in this area (Plate 1).

Several faults, expressed as lineaments on aerial photography in southeastern Stewart Valley, repeat and truncate distinctive tuff beds in the Esmeralda Formation (Plate 1 and Plate 5, location B). Poor outcrop exposure and possible lateral facies changes prevented a definite resolution of the structural relationships in this area but normal dip-separation on each fault is probably less than 30 m (98 feet) and no evidence for lateral slip was obtained.

Gianella and Callaghan (1934b, p. 370-372, no. 18, 19) reported several surface fissures parallel to Stewart Valley Road in northernmost Monte Cristo Valley with "...a displacement with the west side down from a fraction of an inch to six inches" and "...a horizontal component of movement in which the east side has moved south with respect to the

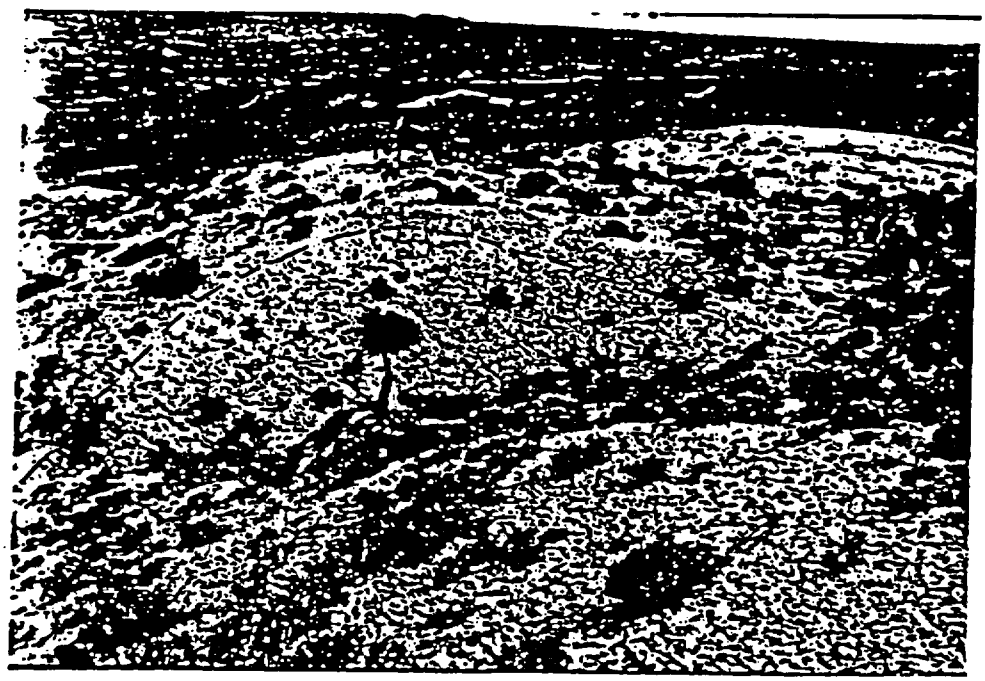

Figure 15 - Small, northwest trending anticline formed in the Esmeralda Formation on the south limb of the major N75°E anticline shown in Figure 14. Fold is typical of the small, tightly spaced folds in the central portion of the SMCFZ. View to the southeast. Limbs dip 45°NE and 25°SW.

west side." These fissures could not be relocated but an inferred fault (Plate 6, Fault A) was mapped based on the right-lateral displacement suggested by Gianella and Callaghan (1934) and a weak photo-lineament south of the bend in Stewart Valley Road.

South Section

The south section extends to the south for approximately 24 km (15 miles) from the "Y" intersection to the south edge of Kibby Flat. In this section, the SMCFZ is expressed as prominent north-to-northwest trending, west-facing scarps in mid(?)-to-late Quaternary alluvium in the northern 12.2 km (7.6 miles) of the section and a northwest trending vegetation lineament across Kibby Flat. Northerly trending faults form a left-stepping en echelon pattern in the central portion of the section (Plate 6). Vertical displacement on the faults has been great enough to locally juxtapose rocks of the Esmeralda Formation against various alluvial units. Bedding in the Esmeralda strikes northwest and dips 6° to 65° east except where northwest trending, gentle to open folds have formed (Plates 2 and 6).

The surface faults associated with the 1932 Cedar Mountain earthquake generally formed along pre-existing fault scarps. Several new faults were formed and some pre-existing faults were not reactivated. During this study, several new faults with surface rupture in 1932 were mapped in addition to those reported by Gianella and Callaghan

(1934a, b, Figure 2, no. 22-25).[4] Fault numbers 20 and 21 (Gianella and Callaghan, 1934b, pgs. 372-373) had short rupture lengths, surface displacements of less than 2.5 cm (6 inches), were not visible on low-sun-angle aerial photography, and thus were not included on the maps in this report.

Fault B (Plate 6) extends for at least 0.6 km (0.4 miles) and Gianella and Callaghan (1934b, p. 374) measured a surface separation of 46 cm (18 inches), up on the east, where it crosses the crest of a ridgeline remnant of Qplb overlying Esmeralda Formation (Figure 16). Measurements of at least 100 cm (39 inches) of normal separation were made during this study. The discrepancy between the two values is either due to an incorrect recording of the value by Gianella and Callaghan or movement related to aftershocks or creep after their investigation. The scarp height increases down the sideslope to the south and decreases to the north indicating a significant component of right-slip (Figure 16). The fault plane is exposed in the Esmeralda Formation on the sideslope of the ridge and dips 65° west.

Fault C (Plate 6) is approximately 5 km (3.1 miles) long. It is expressed as a single scarp, monoclinal ridge, or "molehill" ridge, 45 cm (18 inches) or less in height, to

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[4]Fault numbers 22, 23, 25 of Figure 2 in Gianella and Callaghan (1934a, b) correspond to faults B, C, F, on Plate 6, respectively, and fault 24 is the southernmost graben of E on Plate 6.

Figure 16 - Fault scarp (fault B on Plate 6) formed in pedisediment overlying the Esmeralda Formation during 1932 earthquake. Normal throw at the crest of the ridgeline remnant is approximately 1 meter (39 inches). Scarp height increases downslope indicating a significant component of right-lateral strike-slip. View to the north. Note shovel for scale.

just north of the Monte Cristo Valley Road where it splays into several en echelon complex grabens. Gianella and Callaghan (1934b, p. 362) measured 86 cm (34 inches) of right-lateral separation of a road across the northern portion of the fault. Normal throw in this area is approximately half of the right-lateral component. Normal throw on the graben walls is as much as 120 cm (48 inches) on the west-bounding fault and 60 cm (24 inches) on the east-bounding fault. Within the main graben, smaller grabens, horsts, and compression ridges were formed and locally individual faults exhibit a reversal in the direction of the throw (Figure 17). The larger of these features, including a horst block approximately 15 m (49 feet) in width, are still visible. Many of the small antecedent drainage channels crossing the west-bounding scarp of the graben appear to be displaced in a right-lateral sense. No reliable measurements were obtained during this study, but the right-lateral separation appeared to be $\leq$100 cm (39 inches).

Fault D (Plate 6) was not previously mapped, is located approximately 0.9 km (0.6 miles) east of Fault C, and extends for 5.3 km (3.3 miles). The northern 3.4 km (2.1 miles) of the fault is expressed as a single scarp, whereas the southern portion is comprised of three splays of unequal length. The northern portion and two westernmost splays of the southern portion exhibited surface displacement in 1932. The eastern splay is expressed as a single scarp or narrow graben in late Pleistocene (Qp2a and Qf2b) alluvium. 1932 scarps range from

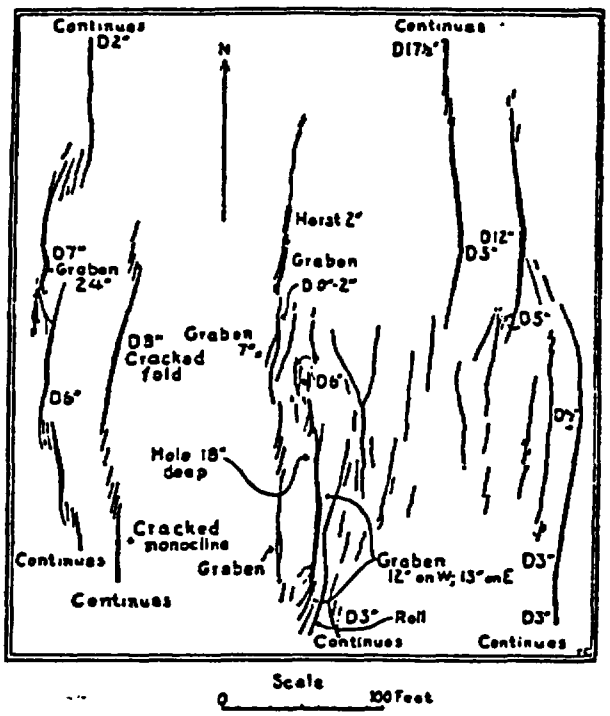


Figure 17 - Sketch from pace traverses showing the pattern of faulting within a portion of the main graben at the south end of fault C on Plate 6. The line weight indicates the relative size of the fissures. "D" represents the downthrown side and surface displacement is in inches. (From Gianella and Callaghan, 1934b, Figure 10).

a monoclinal ridge ≤30 cm (12 inches) in height to a scarp with up to 75 cm (30 inches) of normal throw, down to the west. A section of the 1932 fault reactivated, and exhibited a reversal in throw from, the west-bounding fault of an older graben formed in a late Pleistocene (Qp2a) surface overlying Esmeralda Formation (Figure 18). The Qp2b surface is not displaced by the graben bounding faults. A component of right-lateral slip on fault D in 1932 is indicated by the lateral displacement of antecedent stream channels and the increase in scarp height on south-facing sideslopes cut by the fault. Local ponding of alluvium and erosion along the base of the scarp prevented the accurate measurement of displaced drainage channels. The total scarp height along the central portion of the fault locally exceeds 4.0 m (157 inches) where the higher (older) fan and pediment remnants are preserved on the upthrown block of the fault.

A northwest trending fault is inferred on the south side of a horst located between faults C and D. There is no evidence for surface faulting in 1932 along this inferred fault. The uplifted block is bounded on the west by fault C and on the east by the west-bounding fault of the previously noted "older" graben (Figure 19). An anticline-syncline fold sequence, exposed in the Esmeralda Formation in the uplifted block, is most likely a result of the left-bend formed by the inferred fault and the surface traces of faults C and D. The Qp2a surface overlying the folds is warped into a broad, gentle syncline (Figure 19) suggesting the folds are still

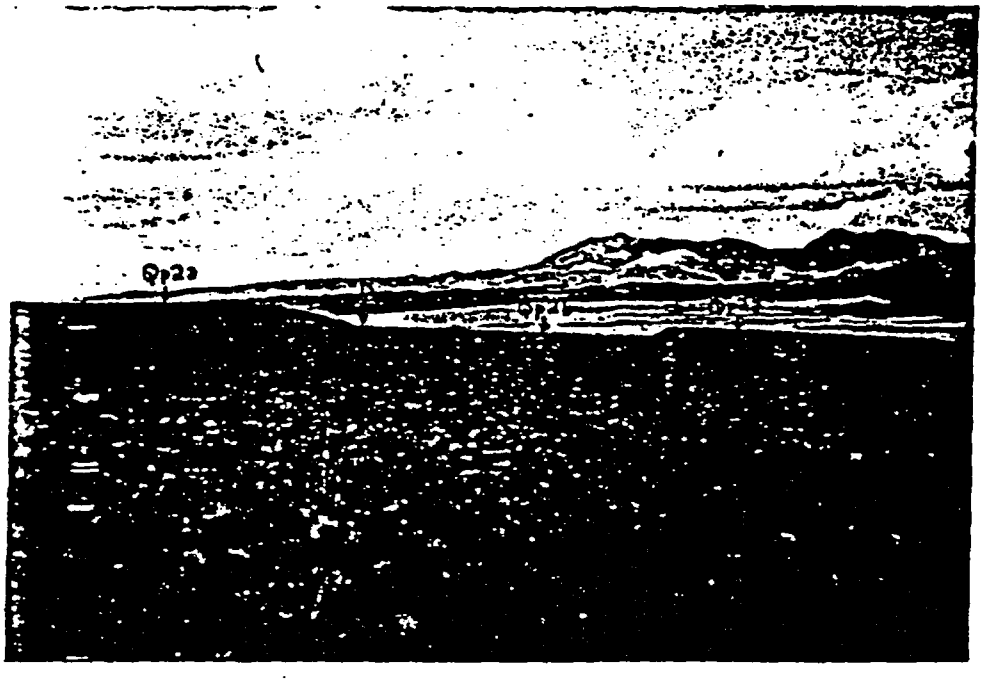


Figure 18 - Graben formed in Qp2a surface but not in Qp2b surface. 1932 surface rupture on fault D (Plate 6) occurred at base of the west wall of the graben but surface displacement was opposite that of the older, graben-forming fault. View to the north in northern Monte Cristo Valley.

