CALIFORNIA STATE UNIVERSITY, NORTHRIDGE

# SUPERPOSED MESOZOIC DEFORMATIONS, SOUTHEASTERN INYO MOUNTAINS, CALIFORNIA

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

Geology

by

Michael Robert Werner

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The Thesis of Michael Robert Werner is approved:

Dr. Perry L. Ehlig (CSULA)

Dr. Richard H. Miller (SDSU)

Dr. George C. Dunne (CSUN)

California State University, Northridge

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ABSTRACT

SUPERPOSED MESOZOIC DEFORMATIONS,

SOUTHEASTERN INYO MOUNTAINS, CALIFORNIA

by

Michael Robert Werner

Master of Science in Geology

Mapping in the Nelson Range and the adjacent southeastern Inyo Mountains shows that this area has experienced at least two episodes of early Mesozoic deformation which produced two sets of folds in Mississippian through Permian rocks. A later Mesozoic event may be represented by faults in this area.

The older fold set ($F_1$) consists of generally northeast-trending, upright to southeast-overturned folds. A younger fold set ($F_2$) is superimposed upon and locally refolds $F_1$ folds. $F_2$ consists of generally northwest-trending, upright to northeast-overturned folds with steeply southwest-dipping axial plane cleavage. Truncation of some $F_2$ folds by the Hunter Mountain batholith in the Nelson Range indicates that both fold sets are pre-Middle Jurassic in age. The style of $F_2$ folds along portions of the intrusive contact suggests that emplace-

ment of the batholith locally appressed and deflected pre-existing folds.

Based on their orientation, sense of vergence and relative age, $F_1$ folds are believed to be temporally and structurally related to the Last Chance thrust system (Middle (?) Triassic to Early Jurassic). $F_2$ folds are correlated with other northwest-trending folds of pre-Middle Jurassic age which occur throughout the ranges immediately east of the Sierra Nevada.

Some northwest-striking faults in this area apparently are intruded by dikes of the Independence swarm and thus may be pre-Late Jurassic in age. If so, these faults may be related to a broad, northwest-trending zone of fractures and left-slip faults of Middle to Late Jurassic age which trends obliquely across this region.

## INTRODUCTION

The Nelson Range trends northwest across the southwest corner of the Ubehebe Peak quadrangle in Inyo County, California (fig. 1). Paleozoic rocks exposed along the southwest flank of the range and in the eastern foothills of the southern Inyo Mountains consist of over 2100 m (6880 ft) of Mississippian through Permian marine stata. These rocks have been intruded in the Nelson Range by the Hunter Mountain batholith of late Early Jurassic age.

Initial regional mapping of the Ubehebe Peak quadrangle and vicinity by McAllister (1952, 1956) demonstrated the predominance of north-and northwest-trending folds of Mesozoic age in the Paleozoic rocks of the Nelson Range and southern Inyo Mountains. Locally developed east-trending folds, as well as northwest-trending folds roughly parallel to the intrusive contact in the Nelson Range, were believed to be the result of forceful batholithic emplacement (McAllister, 1956). Northwest-trending folds were also reported in the Darwin quadrangle to the south (Hall and MacKevett, 1962) and in the Cerro Gordo district to the west (Merriam, 1963), and these also were believed to have formed during emplacement of nearby plutons.

Recent geological investigations in areas adjoining the Nelson Range indicate that many folds pre-date plutonism and that the ranges immediately east of the Sierra Nevada contain one of the most complete records of Mesozoic deformation in eastern California (Dunne and Gulliver, 1976; Dunne and others, 1978). In the Talc City Hills, Gulliver (1971, 1976a, 1976b) reports three phases of Mesozoic folding, at least two of which are genetically related to faults of regional