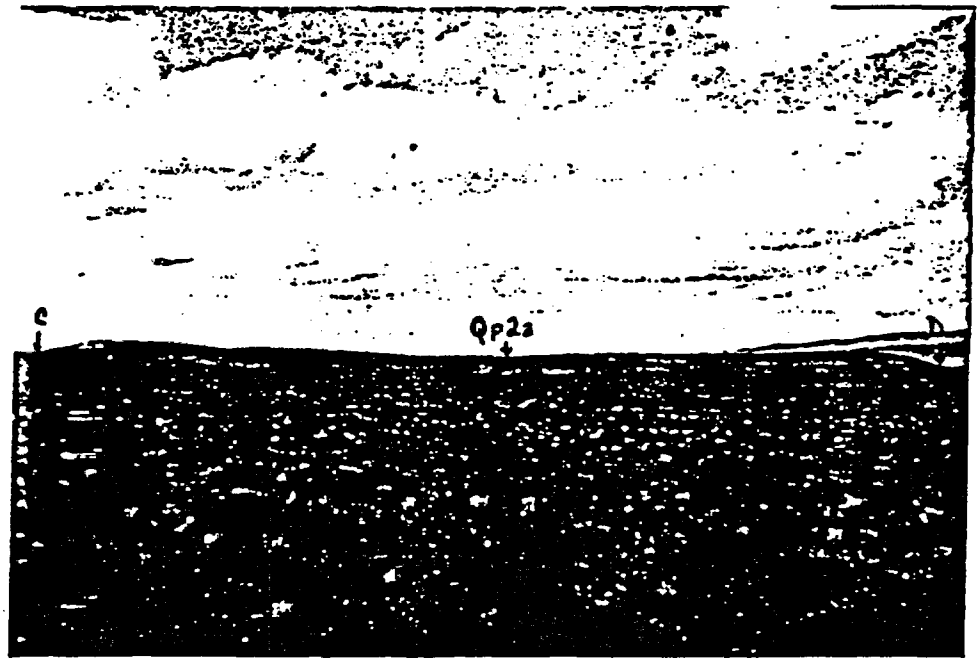


Figure 19 - Horst block between surface traces of faults C and D (Plate 6). Dashed line marks Qp2b surface. Additional fault inferred at base of the horst. Qp2a surface appears to be gently folded into a syncline due to convergence at the left-step between faults C and D. View to the north in northern Monte Cristo Valley.

actively forming.

Fault E (Plate 6) is a branch of the south portion of fault D and is expressed as a discontinuous complex graben 1.7 km (1.1 miles) long. Normal throw of up to 140 cm (55 inches) and 30 cm (12 inches) occurred in 1932 on the west-bounding and east-bounding fault, respectively. En echelon fissures, small open fissures, and small laterally displaced fault slices are preserved in lag deposits of cobbles and small boulders within the graben. Right-lateral slip ranging between 100-200 cm (39-79 inches) on the west side of the graben is indicated by displaced drainage channels and crests of linear lag deposits (Figure 20). Measured displacements indicate lateral to vertical slip ratios from 1:1 to 2:1 on the west-bounding fault.

Fault F (Plate 6) was not detected on the low-sun-angle aerial photographs but is inferred based on the account of Gianella and Callaghan (1934b, p. 375) in which they noted a "molehill" ridge and fissure "...at the northeast side of a low, broad hill northwest of the playa..."

Fault G (Plate 6) is inferred based on the northwest trending vegetation lineament on the west side of Kibby Flat playa (Figure 21). This lineament is transverse to and locally appears to deflect the vegetation bands around the edge of the playa. No displacement of the playa surface was observed but the vegetation lineament is considered to be structurally controlled. The fact that the lineament has the same trend as the SMCFZ and is aligned with the south

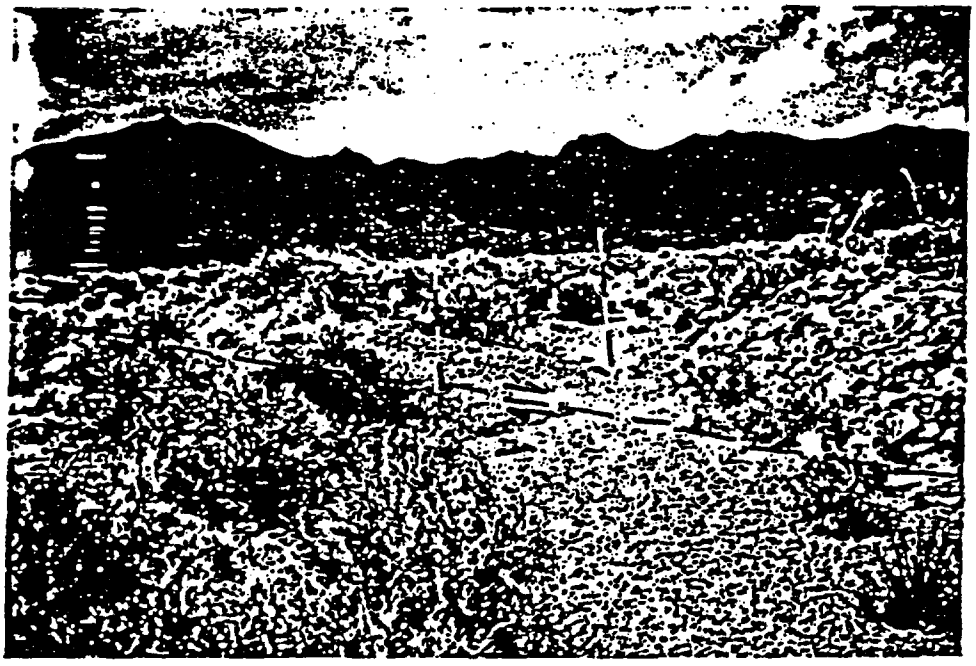


Figure 20 - Edge of small drainage channel displaced 1.6 m (62 inches) on west-bounding fault of graben (fault E, Plate 6) formed during 1932 earthquake. Staffs mark edge of the drainage. View to the west in southern Monte Cristo Valley. Pilot Mountains in the background.

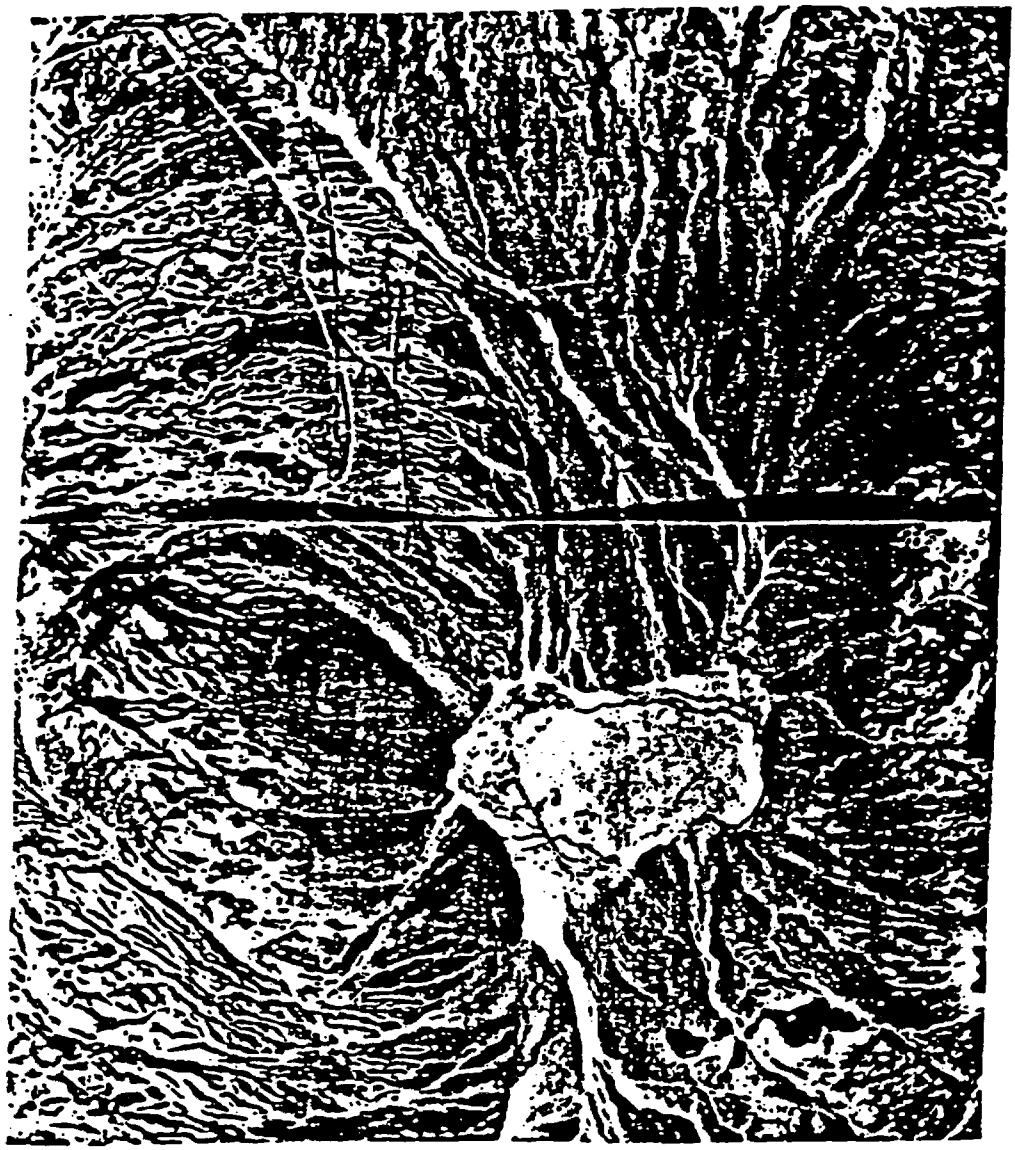


Figure 21 - Photo mosaic showing relationship of vegetation lineament across Kibby Flat playa (a) to faults of the southern section of the SMCFZ. Note apparent deflection of vegetation bands at south end of the playa. Pilot Mountains on the left and southern Cedar Mountains on the right edge of the photo.

end of Fault D and Fault F also supports this relationship. The lineament splays into four branches at the south end of the playa where it apparently terminates. Strike-slip faults commonly end with several splays (Freund, 1974).

Several northeast trending, left-stepping en echelon faults comprise Fault H which branches from Fault C in north-central Monte Cristo Valley (Plate 6). Fault scarps formed in 1932 have a maximum height of 75 cm (30 inches) and a cumulative length of approximately 3.5 km (2.2 miles). A 65° northwest dipping fault plane is exposed in the Esmeralda Formation at one location along the fault. The scarp height is relatively consistent on sideslopes crossed by the fault indicating little or no strike-slip. There is no evidence of a pre-existing scarp prior to the 1932 surface faulting.

Late Quaternary Recurrence on the SMCFZ, South Segment: Detailed trenching studies on the San Andreas fault (Sieh, 1978) and the Wasatch fault (Swan and others, 1980) suggest individual faults or fault segments may have a "characteristic" earthquake in which accumulated stress is released at relatively regular recurrence intervals by earthquakes with similar magnitudes and surface displacements, rather than numerous earthquakes of various magnitudes and displacements. The 1932 Cedar Mountain (M = 7.2 to 7.3) earthquake is the third largest historic earthquake in the Basin and Range Province, exceeded only by the 1872 Owens Valley, California (M = 8±) and 1915 Pleasant Valley, Nevada (M = 7.75) earthquakes. Thus, it is reasonable to assume that the 1932 event

is probably the maximum earthquake for the SMCFZ and the measured displacements that formed in 1932 are "characteristic" for the 1932-sized events on the SMCFZ. Measuring the surface displacement that occurred in 1932 and dividing this value into the total scarp height above the top of the 1932 scarp, will give an estimation of the number of 1932-sized events that occurred after the formation of the surface at the summit of the scarp and prior to 1932. This method assumes that the SMCFZ is activated by a "characteristic" earthquake and that the deformation at depth is propagated to the surface as fault rupture and not folding or detachment. This may not be true for the SMCFZ as will be discussed in a later section on a tectonic model for the fault zone.