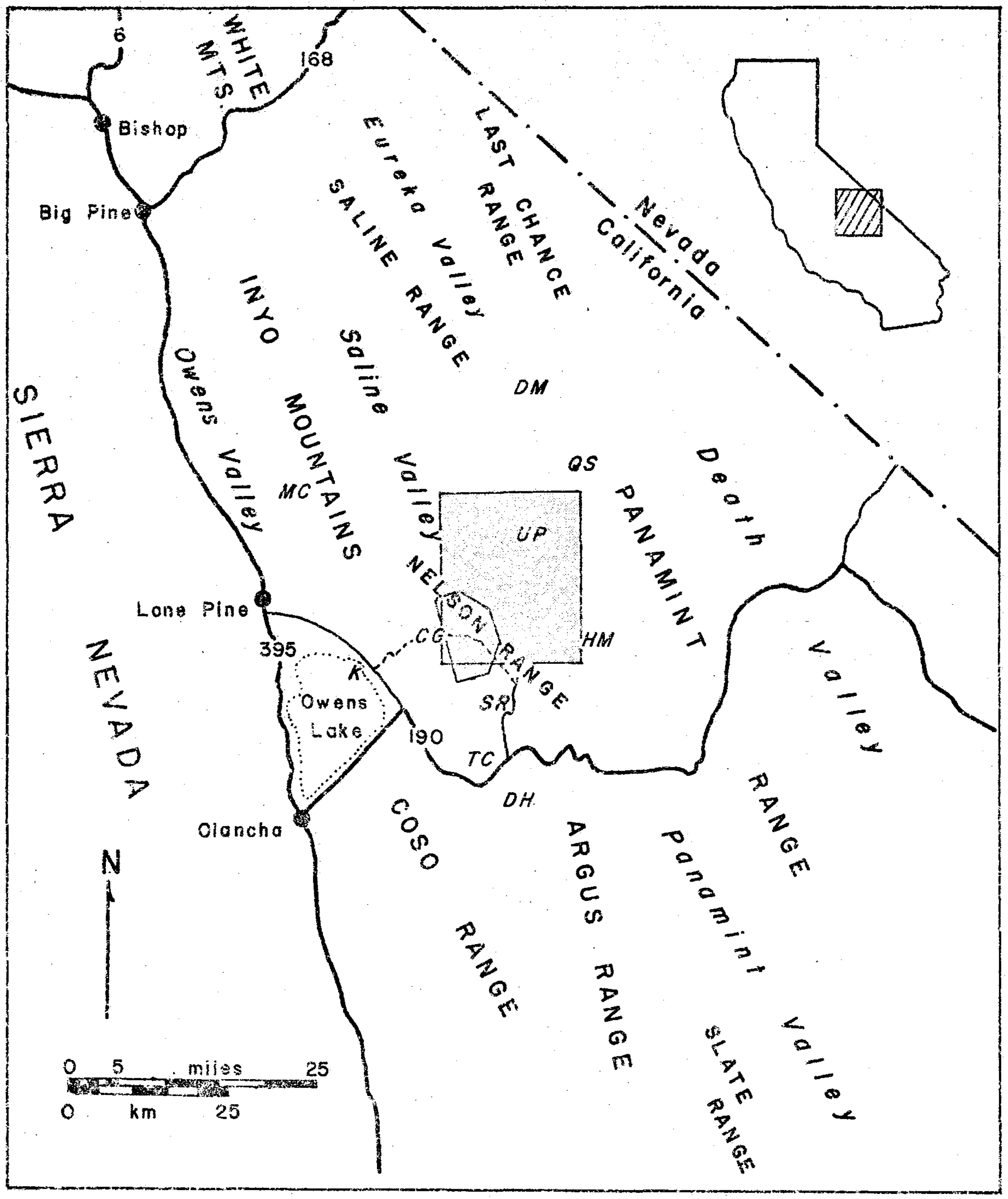



Figure 1. Index map to eastern California. Ubehebe Peak quadrangle and study area shown in gray. Abbreviations are as follows: CG, Cerro Gordo; DH, Darwin Hills; DM, Dry Mountain; K, Keeler; MC, Mazourka Canyon; QS, Quartz Spring; SR, Santa Rosa Hills; TC, Talc City Hills; UP, Ubehebe Peak; HM, Hunter Mountain.

extent. In the Darwin Hills, Dunne and Gulliver (1976) recognized four deformational phases which span the Mesozoic and are characterized by folds and faults of distinctive style and geometry. Elayer (1974) reports four phases of folding near Cerro Gordo in the southern Inyo Mountains, although, subsequent work by Mora (1978) in this same area suggests a somewhat simpler structural history.

The purpose of the present investigation was to determine whether the Nelson Range and nearby eastern slopes of the southern Inyo Mountains share a similar history of multiple superposed Mesozoic deformation as reported in adjoining areas. Preliminary work showed that this area bears the imprint of at least two phases of early Mesozoic deformation (Werner, 1978). These deformational phases are believed to be genetically and temporally correlated with major structural features which define a portion of an early Mesozoic orogenic belt in eastern California (Burchfiel and others, 1970).

## METHOD

The study area (fig. 1) which consists of approximately 100 sq km, was mapped during 70 days of field work between June, 1976 and September, 1977. Field data were plotted directly on portions of the Ubehebe Peak, Darwin, and New York Butte 15' topographic quadrangles which had been enlarged to a scale of 1:15,625 (approx. 4 inches to 1 mile). Field investigation was augmented by the use of color aerial photographs (scale 1:20,000) taken in December, 1975 for the Bureau of Land Management.

Preparation and contouring of stereonet diagrams was aided by the use of the California State University, Northridge (CSUN) CDC 3170 computer using the FORTRAN program PETROFAB (Werner, 1977).

# ROCK UNITS

## PALEOZOIC STRATIGRAPHY

Paleozoic sedimentary rocks which crop out within the study area consist of over 2100 m (6880 ft) of marine limestone, shale, siltstone, sandstone, and conglomerate which range in age from Mississippian to Permian. Mississippian units are primarily fine-grained, siliceous clastics and include the Perdido Formation and Rest Spring Shale. Pennsylvanian and Permian strata, originally mapped as Bird Spring Formation by McAllister (1956), are assigned to the Keeler Canyon and Owens Valley Formations and comprise a monotonous sequence of interbedded limestone and shale.

This investigation indicates that rapid lateral variations in lithology characterize several portions of the late Paleozoic section. Field evidence suggests that these variations can be attributed to local unconformities, possibly due to submarine erosion, as well as facies changes associated with the transition from shelf carbonates to foreland basin deposition which characterized this region during the late Paleozoic (Stevens and others, in preparation).

Figure 2 shows composite stratigraphic columns for both the northern and southern halves of the study area. These stratigraphic columns were compiled from field notes and cross sections. Lithologies are much generalized and thicknesses are approximate. Stratigraphic data presented here are in general agreement with data compiled by Duane Cavit and Robert Husk, California State University, San Jose, who are conducting detailed stratigraphic investigations in portions of the study area.