Geomorphic and stratigraphic evidence indicative of multiple displacements on a fault include: 1) distinct breaks in slope on a scarp profile, 2) benches or terraces along channels dissecting the scarp, and 3) greater displacement of older deposits or surfaces than younger deposits of surfaces (Wallace, 1977). Terraces can be of depositional or erosional origin. The preservation of benches or terraces along stream channels can be the result of local and regional tectonics, or climatically induced changes in fluvial processes. The effects of these processes must be differentiated for purposes of active fault evaluation. The preservation of benches and terraces as the result of surface displacements on the fault provide insight into the number of surface faulting events and rates of uplift on the fault.

Benches and terraces are commonly preserved on the upthrown (upstream) block of typical range-front faults in the Basin and Range Province due to rapid downcutting following surface faulting as the channel attempts to reestablish grade with the channel floor at the base of the scarp. The faults in the south segment of the SMCFZ are located near the toe of the fan piedmont and the upthrown block is generally on the downstream side of the fault. Although this scenario is quite different from the more typical range-front fault, the preservation of geomorphic surfaces on the upthrown block is still useful for evaluating fault history.

Older erosional fan remnants and nonburied fan remnants may occur on the fan piedmont at higher relative elevations than the surrounding younger component landforms strictly due to changes in the erosional-depositional process through time. These older surfaces should be accordant with similar aged surfaces preserved upslope near the head of the piedmont. The older surfaces preserved on the piedmont due to uplift on a fault are higher than the projected elevation of a similar aged surface across the fault. This latter relationship occurs on the south segment of the SMCFZ.

The presence of late Pleistocene scarps without demonstrable Holocene displacement, scarps with late Pleistocene and Holocene displacement that were reactivated in 1932, and the formation of new faults in 1932 indicate recurrent events on a continually evolving fault zone. Along most of the preexisting scarps of faults C and D (Plate 6), the surfaces

on the west (downthrown) block are younger than the highest surfaces on the east (upthrown) block; thus the total scarp height is the minimum displacement of the older surfaces. The highest surfaces on the upthrown block are erosional remnants and the unit designation was based on the relative stratigraphic position of the surface with respect to a known surface (i.e., a surface remnant higher than a known Qf2b surface is mapped as Qp2a even though soil on this remnant is younger than that of the summit of a Qf2a surface). Consequently, the amount of throw between a Qf2a surface on the downthrown block and a Qp2a surface on the upthrown block would be a minimum value and the age of the upper surface would be younger than the initial faulting event that displaced the original Qf2a surface.

Stratigraphic and geomorphic relationships along faults of the SMCFZ south segment suggest at least three and probably five or six pre-1932, post-Qf2a (latest Pleistocene and Holocene) surface faulting events. In addition, several pre-Qf2a, post-Qf1b (late Pleistocene) events are suggested by scarps in older geomorphic surfaces that were not reactivated in 1932. Since the total scarp heights used are minimum values, the total number of post-Qf2a events inferred from these data is probably a minimum value. An alternative is that there were fewer post-Qf2a surface faulting events that produced larger surface displacements than the 1932 earthquake. Evidence for the numbers, magnitude of displacement, and age of pre-1932 surface faulting events on the individual

fault traces of the southern segment of the SMCFZ is presented subsequently.

Fault I (Plate 6) is located at the "Y" road intersection in northwest Monte Cristo Valley and has a throw direction of up on the east. On the east side of the fault, isolated remnants of Qf1b alluvium unconformably overlie east-dipping sediments of the Esmeralda Formation. These remnants lack the well-developed argillic soil formed on the surface of the Qf1b alluvium on the west side of the fault; thus the several meter high scarp represents the minimum post-Qf1b displacement on this fault. Where the fault crosses the middle road in the "Y" intersection, the scarp in the Qf2a surface is less than a meter. This relationship indicates several post-Qf1b, pre-Qf2a surface faulting events which are not exhibited anywhere else in the area and only minor post-Qf2a displacement on fault I. This fault was not reactivated in 1932.

Fault B (Plate 6) is located at the base of a pre-existing scarp on a ridgeline remnant of Qf1b overlying the east-dipping Esmeralda Formation. The total scarp height is approximately three times the vertical separation of the ridgeline in 1932 suggesting two prior events.

Near the north end of fault C (Plate 6) the 1932 scarp is 40 cm (16 inches) high and occurs near the base of a pre-existing scarp 210 cm (83 inches) high in Esmeralda Formation overlain by a Qp2a surface at the scarp summit (Figure 22). If 40 cm (16 inches) is assumed to be the typical surface displacement on this portion of the fault for a 1932-sized

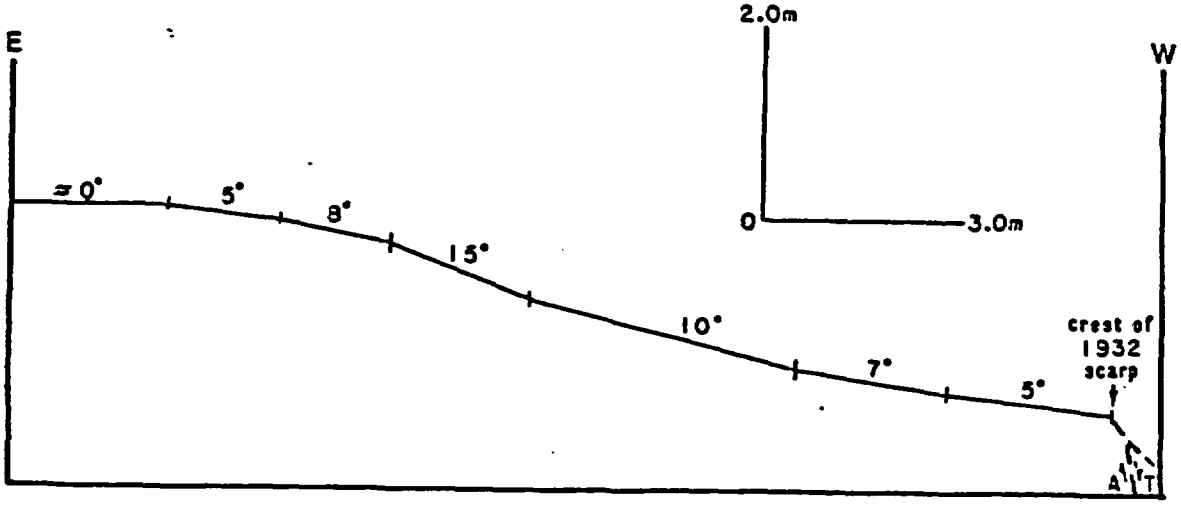
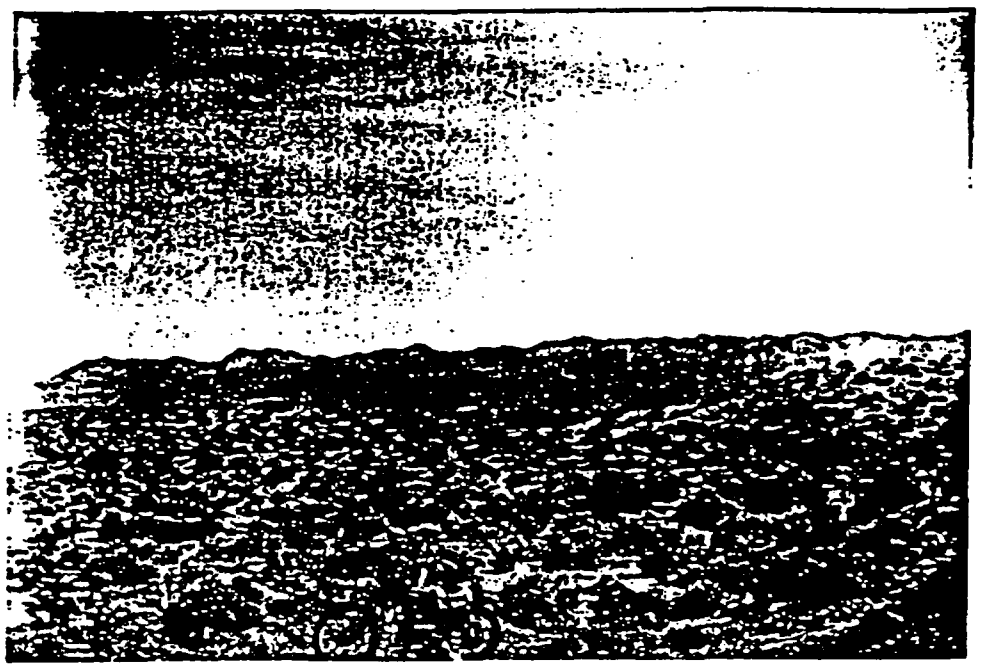


Figure 22 - Photo of 1932 scarp (dashed line) formed on fault C (Plate 6) at the base of a pre-existing scarp. 1932 scarp height is 40 cm (16 inches) at the staff. Scarp profile is from the top of the 1932 scarp to the crest. View to the south.

event, then there may have been as many as five post-Qp2a events prior to 1932. The scarp profile suggests only two displacement events of 90 cm (35 inches) and 1.2 m (47 inches) during this time interval (Figure 22).

The central portion of the 1932 scarp along fault D (Plate 6) formed at the base of a pre-existing scarp which varies in height depending on the age of the preserved remnant on the upthrown block of the fault. Figure 23 shows two higher, older surfaces above the 1932 scarp on the upthrown block of the fault. The lower of these two surfaces (Qp2b) is on the summit of a pre-existing scarp that is twice the height of the 1932 scarp formed at its base, suggesting two post-Qp2b, pre-1932 events. The higher (Qp2a) surface suggests at least one and probably two or three pre-Qp2b, post-Qp2a events. Similar relationships occur at other locations on the central and southern portions of this fault. Additional evidence for these older events is a prominent post-Qp2a graben (Figure 18). The Qp2b surfaces on the graben floor were apparently not displaced by the graben-bounding faults until the 1932 earthquake reactivated a short portion of the west-bounding fault.

There was no prominent pre-existing scarp at fault E (Plate 6) but scarp profiles (Figure 24) across the west-bounding fault of the graben indicate a break in slope above the 1932 scarp. If there was a surface rupture event on this fault prior to 1932 it was mid-to-late Holocene (post-Qf3a) in age since the scarp profile in both Qf3a and Qf2b surfaces