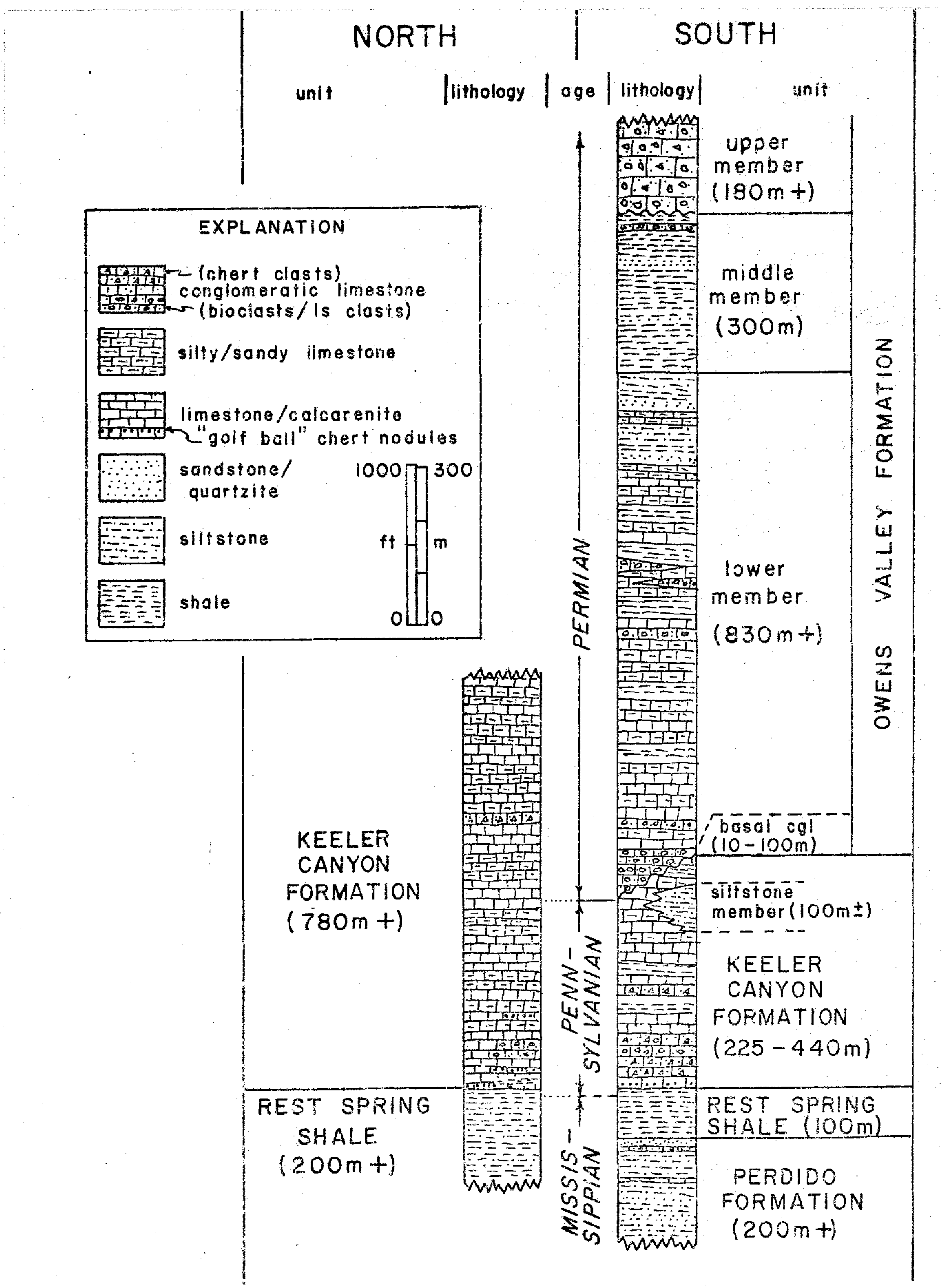



Figure 2. Generalized stratigraphic columns for the Nelson Range (north) and southeastern Inyo Mountains (south).

## Perdido Formation

The name Perdido Formation was first applied by McAllister (1952) to a laterally diverse sequence of siltstone, sandstone, limestone, and conglomerate exposed in Perdido Canyon in the Quartz Spring area (fig. 1). Exposures of Perdido Formation in the study area are confined to two locations in the foothills east and northeast of Conglomerate Mesa. At both locations the Perdido seems to be in depositional contact with the overlying Rest Spring Shale, but the base of each section is faulted.

The thickest exposure of Perdido Formation (approximately 200 m (650 ft)) occurs due east of Conglomerate Mesa. The lowest exposed portions of the section include dark gray, cherty limestone which is similar to and may be gradational with the underlying Tin Mountain Limestone (McAllister, 1952). Most of the remainder of the section is composed of massive, thoroughly fractured, light- to medium-brown siltstone and quartzite with several interbeds of gray limestone, pelmetazoan conglomerate, and black, cherty siltstone. The upper 22 m (72 ft) of the Perdido consists of light-gray, fine-grained, laminated, calcareous sandstone. The more northerly exposure of the Perdido Formation is of limited extent and consists of only the uppermost light gray sandstone.

An Early to Late Mississippian age is reported by McAllister (1952) for the Perdido Formation at its type section. This age range agrees with the findings of Merriam (1963) and Elayer (1974) in the New York Butte quadrangle immediately to the west.

Rest Spring Shale

The Rest Spring Shale was first described by McAllister (1952) in the Quartz Spring area. McAllister (1955, 1956) subsequently applied this name to all dark-colored, shaly Mississippian strata in the Ubehebe Peak quadrangle. Rest Spring Shale has also been recognized in the Darwin quadrangle to the south (Hall and MacKevett, 1962), and the Independence (Ross, 1965) and Dry Mountain (Burchfiel, 1969) quadrangles to the north. In the Cerro Gordo district to the west, Merriam (1963) referred to this unit as the Chainman Shale inferring lighologic correlation with the Chainman Shale near Ely, Nevada (Spencer, 1917).

Stevens and others (in preparation) propose that the Rest Spring Shale in Mazourka Canyon in the Independence quadrangle consists of two members. The lower member is equivalent to the Chainman Shale (Stevens and Ridley, 1974). The upper member, which consists of blocky-weathering andalusite hornfels, is equivalent to the Hamilton Canyon Formation (Langenheim, 1962; Stevens and Ridley, 1974). Elayer (1974) proposed that andalusite hornfels exposed in lower San Lucas Canyon within this study area should also be called Hamilton Canyon Formation because they are similar to this formation in Mazourka Canyon. However, extensive metamorphism of adjacent carbonate rocks of the Keeler Canyon Formation in San Lucas Canyon suggests that these hornfels are the result of thermal metamorphism of shale by the Hunter Mountain batholith and they probably do not represent a distinct stratigraphic unit. Therefore, these hornfels are referred to the more general classification of Rest Spring Shale.

West of Lee Flat, approximately 100 m (330 ft) of Rest Spring Shale overlies Perdido Formation and is overlain by Keeler Canyon Formation with sharp depositional contacts. The Rest Spring section consists of olive gray and black shale with minor thin beds of dark reddish brown siltstone and very fine sandstone. Locally there are small discoidal concretions 1-3 cm in diameter which are aligned parallel to bedding. Tracks and burrows are common on shaly partings in some horizons.

In San Lucas Canyon and the Nelson Range, the Rest Spring Shale consists of massive gray and brown siltstone and shale commonly metamorphosed to hornfels which are characterized by the presence of randomly oriented andalusite crystals. Locally, relict laminations or sandy lenses are seen which are similar to those in the Rest Spring Shale in the south half of the study area; however, shaly partings are rare. The contact with the overlying Keeler Canyon Formation in San Lucas Canyon is relatively sharp. To the northeast the contact becomes gradational and is characterized by interbedded shale and limestone. Near the crest of the Nelson Range these interbeds are well exposed in an overturned section, and the contact has been placed at the lowest stratigraphic occurrence of mappable limestone. Therefore, some Rest Spring-like shale is included in the lower 150 m of the Keeler Canyon Formation. The gradational nature of the upper Rest Spring contact has also been reported near Cerro Gordo (Merriam, 1963) and in the Dry Mountain quadrangle to the north (Burchfiel, 1969).

The base of the Rest Spring Shale is not exposed in the north half of the study area, but cross sections indicate that its thickness exceeds 200 m (650 ft) in San Lucas Canyon. The apparent rapid thinning of the Rest Spring Shale from north to south (fig. 2) is consistent with interpretations that this region occupied a transitional zone between the Antler foreland basin to the north and shelf deposition to the south (Stevens and others, in preparation).

McAllister (1952) reported a tentative age of Early Pennsylvanian for the Rest Spring Shale based largely on stratigraphic position. Poole and Sandberg (1977, fig. 2a) indicate the Rest Spring Shale is entirely Mississippian in age at its type section. However, Dunne and Miller (personal communication, 1978) have recovered Early Pennsylvanian conodonts from the Rest Spring Shale in the Darwin Hills. Therefore, the Rest Spring Shale ranges in age from Late Mississippian to Early Pennsylvanian.

Keeler Canyon Formation

The name Keeler Canyon Formation was first applied by Merriam and Hall (1957) to limestone and interbedded shale of Pennsylvanian and Permian age which is widely exposed in the Inyo Mountains and vicinity. Originally mapped as Bird Spring Formation by McAllister (1956) in the Ubehebe Peak quadrangle, the Keeler Canyon Formation comprises over 60 percent of Paleozoic outcrops in the study area.

In the south, the Keeler Canyon occurs in a discontinuous band in the foothills east of Conglomerate Mesa and west of Lee Flat. Here the contact is considered to be essentially depositional due to the concordance of bedding between the formations and the occurrence of the distinctive "golf ball" beds which characterize the basal Keeler Canyon Formation throughout the region (Merriam and Hall, 1957). Shearing and transposition locally occurs in the Keeler Canyon near the lower contact. This probably is an expression of slight differential movement along the contact during folding due to competency contrast between the Keeler Canyon limestone and the underlying Rest Spring Shale.

The contact of the Keeler Canyon Formation with the overlying Owens Valley Formation is unconformable in the south half of the study area. Though angular discordance cannot be demonstrated, the Keeler Canyon thins dramatically northward from 440 m (1440 ft) to 225 m (740 ft) over a distance of 1 km. Where thinnest, the Keeler Canyon is overlain by a massive bioclastic conglomerate assigned to the Owens Valley Formation. Farther south the contact is less well defined and has been placed at the base of a mappable limestone conglomerate which is east of and stratigraphically lower than the contact chosen by Hall and McKevett (1962). Though this conglomerate is tentatively correlated with the much thicker conglomerate to the north, the units are not continuous due to faulting.

Lithology of the Keeler Canyon Formation in the southern half of the study area is predominantely light to dark gray, thin to medium bedded silty limestone and calcarenite with thin interbeds of gray and pale-red calcareous shale. Spherical chert nodules which range between 1 and 5 cm in diameter and are referred to as "golf balls" (Merriam and Hall, 1957) characterize the lowest 10 to 15 m. Strata immediately overlying these basal beds commonly contain thin to medium, lenticular beds of granule to pebble conglomerate comprised of clasts of pelmetazoan debris, angular chert, limestone and shale in a limestone matrix. Chert clasts are more abundant to the south and are similar to chert pebble conglomerates reported in the basal Keeler Canyon Formation in the Santa Rosa Hills to the south (Hall and MacKevett, 1962). A 100 m

(330 ft) thick unit of massive, light-to moderate-brown siltstone and quartzite with very thin interbeds of gray limestone occurs in the southernmost part of the area. This unit, mapped as the siltstone member of the Keeler Canyon Formation, is missing to the north and may have been removed by erosion prior to deposition of the Owens Valley Formation.

North of Lee Flat, the Keeler Canyon Formation represents the entire exposure of Paleozoic carbonates from San Lucas Canyon eastward to the intrusive contact with the Hunter Mountain batholith. As previously described, the lower contact between the Rest Spring Shale and the Keeler Canyon becomes increasingly gradational eastward towards the crest of the Nelson Range, and no upper contact is exposed. The maximum thickness of continuous exposure is estimated to be in excess of 780 m (2560 ft).

Northern exposures of the Keeler Canyon Formation are similar to those in the south and consist of thin-to medium-bedded, light- to dark-gray limestone, silty laminated limestone, and calcarenite. Graded, cross-laminated and convoluted bedforms locally are abundant. Massive, medium-to thick-bedded lenses of limestone conglomerate occur higher in the section than they do to the south. These conglomerates are concordant with underlying and overlying beds but truncate strata laterally suggesting that they represent portions of submarine channels. "Golf balls" are present near the base of the Keeler Canyon Formation in San Lucas Canyon but are discontinuous and occur as much as 180 m above the contact with the Rest Spring Shale near the crest of the Nelson Range. Dark colored shaly interbeds similar to the underlying Rest Spring Shale occur within the lower part of the Keeler Canyon Formation throughout the Nelson Range.

Contact metamorphism caused by the Hunter Mountain batholith has altered much of the Keeler Canyon Formation in the Nelson Range to brown and white calcsilicate hornfels and marble. "Golf balls" and chert pebble clasts are commonly altered to blebs and clots of talc and actinolite-tremolite. Although some sedimentary features locally are preserved, most of the original stratigraphic detail has been obliterated. Assignment of carbonates within this calcsilicate terrain to the Keeler Canyon Formation is based on: 1) the proximity of the contact with the Rest Spring Shale, 2) the presence of "golf balls", 3) the apparent structural continuity of the carbonates above the Rest Spring Shale, and 4) the lack of any distinctive lithologic breaks up-section that would indicate the presence of any other stratigraphic units (i.e., Owens Valley Formation).

On the basis of fuslinid data, the basal "golf ball" beds of the Keeler Canyon Formation in the Darwin quadrangle are reported to be Middle Pennsylvanian (Hall and MacKevett, 1957). Fauna of probable Wolfcampanian age (E. Permian) near the northwest corner of the Darwin quadrangle (Hall and MacKevett, 1957) may actually lie within the Owens Valley Formation as defined in this paper. Keeler Canyon beds immediately below the basal conglomerate of the Owens Valley Formation are reported to be Virgilian (Late Pennsylvanian) in age (Duane Cavit, personal communication, October, 1977).

Age of the Keeler Canyon Formation in the Nelson Range has not been established directly. C. H. Stevens (personal communication, 1977) reported Late Pennsylvanian fusilinids about 60 m (200 ft) above the lower contact in San Lucas Canyon. Based on the thickness of the Keeler Canyon Formation in the Nelson Range and on reported Early Permian fossils in the Cerro Gordo area to the west (Merriam, 1963), the age of the Keeler Canyon in the north half of the study area probably ranges in age from Late Pennsylvanian to Early Permian.