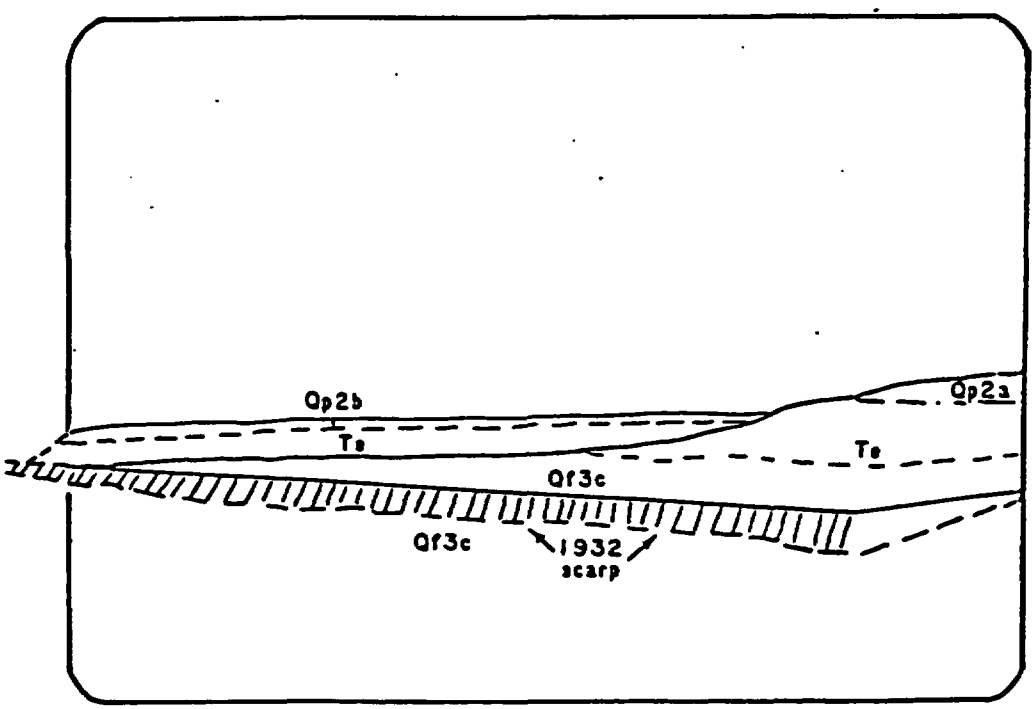
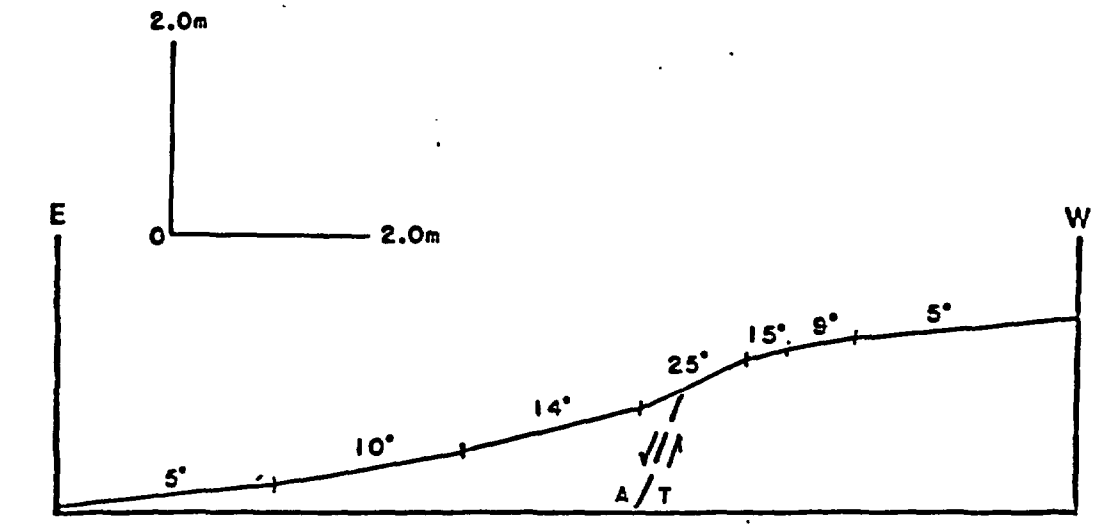


Figure 23 - Photo showing 1932 fault scarp and two uplifted pediment remnants on fault D (Plate 6) in Monte Cristo Valley indicating several 1932-sized surface faulting events post-Qp2a.



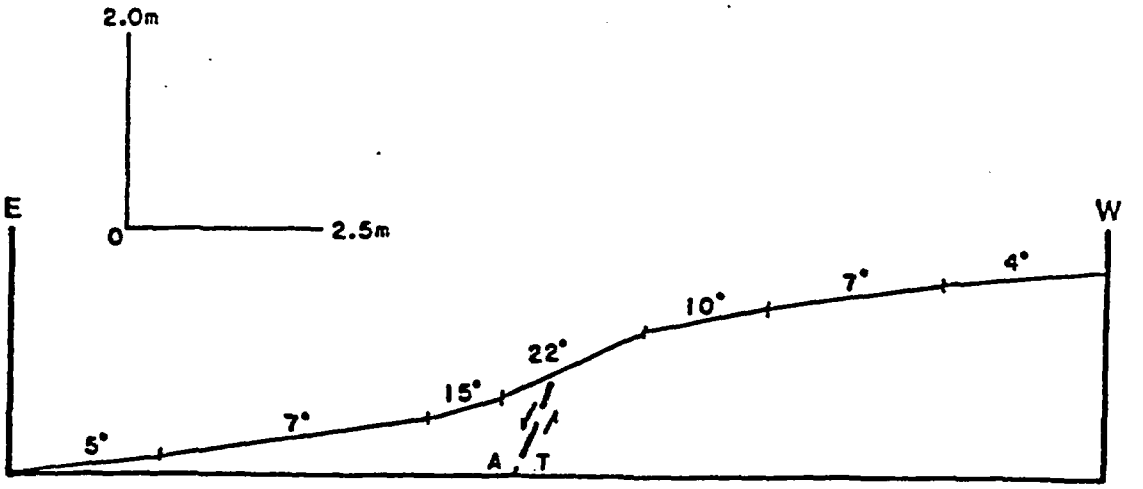


Figure 24 - Scarp profiles across the west-bounding fault of a graben (fault E, Plate 6) formed during the 1932 earthquake. Break in slope at crest of the scarp on both profiles suggests a small scarp existed prior to 1932.

are similar. In addition to the break in slope above the top of the 1932 scarp, the alluvium on the shoulder is considerably coarser than the summit suggesting disruption of the desert pavement prior to 1932. The distal edge of linear lag deposits of gravel and cobbles deposited prior to 1932 generally occurs within the boundaries of the 1932 graben, and deposition may have been controlled by a minor pre-existing depression. Two Holocene age events indicate a recurrence interval on the order of several thousand years.

Additional Late Cenozoic Structural Geology

In addition to the SMCFZ, faults mapped on Plates 1 and 2 include the Bettles Well fault and range-front faults on the west flank of the northern Cedar Mountains, east flank of the Pilot Mountains, and southeast flank of the Gabbs Valley Range.

The Cedar Mountains range-front fault is marked by a sharp break in slope, angular discordance between the Esmeralda Formation on the west side of the fault and Tertiary volcanic rocks on the east, and the local emplacement of intrusive breccias along the fault trace. A late Pleistocene (Qp2) pediment surface is not faulted where it crosses the inferred trace of the fault. Near Stewart Spring the fault makes a northeast bend following the range front and several branches splay off to the northwest. These branch faults are expressed as lineaments on aerial photography and slickensided surfaces and displaced bedding in the

debris-flow breccias (Tcmb) that outcrop north of the road to Stewart Spring. The easternmost of these faults has horizontal striae on the fault surface and there is right-lateral separation of the Te-Tcmb contact of approximately 100 m (300 feet). The 1932 surface faulting west of Stewart Spring (Gianella and Callaghan, 1934b, p. 370) probably occurred on these faults.

North of Stewart Spring (Plate 5, location H) several northeast trending, steeply dipping faults crosscut the northwest striking, east dipping Esmeralda Formation. These faults are expressed by slickensided surfaces, angular discordance of bedding on opposite sides of the fault, truncation of a tuff marker bed, and juxtaposition of varying lithologies. Apparent vertical separation is down to the north and a late Pleistocene pediment surface (Qp2) overlying one of the faults is not displaced.

A fault along the southeastern range front of the Gabbs Valley Range is inferred based on the sharp break in slope at the base of the range which is in line with a splay of the Bettles Well fault mapped by Ekren and Byers (1978). Late Pleistocene fan remnants (Qf2a) are not displaced by the fault.

The Bettles Well fault was mapped in the Pilot Mountains and Gabbs Valley Range by Ferguson and Muller (1949) and Nielsen (1964). More detailed work in the Bettles Well 7.5' quadrangle by J.S. Oldow (written communication, 1983) and the Sunrise Flat 7.5' quadrangle by Ekren and Byers (1978)

has better defined the fault trace. Neilsen (1965) noted 1.6 km (1 mile) of right-lateral separation of a Mesozoic thrust fault and steeply dipping late Oligocene - early Miocene volcanic rocks on the Bettles Well fault in the northeastern Pilot Mountains. The mapping by Oldow (written communication, 1983) indicates 2.0-2.4 km (1.24 - 1.5 miles) of right-lateral shift (slip plus drag) of a fault contact in the volcanic rocks in this area.

Interpretation of aerial photographs and field reconnaissance during this study has delineated the trace of the fault in the southwestern corner of the Stewart Spring 7.5' quadrangle and a probable southern extension in the southwestern corner of the Kibby Flat 7.5' quadrangle. Evidence for the fault in the Stewart Spring quadrangle is a topographic lineament and sheared and brecciated limestone of the Luning Formation. In the Kibby Flat quadrangle (Plate 6, fault J) a northwest trending fault is inferred from a photolineament delineated by scarps in late Pleistocene (Qf2b) fan deposits and the rapid change in dip of a basalt flow from essentially horizontal to 60° south near the inferred trace of the fault. Additional evidence for a buried fault is the approximately 2 km (3.2 miles) of apparent right-lateral separation of the Pilot Mountains eastern frontal fault west of the mapped area (Plate 7).

The Pilot Mountains are bounded on the east by a major north to northwest trenddng fault that makes a westerly bend in strike at the north end of the range and is truncated by

the Bettles Well fault (Plate 6). This fault juxtaposes the Triassic Luning Formation against late Oligocene-early Miocene andesitic lavas and breccias. South of the Good Hope Mine, the fault is situated within the Luning and Mina Formations. North of the mine several northwest trending faults parallel the main fault and at least one of these faults was reactivated during the 1932 earthquake (Ferguson and Muller, 1949, p. 37). This reactivation is considered to be a secondary effect of the earthquake.

There is no evidence for displacement of the Quaternary alluvial fans on the east flank of the Pilot Mountains but south of the Good Hope Mine the slope of the fan surfaces and the rate of channel downcutting are considerably greater than throughout the rest of the study area. The soils formed on the fan surfaces indicate a latest Pleistocene-early Holocene (Qf2b) age for these fans although the fan morphology is more typical of older (Qf2a) fans. This relatively abnormal fan morphology is in part due to the close proximity of the range front to the basin floor, but is also of tectonic origin. Despite the lack of fault scarps in the alluvial fans, the steep gradients and geomorphology of the piedmont are indicative of relatively high uplift rates (Bull, 1977) and may be related to eastward tilting of the range. The very steep range front and prominent scarp on the west flank of the Pilot Mountains indicates a much higher uplift rate than on the east flank which would result in an eastward tilt of the range. Another possible contributing effect could be subsidence of

Kibby Flat which would lower the base level of the system and cause increased downcutting and transfer of the loci of deposition downfan.

There are several northwest trending gentle folds in the Esmeralda Formation in northeastern Monte Cristo Valley, including two folds north of the mapped area that were mapped by Mottern (1962). These folds are probably too far away from the SMCFZ to be directly related to it and no other faults have been mapped in the area to which these folds can be genetically related. They most likely are associated with buried faults in the area and/or uplift of the northern Cedar Mountains.

Several minor northeast trending fault scarps are located south of Kibby Flat in late Pleistocene (Qf2a) fans and older units. These faults are interpreted as secondary faults conjugate to the main trace of the SMCFZ. One fault is formed in a latest Pleistocene (Qf2b) fan and can be traced as a vegetation lineament across recent alluvium (Qf3c) onto the playa floor. Two short vegetation lineaments on the playa floor are parallel to this fault and may be fault related. No displacement on the playa floor could be detected but minor reactivation in 1932 is possible.

TECTONIC MODEL

Any model for the post mid-Miocene tectonic development of Stewart and Monte Cristo valleys must take into account several factors: the apparently simultaneous development of northwest trending folds and right-normal oblique-slip faults of the SMCFZ; the discrepancy between the amount of demonstrable late Quaternary faulting on the north and south segments of the SMCFZ; and the relatively undeformed central section of the area. Other anomalous structural relationships include: the absence of 1932 surface faulting in Stewart Valley when it occurred to the north, south, and east; the formation of the major N75°E trending fold normal to the structural trend of Stewart Valley; and the variation in the apparent number and size of late Quaternary surface faulting events on individual faults along the southern segment of the SMCFZ.

The structural style and trends in Stewart and Monte Cristo valleys indicate the SMCFZ is a right-lateral wrench fault zone. The seemingly incompatible structural elements previously noted can be best explained by various aspects of wrench tectonic models based both on laboratory models (Tchalenko, 1970; Wilcox and others, 1973; Odonne and Vialon, 1983) and the style and trend of structures associated with major well-developed wrench faults (Moody and Hill, 1956; Garfunkel, 1966; Crowell, 1974; Freund, 1974; Mann and others, 1983).

Wrench Folds

Folds related to wrench faults have initial trends subparallel (<45°) to the fault. With increasing deformation, the angle between the fold axes and the controlling fault decreases and the fold axes may become curved (Wilcox and others, 1973; Odonne and Vialon, 1983). Garfunkel (1966) did a geometrical analysis of "drag" folds associated with wrench faults and concluded that formation of the folds indicates that shortening normal to the fault must have occurred. Clay-cake (Wilcox and others, 1973) and paraffin wax (Odonne and Vialon, 1983) models show that folds may occur without convergence when plastic materials overlie solid materials experiencing brittle deformation along a wrench fault; but convergent wrenching exhances compressional wrench-related structures. A left bend on a right-slip fault causes convergence which forms folds subparallel to the fault (Figure 25).