Owens Valley Formation

The Owens Valley Formation was named by Merriam and Hall (1957) for beds exposed on the west flank of the southern Inyo Mountains. Initially mapped as Bird Spring Formation by McAllister (1956) in the Ubehebe Peak quadrangle, the Owens Valley Formation is recognized only in the southern half of the study area where it crops out at Conglomerate Mesa and in the hills west of Lee Flat. Four distinct lithologic units are recognized within the Owens Valley, the upper three of which correspond to the three informal members recognized by Elayer (1974) on the east side of the Inyo Mountains. The lowest unit, designated as the basal conglomerate is part of the lower member, but is discussed separately due to its bearing on stratigraphic interpretations.

The basal conglomerate of the lower member rests with sharp erosional(?) contact on the Keeler Canyon Formation northeast of Conglomerate Mesa. Here the basal conglomerate is approximately 100 m (330 ft) thick, but thins rapidly to the northwest and southeast. Dramatic thinning of the underlying Keeler Canyon Formation where this basal conglomerate is thickest is interpreted to indicate that this is an erosional contact. Correlation of this conglomerate with a much thinner conglomerate to the southeast is tentative because of intervening faults. The basal conglomerate consists of massive, medium -to thick-bedded, granule to cobble conglomerate. Clasts are predominantly pelmetazoan debris with minor amounts of pale-red shale. Both lithology and stratigraphic relationships suggest that this unit represents debris flows deposited within a submarine canyon.

The remainder of the lower member of the Owens Valley Formation conformably overlies the basal conglomerate. Thickness of this member, including the basal conglomerate, is in excess of 830 m (2720 ft). This unit consists primarily of thin to medium bedded, brown and dark gray silty limestone, calcarenite, shale, and limestone conglomerate. The conglomerate is comprised of granule to pebble sized clasts of pelmetazoan debris, limestone, and shale, and occurs predominantly in the upper part of this member as lensoidal beds which can be traced laterally up to 300 m. The amount of silt and sand in the limestone increases up-section, and beds of calcareous siltstone and fine sandstone are abundant in higher strata.

The middle member of the Owens Valley Formation consists of approximately 300 m (980 ft) of fissle red and green shale with minor interbeds of siltstone and limestone conglomerate. The contact between the middle and lower members is poorly exposed, but seems to be gradational based on the general increase in fine sand and silt in the upper part of the lower member.

The upper member of the Owens Valley Formation is composed of approximately 180 m (590 ft) of brown, thick bedded, limestone and chert pebble conglomerate which caps Conglomerate Mesa. The contact with the underlying middle member is fairly sharp and somewhat sheared.

Though initially thought to be a major low angle fault by McAllister (1956), this contact is essentially depositional (see STRUCTURAL GEOLOGY: Faults).

Merriam and Hall (1957) report that the Owens Valley Formation ranges in age from late Early to Late Permian. Cavit (personal communication, 1977) has found Wolfcampanian (Early Permian) fossils in the basal conglomerate unit as well as in other conglomerates within the lower member.

## MESOZOIC INTRUSIVE ROCKS

### Hunter Mountain Batholith

The Hunter Mountain Quartz Monzonite, first named by McAllister (1956) for exposures near Hunter Mountain (fig. 1) in the Ubehebe Peak quadrangle, is a composite batholith which crops out over several hundred square kilometers in eastern California. K-Ar age dates reported by Burchfiel and others (1970) in the Panamint Range indicate a minimum age of 165 m.y. Recent U-Pb dating of Hunter Mountain rocks in the Darwin Hills and Inyo Mountains (Dunne, personal communication, 1977) indicate that the composite batholith ranges in age from approximately 160 m.y. (Middle Jurassic) to at least 175 m.y. (late Early Jurassic) (time scale of Van Hinte, 1976).

In the study area, the Hunter Mountain batholith crops out along the entire northeast face of the Nelson Range where it is in intrusive contact with the Keeler Canyon Formation. For the most part, the intrusive contact is sharp and concordant with bedding or transposed bedding in the country rock, and there is no evidence of extensive shearing or cataclasis. Evidence of stoping of country rock, lithic inclusions, and schlerien are rare. Mafic and aplitic dikes parallel to jointing are locally common in the batholith near the contact. Discordant apophyses and dikes of granite and aplite have intruded the country rock near the contact. McAllister (1956) proposed that pre-intrusive rocks along the crest of the range form the roof of the batholith, and extensive areas of hornfels in San Lucas Canyon suggest that the batholith extends at a shallow depth beneath much of the northwest end of the Nelson Range.

Thin section and stained slab analyses of six samples from the batholith show that medium-grained augite-hornblende-quartz monzonite (classification scheme of Streckeisen, 1976) is the most common rock type within the study area.

Carbonate rock in the contact aureole of the batholith generally has been altered to marble and calcsilicate hornfels characterized by the presence of quartz, epidote, albite and actinolite-tremolite. Siltstones and shales are altered to andalusite hornfels. Garnet is rare as are diopside and wollastonite. This mineral assemblage is consistent with metamorphism in the albite-epidote hornfels facies with emplacement temperatures probably not in excess of 450$^{\circ}$ C (Hyndman, 1972).

Andesite Porphyry Dikes

Thoroughly altered, dark-green, andesite porphyry dikes are common throughout the study area. These dikes average 2 to 5 m in width and are usually vertical. They have an average trend of N50$^{\circ}$W, but range in trend between N10$^{\circ}$W and N80$^{\circ}$W. These dikes may correlate with the Independence Dike swarm (Moore and Hopson, 1961) of Late

Jurassic age (Chen, 1977).

Because of their composition it is unlikely that these dikes are related to Quaternary basalt flows widely exposed in this vicinity.

## CENOZOIC UNITS

Cenozoic rock units include olivine basalt, fanglomerate and stream gravel.

Dark-yellowish-brown olivine basalt of Quaternary age (McAllister, 1956) occurs in discontinuous patches along the southern flank and crest of the Nelson Range and along the southern margin of the study area where it is continuous with extensive flows which cap Malpias Mesa. The present sub-horizontal orientation of these flows suggests that reorientation of pre-Cenozoic structural trends during late Cenozoic basin and range deformation has been minor (Dunne and others, 1978).

In the Nelson Range, basalt locally overlies thin accummulations of fanglomerate and lakebed deposits which McAllister (1956) has correlated with the Coso Formation of Late Pliocene or Early Pleistocene age. These deposits, which rest unconformably on Paleozoic and Mesozoic basement, and the overlying basalt were not mapped separately.