The subparallel, counterclockwise pattern of folds with respect to the faults of the SMCFZ is similar to major right-lateral wrench faults such as the Alpine fault in New Zealand and Barisan Mountains (Semangko) fault in Sumatra (compare Plate 5 with Figure 26). Although the overall fault-fold pattern of the SMCFZ is similar to other wrench zones, the folds do not exhibit the true en echelon pattern of shear box models. There are several northeast trending folds, and the numerous tightly spaced folds in the central section are anomalous with respect to the number and spacing of folds

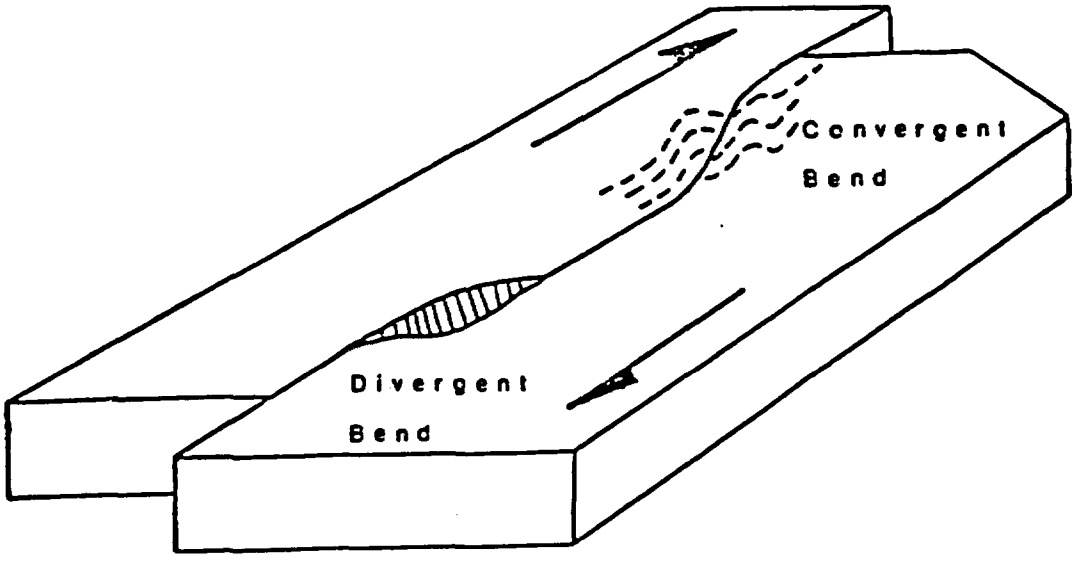


Figure 25 - Right-slip on fault with distinct bends results in tensional structures at diverging bends and uplift and en echelon folds at converging bends (after Crowell, 1974).

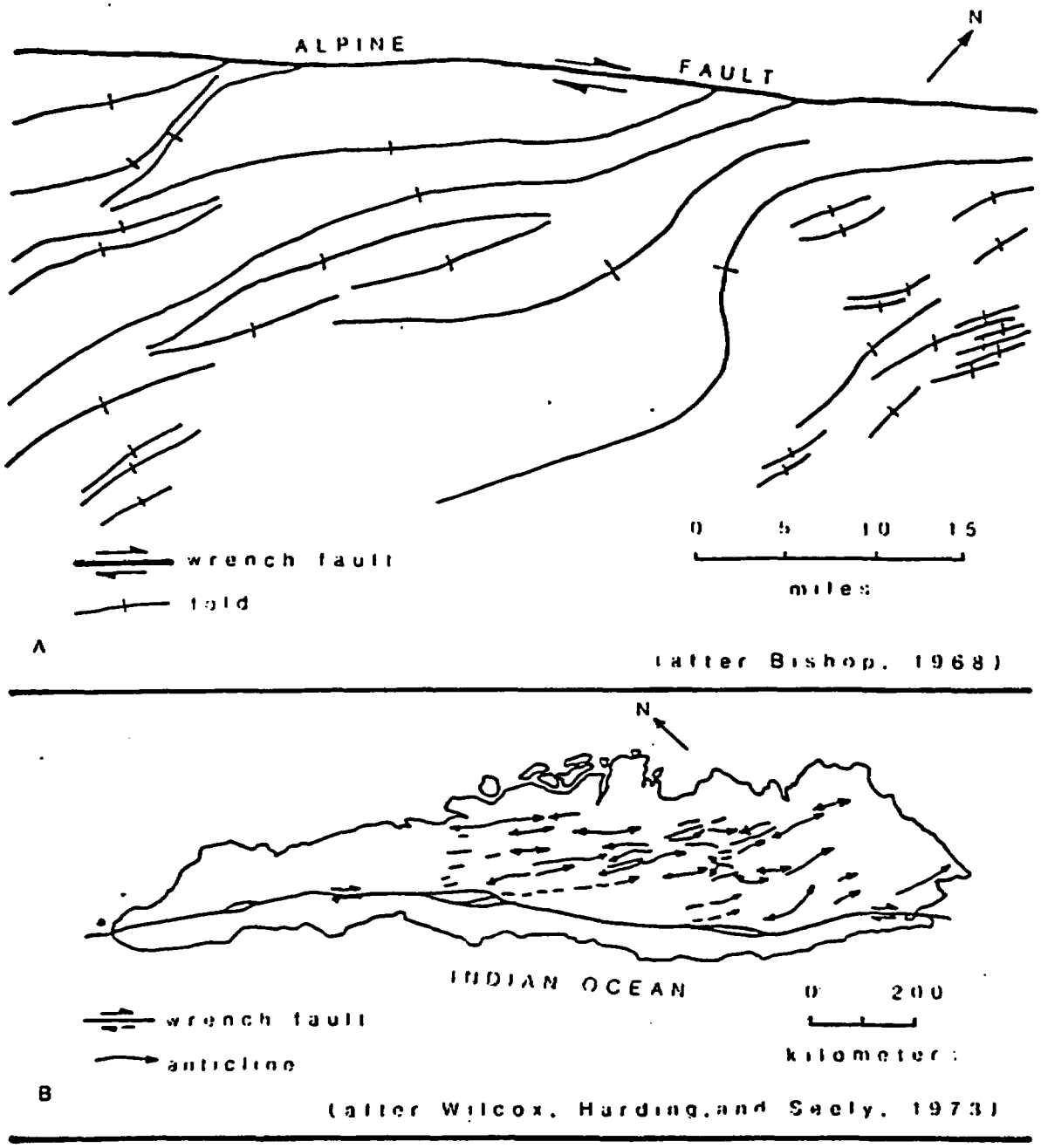


Figure 26- En echelon wrench folds along A) the central portion of the Alpine fault, New Zealand and B) the Barisan Mountains (Semangko) fault, Sumatra.

in the rest of the area. Wilcox and others (1973) noted that in nature, fold trends, fold spacing, and fold shape may be variable due to factors such as the magnitude of convergence and vertical displacement within the wrench zone. Variable lithologies, the thickness of sediments overlying the faulted basement, and rate of deformation also affect fold development. It is also possible that some of the folds in easternmost Stewart Valley may in part be related to drag effects on the Cedar Mountains range-front fault.

The major N75°E trending fold that forms the boundary between the north and central sections of the study area is best explained by a left-step or sharp bend in the fault zone at depth causing strong convergence transverse to the fault trend (Figure 27). This structural geometry is similar to that formed at a left-step on the more well-developed Coyote Creek fault in southern California (Sharp and Clark, 1972).

En Echelon Faults

The south segment of the SMCFZ consists of northerly trending oblique-slip faults B, C, D, E (Plate 6) in a left-stepping en echelon pattern between northwest trending faults A, F, G (Plate 6) to form an overall "lazy Z" shaped (right-stepping) fault pattern. The en echelon oblique-slip faults resemble Reidel shears which form at low angles (<30°) to the slip direction during the initial stage of shear box modeling of wrench faults (Tchalenko, 1970; Wilcox and others, 1973). Conjugate Reidel shears form at high angles to the slip

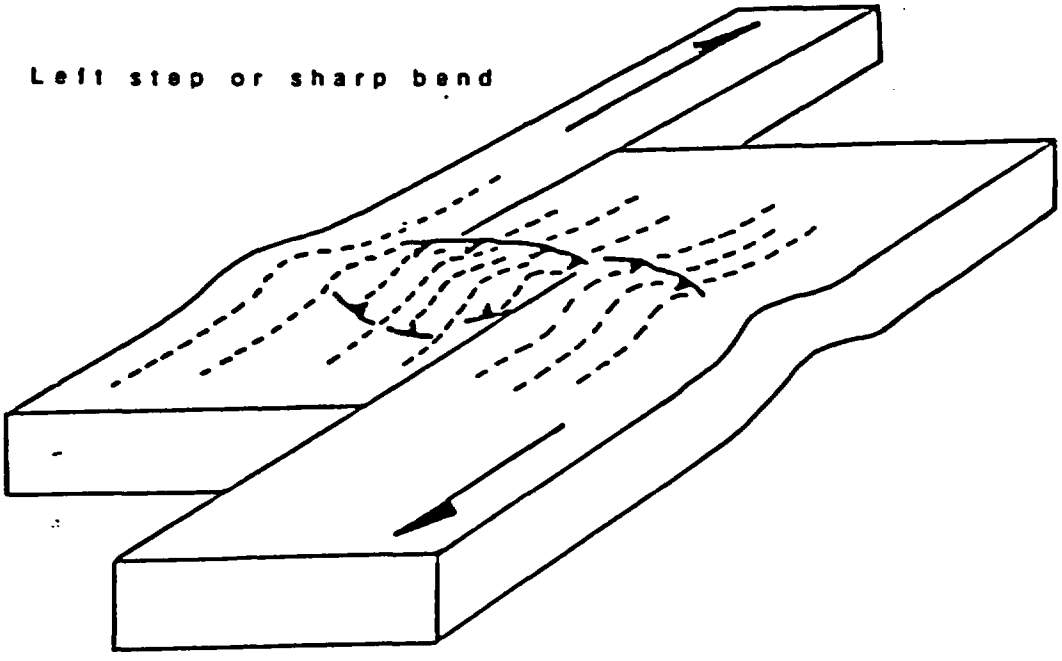


Figure 27 - Strong convergence at a left step or sharp bend on a right-slip fault results in folds and possible reverse or thrust faults (after Crowell, 1974).

direction in the shear box experiments. The mapped fault pattern is similar to the "peak structure" stage of evolution where individual Reidel shears accomodate up to 50% of the total slip at depth (Tchalenko, 1970).

On a small scale, Gianella and Callaghan (1934a, b) measured nine shears formed in 1932 with trends of N45°E to N83°E. These are analogous to the conjugate Reidel shears. Sixteen measured shears had trends between N5°W and N25°W and are analogous to Reidel shears. Thirty measured shears had trends between NS and N28°E. Some of these may be Reidel shears but the majority of these shears occurred east and north of Stewart Valley on pre-existing faults which had trends more characteristic of basin-range structure than Walker Lane structure. These faults are herein interpreted to be secondary to the main fault that was the earthquake source. The northeast trending normal and oblique-slip faults in eastern Stewart Valley, south of Kibby Flat playa, and in northern Monte Cristo Valley (faults H, Plate 6) may be larger scale conjugate Reidel shears.

An alternative to the Reidel shear origin for the en echelon faults is a right-stepping bend in the fault at depth causing divergence and tensional structures (Figure 25). The "lazy Z" pattern with oblique-slip faults connecting the ends of the master faults is characteristic of the initial stages of development of pull-apart basins on right-lateral wrench faults (Mann and others, 1983). The fact that the area of oblique-slip faulting is both a structural and topographic

high supports the Reidel shear origin for the faults.

Evidence for Detachment

Detachment faults associated with the major, well-developed right-lateral wrench faults in the nearby Gabbs Valley and Gillis Ranges occur at many stratigraphic levels in incompetent nonwelded tuffs within the Tertiary section and a major detachment occurs at the Mesozoic-Tertiary contact throughout most of this area (Hardyman and others, 1975; Hardyman, 1978; Ekren and others, 1980). Two of the ash-flow tuffs which are detached in the Gabbs Valley and Gillis Ranges also outcrop in the northern Cedar Mountains. Therefore, it is probable that these units underlie Stewart Valley and may be detached. The Tertiary volcanic rocks outcropping along the aforementioned major wrench faults exhibit brittle deformation. This suggests the folds mapped in the Esmeralda Formation do not occur in the underlying volcanic rocks and are "rootless."