Holocene alluvium consists of stream gravel and fanglomerate derived from bedrock in the Nelson Range and Inyo Mountains.

# STRUCTURAL GEOLOGY

## FOLDS

### Introduction

Folds are common in Paleozoic rocks throughout the study area. Fold axes trend predominately northwest-southeast parallel to the "structural grain" of this region (Dunne and others, 1978). However, folds with axes which trend northeast-southwest, at a high angle to the "structural grain", occur in both the Nelson Range and southeastern Inyo Mountains.

In the Nelson Range a major northeast-trending syncline is refolded by northwest-trending folds which have northwest-striking, southwest-dipping axial plane cleavage. Based on this refolding relationship, two generations of folds are recognized in the Nelson Range. The older northeast-trending folds are designated $F_1$, and the younger northwest-trending folds are designated $F_2$. Both $F_1$ and $F_2$ folds predate the Early to Middle Jurassic Hunter Mountain batholith.

In the southeastern Inyo Mountains several areas are characterized by folds or warps with southwest trending axes. The age or ages of these folds relative to northwest-trending folds in the south half of the area are not evident. Similarly, since Paleozoic strata do not crop out continuously between the northern and southern parts of the study area, the relationship of any of these folds with folds or the batholith in the Nelson Range cannot be established directly.

For purposes of structural analysis of folds, the study area has been divided into 15 structural domains (fig. 3). Each domain was uniformly sampled and the divisions are based on homogeneity of the

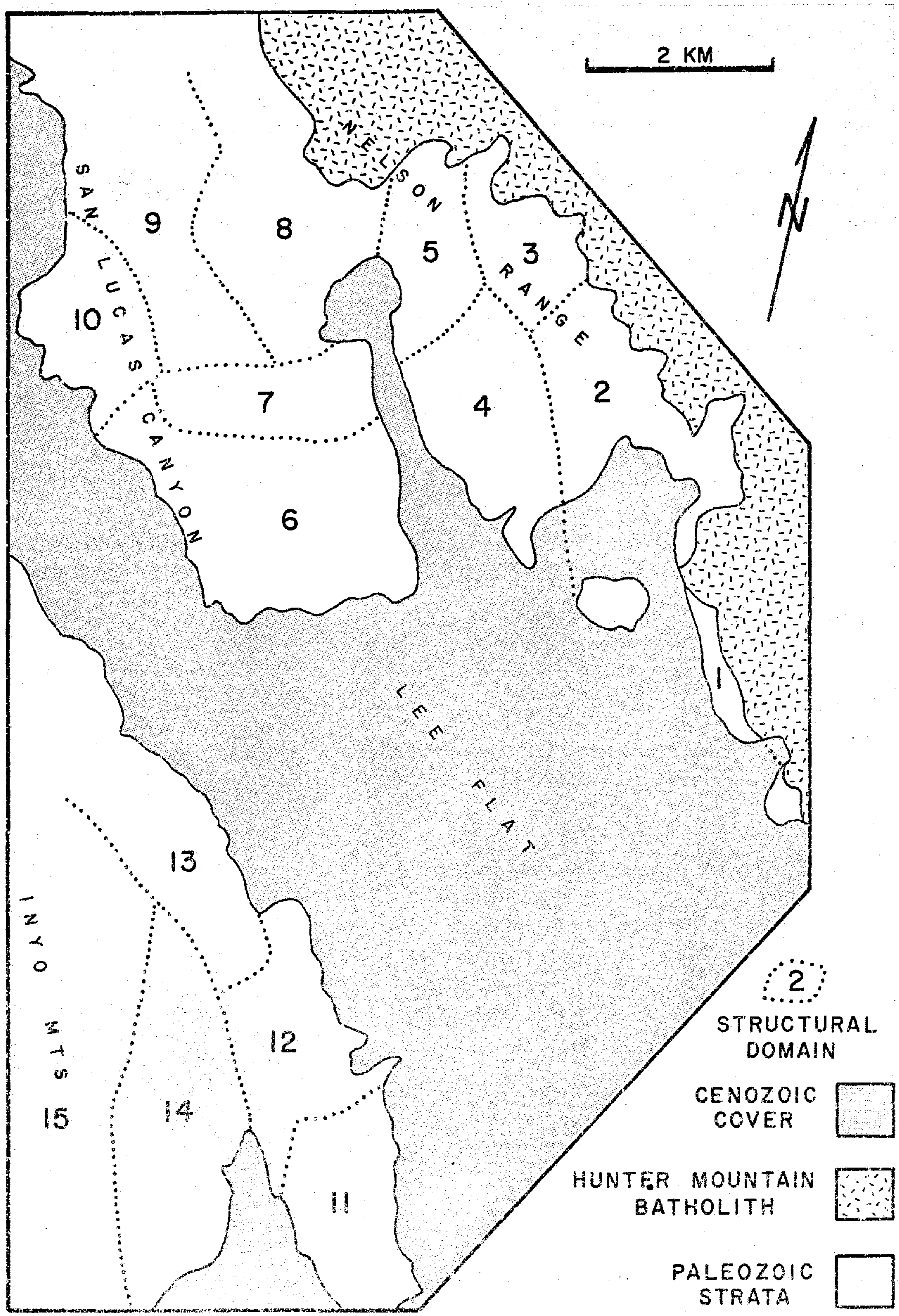



Figure 3. Index map to structural domains.

megascopic fabric of the rocks. Although some domain boundaries coincide with faults, the majority of the divisions are arbitrary and are based on gradual changes in fold geometry.

Planar fabric elements represented in stereonet plots (fig. 4) consist of bedding ($S_0$), foliations related to the axial planes of $F_1$ folds ($S_1$), and axial plane cleavage related to $F_2$ folds ($S_2$) and other northwest-trending folds. Linear fabric elements consist of minor folds (amplitude between 5 cm and 2 m), bedding-cleavage intersections, boudins, mullions and mineral lineations. Only in a few instances could lineations be assigned to a particular fold generation with confidence, therefore, they are not differentiated by age in the stereonet plots or on the map (Plate 1).

Fold terminology used in the following pages is that of Fleuty (1964).