This is supported by wrench models of Wilcox and others (1973) and Odonne and Vialon (1983). In some of their experiments Wilcox and others (1973) placed a thin sheet of plastic film interlayered in the clay. The results showed that the sheet enhanced the number and rate of formation of folds and the shallower the sheet, the smaller and more closely spaced the folds. The plastic sheet allowed the material above it to deform independently from the underlying material. This would be analogous to a detachment zone within the Esmeralda

Formation above the faulted basement. Odonne and Vialon (1983) placed grease between sheets of paraffin wax overlying the rigid basal blocks of their model. The grease coatings act as incompetent layers along which bedding plane slip may occur. This is also analogous to detachment and produced similar results.

There is an abundance of incompetent lithologies within the Esmeralda Formation such as paper shales and nonwelded air-fall tuffs which provide numerous potential zones for detachment. If detachment fault(s) are present at the base of, and/or within the Esmeralda Formation, some of the anomalous structure may be explained. The lack of faults and significant folding throughout much of the central section can be accounted for if the lateral slip on the wrench fault at depth is being accomodated by horizontal displacement on an overlying detachment fault(s). Uplift during development of the N75°E anticline may have caused ramping of the detachment(s) towards the surface causing the numerous tightly spaced folds to form with continuing displacement on the wrench zone. Detachment could also have influenced the trends of folds in the northern section. An example of this style of detached folding is the en echelon mesoscopic scale folds and shears in detached surficial turf layers related to oblique-slip faulting during the 1969 Pariahuanca earthquake in central Peru (Philip and Megard, 1977).

Figure 28 presents a model for the development of detachment faults within the SMCFZ. This model is a modified and

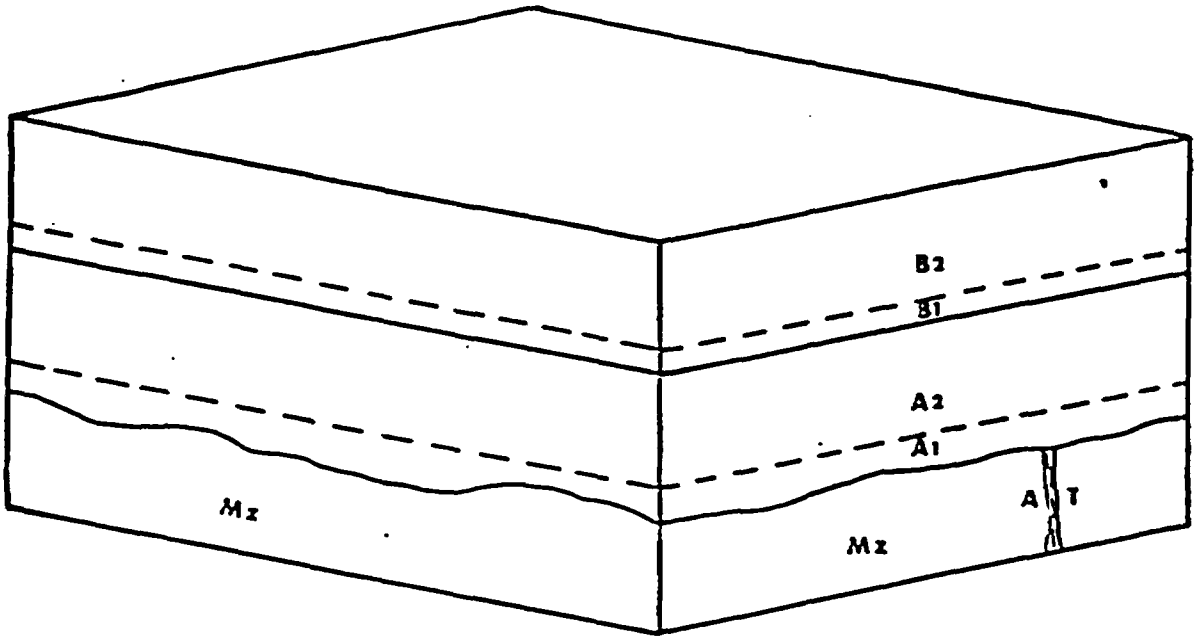


Figure 28a - Simplified diagram depicting strike-slip fault in Mesozoic rocks overlain by Tertiary welded ash-flow tuffs and lavas (A2) with a basal nonwelded ash-flow tuff (A1) and Tertiary sediments (B2) with incompetent strata (B1) at or near the base (after Hardyman, 1978).

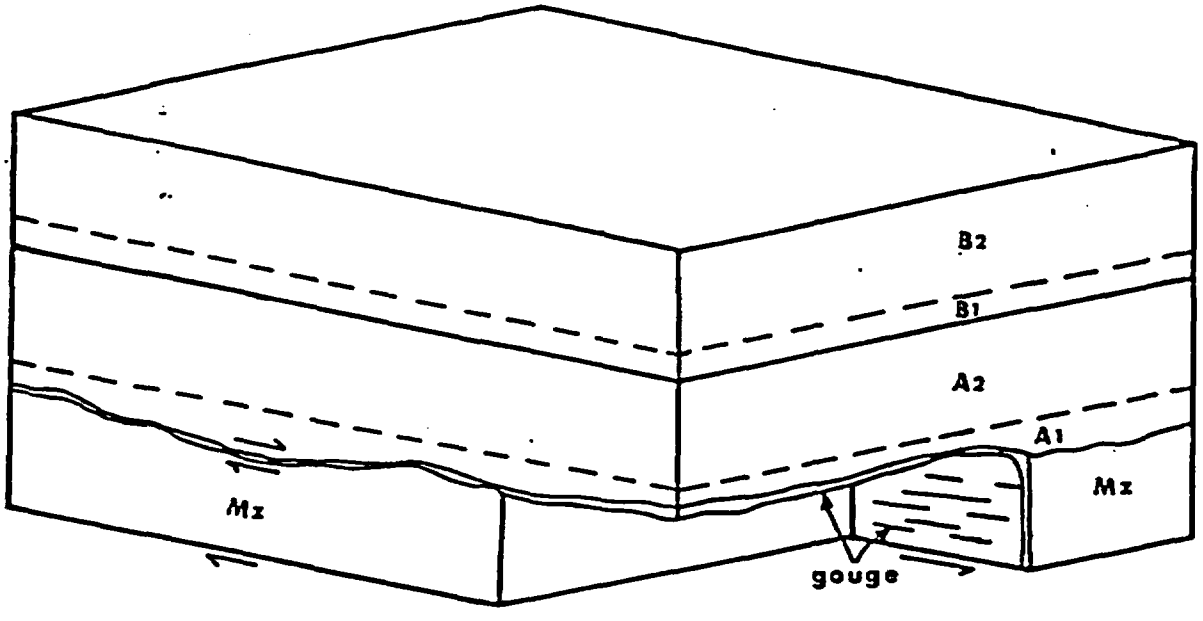


Figure 28b - Lateral movement on strike-slip fault migrates toward surface but is taken up on horizontal surface occupied by nonwelded ash-flow at base of the section (after Hardyman, 1978).

Molinari, 1984

simplified version of the model developed by Hardyman (1978) to explain the origin of detachment faults observed in the Gabbs Valley and Gillis Ranges. One major difference between the two models is the difference in structural style within the detached blocks. In the Gabbs Valley and Gillis Ranges, the lavas and welded tuffs exhibit brittle, tensional deformation. Numerous listric normal faults, with trends subparallel to the strike-slip faults, form in the detached blocks rotating and extending the Tertiary strata. In Stewart Valley, the sediments of the Esmeralda Formation exhibit a more plastic deformation with folds being the dominant structures in the detached blocks. These differences in structural style are most likely a result of divergent versus convergent wrenching in addition to the different lithologies. The fact that the folds are confined to the Esmeralda Formation and do not affect the lavas and volcanic breccias along the northern segment of the SMCFZ attests to a lithologic influence on fold formation.

Characteristic Earthquake for the SMCFZ

As previously noted, detailed studies of several faults and the presence of multiple fault scarps along many of the range-front faults in the Basin and Range Province and elsewhere suggest individual faults experience "characteristic" earthquakes with relatively consistent recurrence intervals, surface fault rupture lengths, and displacements. There has obviously been numerous earthquakes generated by slip on the SMCFZ during the late Pleistocene and Holocene with magnitudes

large enough (Ms >5.5 to 6.0) to produce surface faulting. It was not possible within the scope of this study to perform the detailed trenching and soil stratigraphic work needed to determine the late Pleistocene and Holocene fault rupture history of the SMCFZ but some general conclusions regarding a "characteristic" earthquake can be inferred based on comparisons of the 1932 surface faults with the tectonic geomorphology and structural features in the study area, and wrench fault models.

The 1932 Cedar Mountain earthquake produced new surface faults and reactivated pre-existing faults in the southern portion of the SMCFZ. In addition, faults which are clearly of late Pleistocene age and have the same trend as the 1932 faults were not reactivated. Topographic profiles and total scarp heights on the individual faults of the southern SMCFZ suggest a variable number of surface faulting events and amount of displacement per event. The irregular nature of surface faulting is suggestive of earthquakes with varying magnitudes, fault rupture patterns, and displacements.

An alternative is that large magnitude earthquakes similar to the 1932 event occur on the SMCFZ with relatively regular recurrence intervals; but due to the lack of a throughgoing wrench fault at the surface, subsurface displacement is not completely reflected at the surface. If en echelon faults of the southern SMCFZ are indicative of the "peak structure" stage of evolution, then measured 1932

surface displacements would correspond to 50% or less of the total fault slip at depth and previous surface faulting events would reflect even less of the total slip (Tchalenko, 1970). In addition, Wilcox and others (1973) noted that numerous fault blocks form early and once separated by faulting, tend to deform independently. This could account for the variation in the apparent number and size of prehistoric surface faulting events on the south segment of the SMCFZ, and lack of 1932 surface faulting along some pre-existing scarps. Folding and/or detachment would also affect the surface fault rupture pattern and displacements. If this is the case on the SMCFZ, then the number of surface faulting events suggested by the scarp profiles may be accurate but using the scarp heights to determine the number and magnitude of prehistoric surface faulting events will probably yield inaccurate results.

Summary

The structural pattern, timing, and degree of development of the SMCFZ is indicative of a right-lateral wrench fault in the initial stage of the early phase of development. This interpretation is supported by the shear box experiments of Tchalenko (1970) and Wilcox and others (1973) which showed that: 1) Reidel shears form in the initial (peak structure) stage of the early phase of wrench-zone deformation; 2) the Reidel shears are either rotated towards the slip direction and interconnect or are cut by new faults parallel to the

slip direction during formation of a main, continuous wrench fault in the last (residual structure) stage of the early phase of deformation; 3) en echelon folds form very early in the development of a wrench zone and may be crosscut later by faults, and 4) both convergence and divergence (and the associated structures) commonly develop locally along the wrench zone. In addition to the fault-fold pattern, evidence for a major component of right-slip includes the low-angle rake of striae on exposed fault planes in the northern section, fault-related geomorphology, measured lateral displacements associated with the 1932 earthquake, and a composite focal mechanism of microearthquakes in the area (Figure 29).