## Folds in the Nelson Range

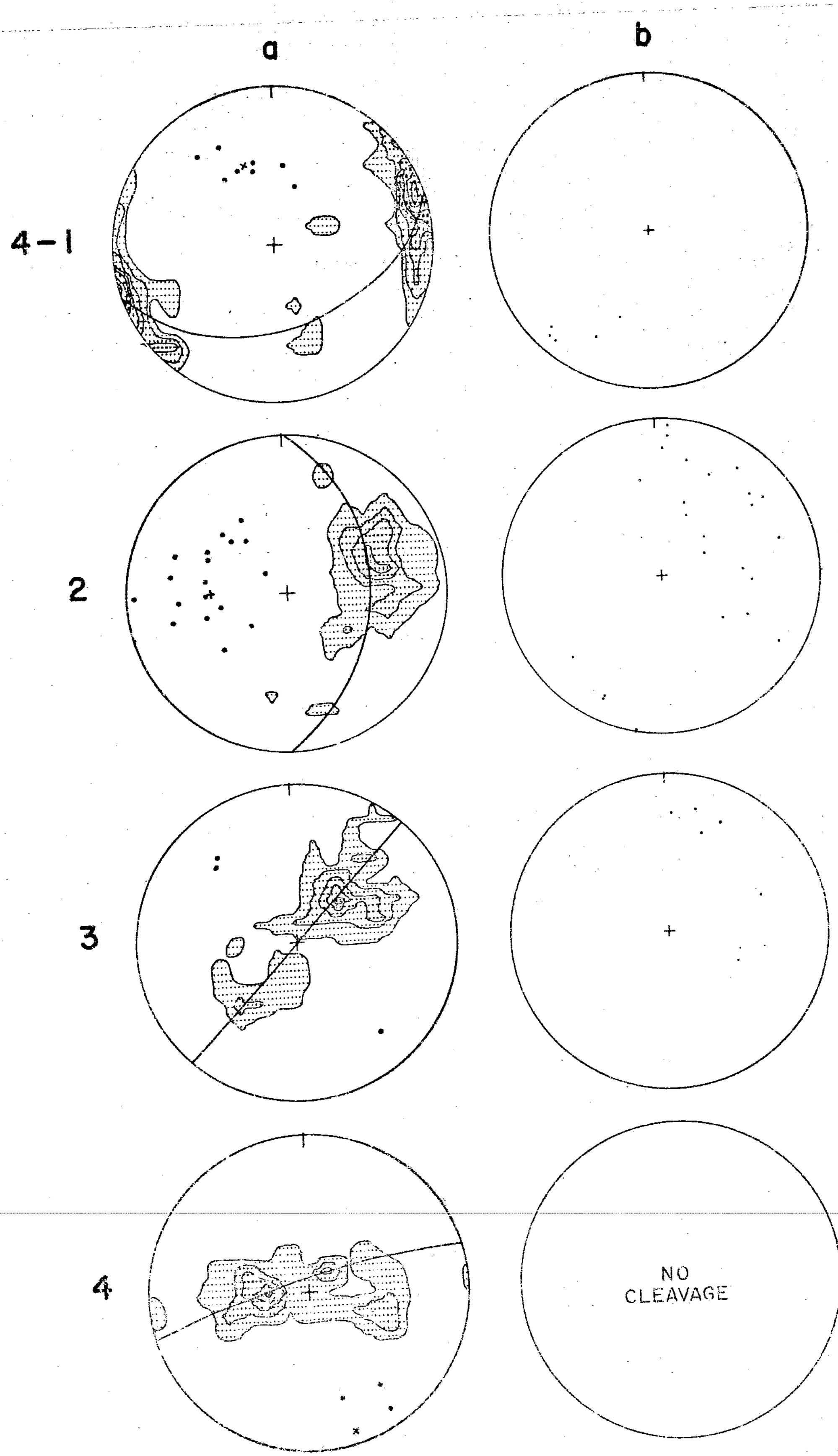
### $F_1$ - The San Lucas Canyon syncline

The major $F_1$ generation fold, here named the San Lucas Canyon syncline (Plate 1), is a southeast-overturned syncline which is well exposed along San Lucas Canyon Road (fig. 5a) where it trends N. 15$^{\circ}$ E. and plunges 20$^{\circ}$. Strata of the basal Keeler Canyon Formation and the underlying Rest Spring Shale which comprise the overturned limb can be traced over 3 km from upper San Lucas Canyon northeast into the Nelson Range. Based on structural trends and similar stratigraphy, this overturned limb can be projected another kilometer beneath the alluvium of upper Lee Flat and toward the intrusive contact with the

Figure 4. Structural analysis of a) bedding planes and lineations, and b) axial plane foliations. Numbers 1-15 correspond to structural domains shown in Figure 3. North is towards top of page.

a) Contoured equal-area, lower hemisphere stereonet diagrams of poles to bedding ($S_0$). Contours are 2, 4, 6, 8 and 10 percent points per one percent area. Dots are trend and plunge of lineations. Solid line and "X" are best fit great circle and pi point, respectively. Number of points contoured: 1(30), 2(115), 3(100), 4(102), 5(147), 6(136), 7(93), 8(140), 9(145), 10(72), 11(109), 12(105), 13(127), 14(140), 15(150).