The number and size of folds and amount of vertical displacement on the faults in the north and central sections of the area suggest a greater amount of deformation on the north segment than the south segment of the SMCFZ. The lack of evidence for surface faulting in 1932 and sparse youthful, fault-related geomorphology along the north segment suggests it has been relatively inactive during the latest Pleistocene and Holocene. These relationships may be the result of late Miocene to late Pleistocene deformation in the north and central sections of the area followed by the formation of the south segment of the SMCFZ during the latest Pleistocene and Holocene. An alternative is that the SMCFZ is a continuous right-lateral wrench zone at depth and the discrepancies between the amount and timing of deformation observed at the

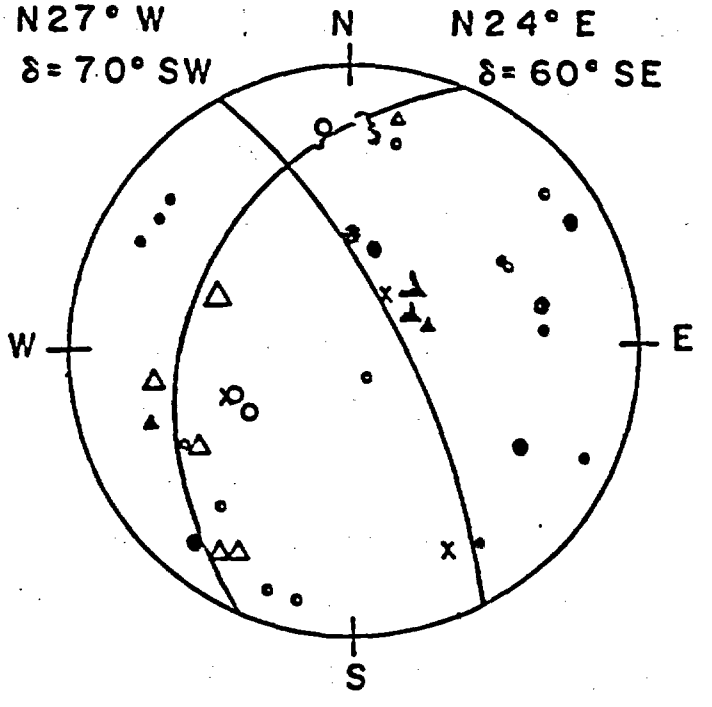


Figure 29 - Composite focal mechanism from microseismicity recorded in January and February 1969 (Gumper and Scholz, 1971) in the Stewart and Monte Cristo valleys area. The focal mechanism indicates right-oblique slip on a N27°W trending fault. This is in strong agreement with the mapped faults in the area. Solid symbols represent compressions; open symbols, dilatations; triangles and crosses, data considered to be near nodal plane. The larger symbols are the most reliable.

surface is the result of the process of developing a single throughgoing fault, folding, and detachment.

Figure 30 is a simplified map view diagram depicting the progressive development of a left-step on a right-lateral wrench fault. Convergence occurs with increasing displacement causing folding transverse to the fault trend. One of the master faults becomes inactive while the other master fault continues to form, eventually merging with the "dead" fault to form a throughgoing wrench fault. The development of the SMCFZ is considered to be analogous to this model which would explain the previously noted apparent variations in timing and rate of deformation along the SMCFZ. The SMCFZ is thought to currently be in the second or third stage of this model (see Figure 30). The lack of 1932 surface faults in Stewart Valley is considered to be due to the poorly developed nature of the newly forming fault at depth, folding, and/or detachment in the Esmeralda Formation. Even with the additional 1932 surface faulting detected during this study, the surface deformation cannot accurately reflect the amount of fault rupture that must have occurred at depth to generate the Ms = 7.2-7.3 earthquake in 1932. The distribution of earthquake epicenters throughout the area (Figure 31) and the 1932 surface faulting to the north of the SMCFZ in southern Gabbs Valley (Gianella and Callaghan, 1934a, b; and interpretation of low-sun-angle aerial photographs for this study) also supports the interpretation of a major throughgoing fault at depth.

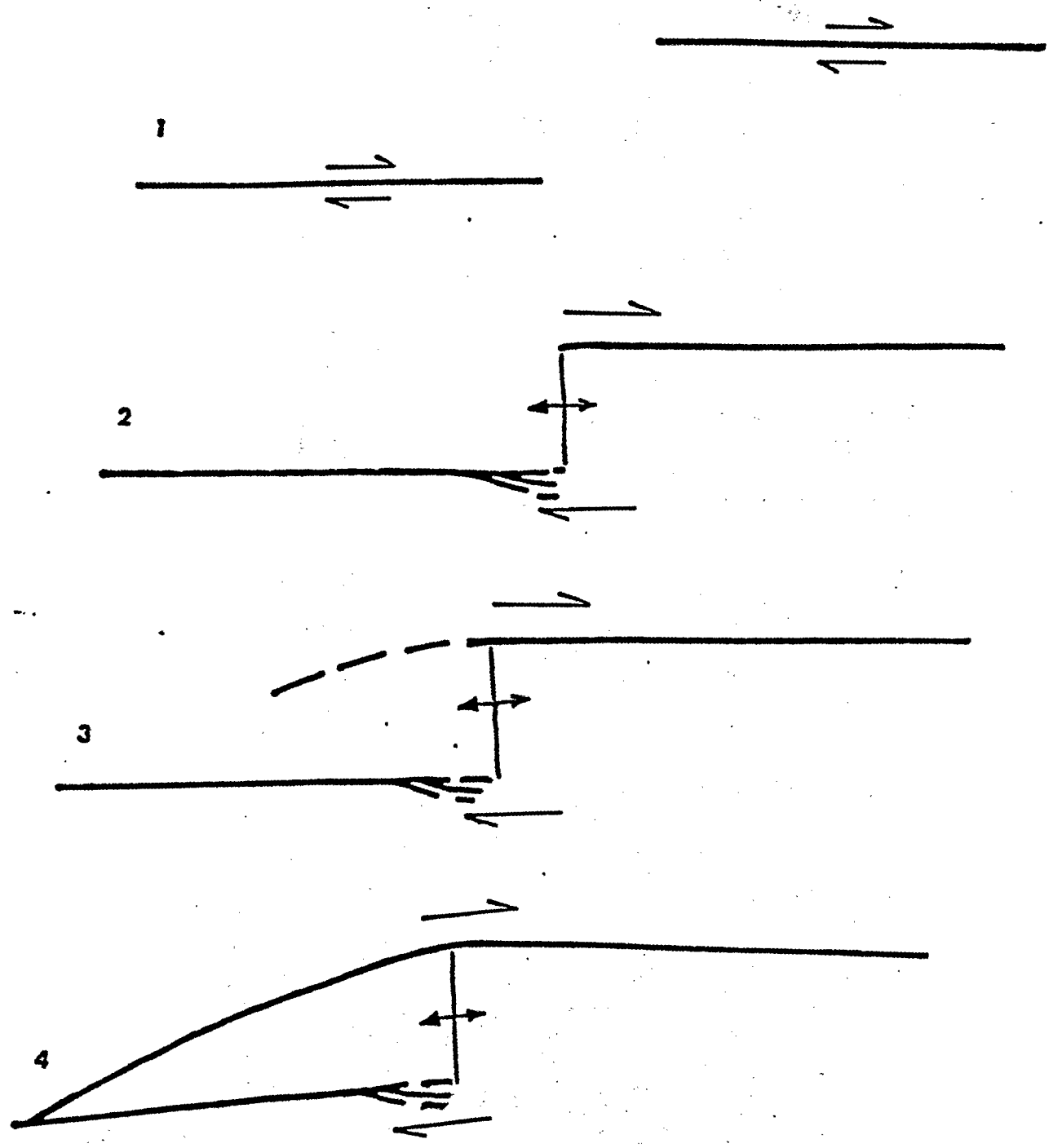


Figure 30 - Diagram depicting progressive development of fold (2) normal to fault trend and new fault splay (3 and 4) due to strong convergence at left step of a right-slip fault. Detachment may occur along new splay during early stage of development (see figure 28).

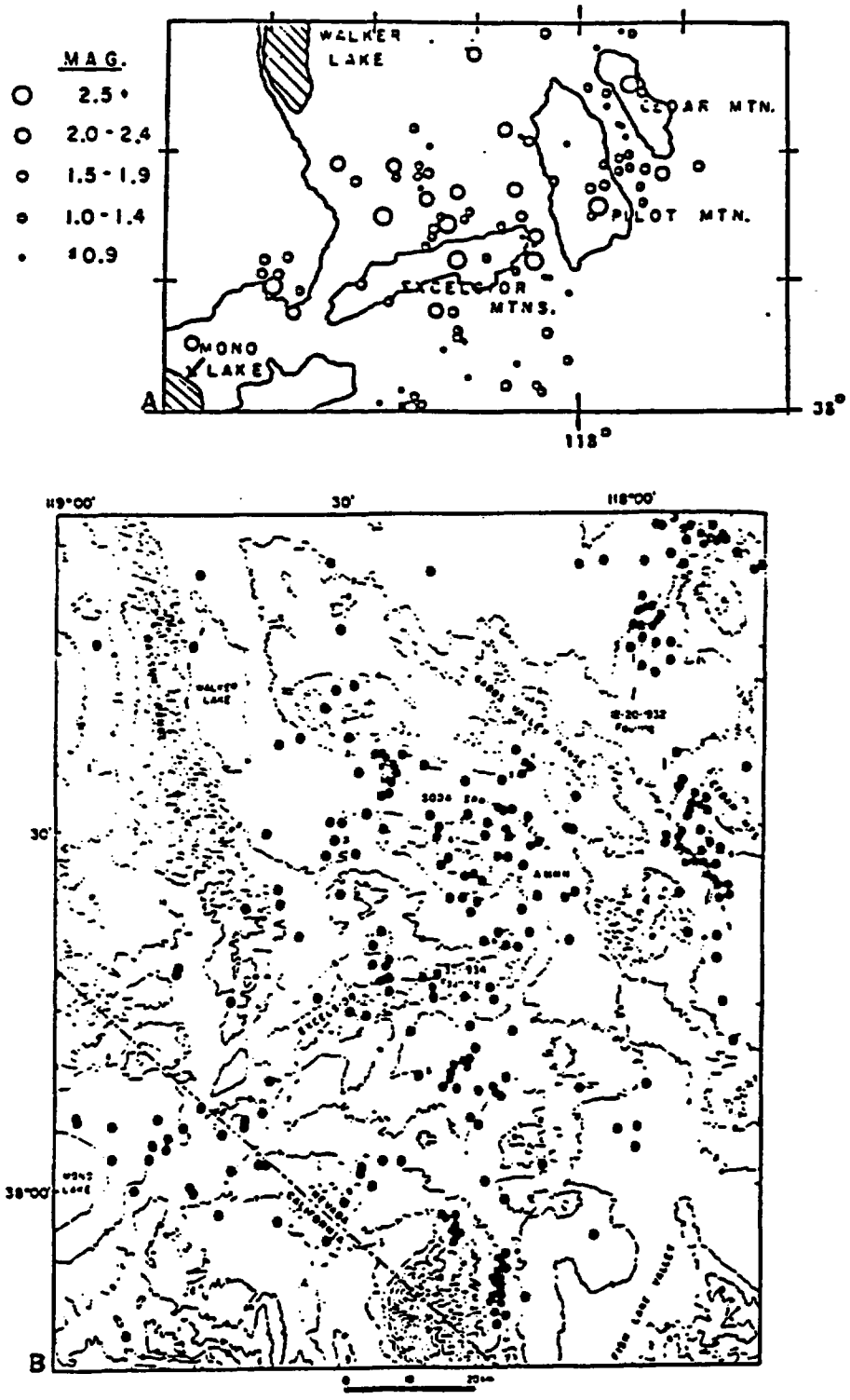


Figure 31 - Seismicity maps for the periods of a) January and February 1969 (Gumper and Scholz, 1971) and b) 1970 to 1972 (Ryall and Priestley, 1975) for west-central Nevada showing a good correlation between the location of the epicenters with mapped faults in Stewart and Monte Cristo Valleys.

The overall structural pattern of Stewart and Monte Cristo valleys is a small scale analog of the large scale tectonic pattern of the western United States and is consistent with the tectonic models of Wright (1976) and Hill (1982). These models emphasize the relatively consistent relationship of structural trends and style of deformation resulting from north-south compressive stress: normal faults strike northerly, reverse faults and associated folds strike easterly, and strike-slip faults have a conjugate relationship of northwest striking right-slip faults and northeast striking left-slip faults.

RELATIONSHIP OF THE SMCFZ TO THE LATE CENOZOIC STRUCTURAL GEOLOGY OF THE CENTRAL WALKER LANE REGION

The late Cenozoic tectonics of the central Walker Lane are characterized by major northwest trending right-slip faults in the Gabbs Valley Range, Gillis Range, and Pilot Mountains and east-to-northeast trending left-slip faults in the Excelsior Mountains, Garfield Hills, and Candelaria Hills (Plate 7). The onset of extensional tectonics in the central Walker Lane occurred at least 23-24 m.y. ago and slip directions appear to have been consistent up to the present, indicating a uniform stress regime during this time (Speed and Cogbill, 1979a; Ekren and others, 1980). This is anomalous with respect to data from throughout the rest of the northern Basin and Range which indicates a 45° rotation in the least principal stress orientation from WSW-ENE to WNW-ESE approximately 10 m.y. ago (Zoback and others, 1981). Initiation of the SMCFZ is post-11.0 m.y.b.p. and may have coincided with this inferred reorientation of the stress regime. The 8° angular unconformity between the Mount Ferguson breccias (Tlfb) and younger sedimentary rocks of the Esmeralda Formation on the east flank of the Gabbs Valley Range (Figure 12) is probably a result of uplift on the Bettles Well fault and predates the inception of the SMCFZ.

The folding in the Esmeralda Formation in Stewart Valley is unique in that it is the only documented case of major late Cenozoic wrench folds in western Nevada. Folds also

occur in mid-to-late Miocene fluvial-lacustrine sediments in the Silver Peak-Lone Mountain-Blair Junction (Robinson, 1964); Moiola, 1969; Moore, 1981) and Pine Grove Hills-Coal Valley (Gilbert and Reynolds, 1973) areas of western Nevada. Folds are not as prevalent in these areas and are formed due to drag or warping associated with movement on normal faults.

Right-slip on a N30°W fault trend indicates the SMCFZ is related to the system of left-stepping, en echelon right-slip faults in the Gillis and Gabbs Valley Ranges (Plate 7). These faults are well developed throughgoing wrench faults with a cumulative right-lateral separation of at least 32 km (20 miles) and maybe as much as 48 km (30 miles) (Hardyman and others, 1975; Hardyman, 1978; Ekren and others, 1980). The early stage of development and relatively young age of the SMCFZ suggests a southeastward expansion of this fault system.

The east-west trending Pancake Range lineament of Ekren and others (1976) crosses the study area in northern Monte Cristo Valley. The lineament is expressed in the area by topographic breaks between the north and south portions of the Cedar Mountains and between the Gabbs Valley Range and Pilot Mountains, and the east-west trending outcrop pattern of Mesozoic and Tertiary rocks in the area. The detailed mapping of this study shows the post Esmeralda Formation (mid-to-late Miocene) structural trend is northwest-southeast, not east-west, and the SMCFZ is continuous across the inferred trace of the lineament. If the lineament has any structural

significance it is related to Mesozoic structures, which have east-west trends in the area.

The topographic boundary between the central Walker Lane and the typical Basin and Range physiography is located south of the Monte Christo Mountains, Paradise Range, Shoshone Range, Toiyabe Range, and San Antonio Mountains (Plate 7) as noted by Gianella and Callaghan (1934a). The Walker Lane is more commonly considered a structural zone characterized by northwest trending right-slip faults and conjugate northeast trending left-slip faults (i.e., Shawe, 1965; Hardyman, 1978; Slemmons and others, 1979; Rowan and Wetlaufer, 1981). The boundary between the Walker Lane and typical Basin and Range, as related to late Cenozoic structural trends, is marked by the transition from the northwest trending right-slip faults of the Walker Lane and north-northeast trending basin-range normal faults (Plate 7).

The structural boundary diverges from the topographic boundary in northern Stewart Valley and the SMCFZ is the southeasternmost structure exhibiting Walker Lane structural characteristics. The range-front fault on the west flank of the Cedar Mountains does not have the characteristic north-northeast trend of basin-range faults but the stratigraphic relationships along the fault suggest a significant normal component of slip and the sinuous trace of the fault limits the amount of right-slip, unless significant reverse faulting has occurred at major bends in the fault trace. Minor folds and normal faults in southernmost Ione Valley

have both northwest and northerly trends and the faults do not exhibit any lateral displacement (Henderson, 1962). The normal faults in the southern Cedar Mountains have a basin-range trend and juxtapose Jurassic and Triassic rocks with Tertiary ash-flow tuffs (Albers and Stewart, 1972). No faults have been mapped in the Royston Hills, Cirac Valley, or the northwest arm of Big Smoky Valley between the Royston Hills and the Toiyabe Range (Kleinhampl and Ziony, 1967). The northern portion of the Royston Hills have a northwest trend and could conceivably be bounded by a fault on the northeast but no distinct fault was observed on high quality, 1:250,000 scale color Skylab aerial photography of this area.

The SMCFZ is located on the east margin of the elongate, northwest trending aeromagnetic anomalies associated with the major wrench faults in the Gabbs Valley Range, Gillis Range, and Pilot Mountains (U.S.G.S., 1971a, b). While not as distinct, a similar relationship is indicated on Bouguer gravity maps of the area (Healy and others, 1980; Healy and others, 1981). These regional geophysical and structural relationships support the interpretation of the SMCFZ as the southeast structural boundary of the central Walker Lane and suggest the wedge-shaped area between the SMCFZ and the topographic boundary is a broad transition zone between the two structural provinces. This transition zone has apparently been less tectonically active since the late Miocene than the Walker Lane and basin-range normal faults to the north.

The northwest trending faults of the central Walker Lane

are not known to occur south of the Pilot Mountains or Monte Cristo Valley. The northern portion of the Furnace Creek fault zone is located south of the Candelaria Hills in Fish Lake Valley and is considered to be the southerly continuation of the Walker Lane structural zone. Faults in the Monte Cristo Range trend both northwest and east-northeast, are relatively short and discontinuous, and neither fault set displaces the other. The overall pattern is somewhat arcuate and many of the faults may be related to the formation of a pre-Gilbert andesite caldera (J. Stewart, written and oral communication, 1983). South of the Monte Cristo Range, a relatively continuous series of right-stepping, en echelon, east-northeast trending normal and normal-left-slip faults extend from the Volcanic Hills on the west to Lone Mountain on the east. At Lone Mountain the fault trend changes to the typical north-northeast trend of basin-range normal faults and continues along the west flank of the San Antonio Mountains (Plate 7).

Albers and Stewart (1972, p. 42-44) suggested the Walker Lane may have been displaced in a right-lateral sense along the series of east-northeast trending, right-stepping fault zones located south of the Monte Cristo Range. The east-northeast faults coincide with the trace of the Warm Springs lineament of Ekren and others (1976, p. 6) who argued against the displacement of the Walker Lane along this zone.

This author's interpretation of the existing geologic data in this area is that if the Walker Lane has been

displaced in a right-lateral sense, it occurred prior to the onset of late Oligocene volcanism in the central Walker Lane since the post-Oligocene faults in the Volcanic Hills-Silver Peak Range-Lone Mountain area are primarily normal faults with a component of left-slip. The late Cenozoic distribution and style of faulting in the southwest portion of the central Walker Lane is a result of a large scale, en echelon right-step in a right-slip fault system. The conjugate east-northeast trending, late Cenozoic left-normal-slip faults of the Garfield Hills-Excelsior Mountains-Candelaria Hills area formed due to extension that resulted from the "pull-apart" that occurred at the right-step from the Furnace Creek fault zone to the series of northwest trending, right-lateral strike-slip faults in the Gabbs Valley Range and Pilot Mountains. This interpretation is consistent with models of pull-apart basins (Crowell, 1974; Mann and others, 1983). Since the right-lateral movement in the central Walker Lane is distributed over several faults, a single topographic and structural basin that characteristically occurs did not form and the resultant extension was distributed over a large area and numerous faults. The distribution of Pliocene basalt throughout the area (Stewart and others, 1982) and formation of small, local pull-apart basins such as the Candelaria trough (Speed and Cogbill, 1979c) and Rhodes Salt Marsh also support this interpretation.

LATE CENOZOIC GEOLOGIC HISTORY OF STEWART AND MONTE CRISTO VALLEYS

During the late Oligocene and early Miocene in the Stewart Valley - Monte Cristo Valley area, emplacement of silicic ash-flow tuffs, derived from sources in the present day northern Gillis Range (Ekren and others, 1980) and Monte Cristo Range (J. Stewart, oral communication, 1983), was followed by widespread andesitic volcanism. Minor tectonism in the Cedar Mountain area prior to deposition of the Esmeralda Formation is marked by the angular unconformity between the andesitic lavas and Esmeralda Formation north of Stewart Spring. Deposition of the fluvio-lacustrine sediments of the mid-to-late Miocene Esmeralda Formation marks the inception of basin formation in the study area, approximately 15.5 m.y. b.p.. The interfingering of the Esmeralda Formation with the volcanic debris-flow breccias of the lavas of Mount Ferguson on the west side of Stewart Valley, and abundant petrified wood preserved at the contact where the breccias overlie the Esmeralda Formation, indicate the western margin of the depositional basin was approximately coincident with the present topographic boundary of the valley. Previous work on the north and east flanks of the northern Cedar Mountains (Henderson, 1962; Mawby, 1965) indicate that low hills existed in the area of the present day northern Cedar Mountains and separated the Ione Valley and Stewart Valley basins during the deposition of the Esmeralda Formation. These hills were not considered to be a major source area for the sediments. The relative

abundance of coarse clastic material in easternmost Stewart Valley, south of Stewart Spring, and the presence of clasts in the Esmeralda Formation of ash-flow tuff which outcrops just east of the range-front fault in this area, suggest the paleo-Cedar Mountains may have been shedding material to the west and the eastern margin of the Stewart Valley basin was similar to the present day valley. Paleocurrent work in this area is needed to verify this interpretation.

Minor syn-Esmeralda uplift of the Cedar Mountains is suggested by an angular unconformity within the formation in southern Ione Valley (Henderson, 1962), and along the western margin of the Stewart Valley basin deformation is indicated by the slight angular unconformity shown in Figure 12. No angular discordance was noted in the Esmeralda Formation in Stewart Valley thus inception of the SMCFZ post-dates the youngest Esmeralda sediments (less than 11 m.y.). Deformation on the SMCFZ is represented by the folding and faulting in Stewart and Monte Cristo valleys which has continued to the present. Continued uplift of the Pilot Mountains, Cedar Mountains, and southern Gabbs Valley Range was accompanied by subsidence of central Monte Cristo Valley.

Exterior drainage of Stewart Valley into Gabbs Valley during the late Quaternary has formed the series of pediment remnants which dominate the present topography. The cyclic pattern of widespread planation of the Esmeralda Formation followed by periods of downcutting is the result of climatic changes, possibly in conjunction with a relative decrease of

the base level in Gabbs Valley. The mid(?)-to-late Holocene was primarily a period of deposition but erosional processes are dominant at the present time. A similar series of depositional events are preserved in Monte Cristo Valley. The dominance of erosional processes in Stewart Valley and depositional processes in Monte Cristo Valley is primarily a function of the difference between semi-bolsons and bolsons (open and closed basins) rather than differences in tectonic or climatic history of the two valleys.

CONCLUSIONS

The geologic mapping completed during this study is the first detailed mapping done in Stewart and Monte Cristo valleys and extends previous work completed to the west in the Gabbs Valley Range and Pilot Mountains (Ekren and Byers, 1978; Oldow, 1981, 1983 written communication). Late Cenozoic deformation in the area is subsequent to deposition of the mid-to-late Miocene Esmeralda Formation (post-11.0 m.y. b.p.) and has continued until the present as evidenced by the 1932 Cedar Mountain earthquake and microseismicity.

This study has delineated and defined the Stewart-Monte Cristo fault zone (SMCFZ); a N30°W trending zone of discontinuous right-normal-slip faults that extends southward along the east flank of the Gabbs Valley Range to the southwest edge of Kibby Flat in Monte Cristo Valley. Total displacement across the northern segment of the fault zone is at least several hundred meters. Fault-related geomorphology of the southern segment provides evidence for at least three, and possibly five or six surface faulting events during the latest Pleistocene and Holocene. This segment was reactivated in 1932 with scarp heights of up to 1.4 meters and 1 to 2 meters of right-lateral separation of antecedent stream channels. In addition to refining the mapped traces of the 1932 surface faults mapped by Gianella and Callaghan (1934a, b), three additional major 1932 faults were delineated (faults D, E, and H on Plate 6).

Numerous north to northwest trending, gentle to open folds in the Esmeralda Formation are located east of the SMCFZ. These folds are interpreted to be wrench folds genetically related with right-slip on the SMCFZ. The geometric relationship of the folds to the SMCFZ is similar to other well-documented wrench faults and is consistent with laboratory models of wrench faults. This is the first documented occurrence of widespread wrench-style folding in late Cenozoic strata within the Basin and Range.

Detachment faulting similar to that documented in the nearby Gabbs Valley and Gillis ranges may be occurring at the base and/or within the Esmeralda Formation. Evidence for detachment is the lack of a throughgoing surface fault of sufficient length to generate the 1932 (M = 7.2-7.3) earthquake. The widespread distribution of 1932 surface faults and microseismicity supports the interpretation of a throughgoing fault at depth. In addition, detachment would enhance fold formation and may have generated anomalous fold geometry in Stewart Valley.

The SMCFZ is related to the system of major left-stepping, right-slip faults located to the west in the Gabbs Valley and Gills ranges and Pilot Mountains. It is probably the southeasternmost fault of the central Walker Lane structural zone. The relatively young age of inception of the SMCFZ, and relatively high rate of late Pleistocene and Holocene deformation suggests a southeastward expansion of the central Walker Lane.

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