b) Lower hemisphere, equal-area projections of poles to axial plane foliations. Small "X" represent $S_1$, dots represent $S_2$ in domains 1-9 and all other foliations in 10-15.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b
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2
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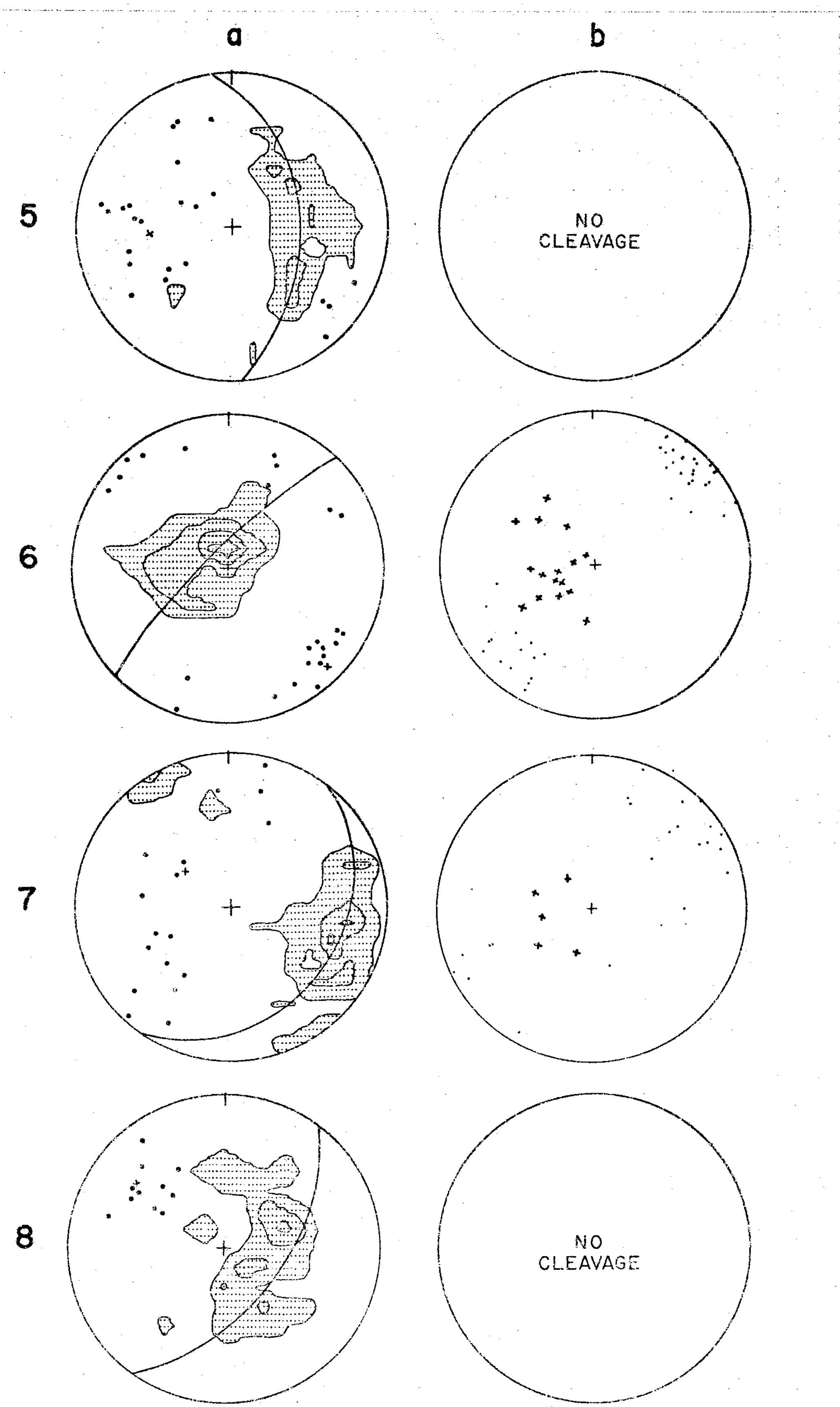
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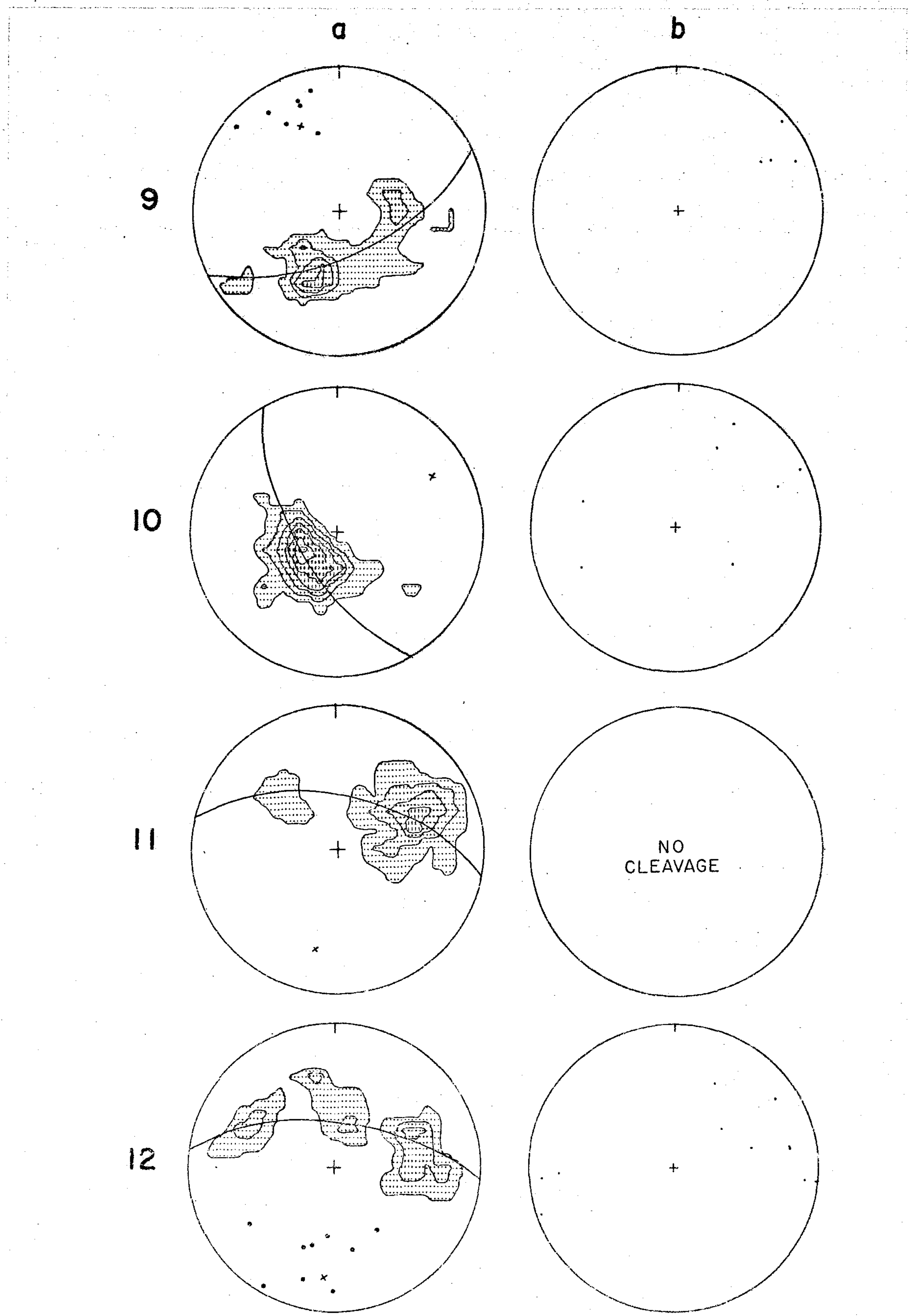
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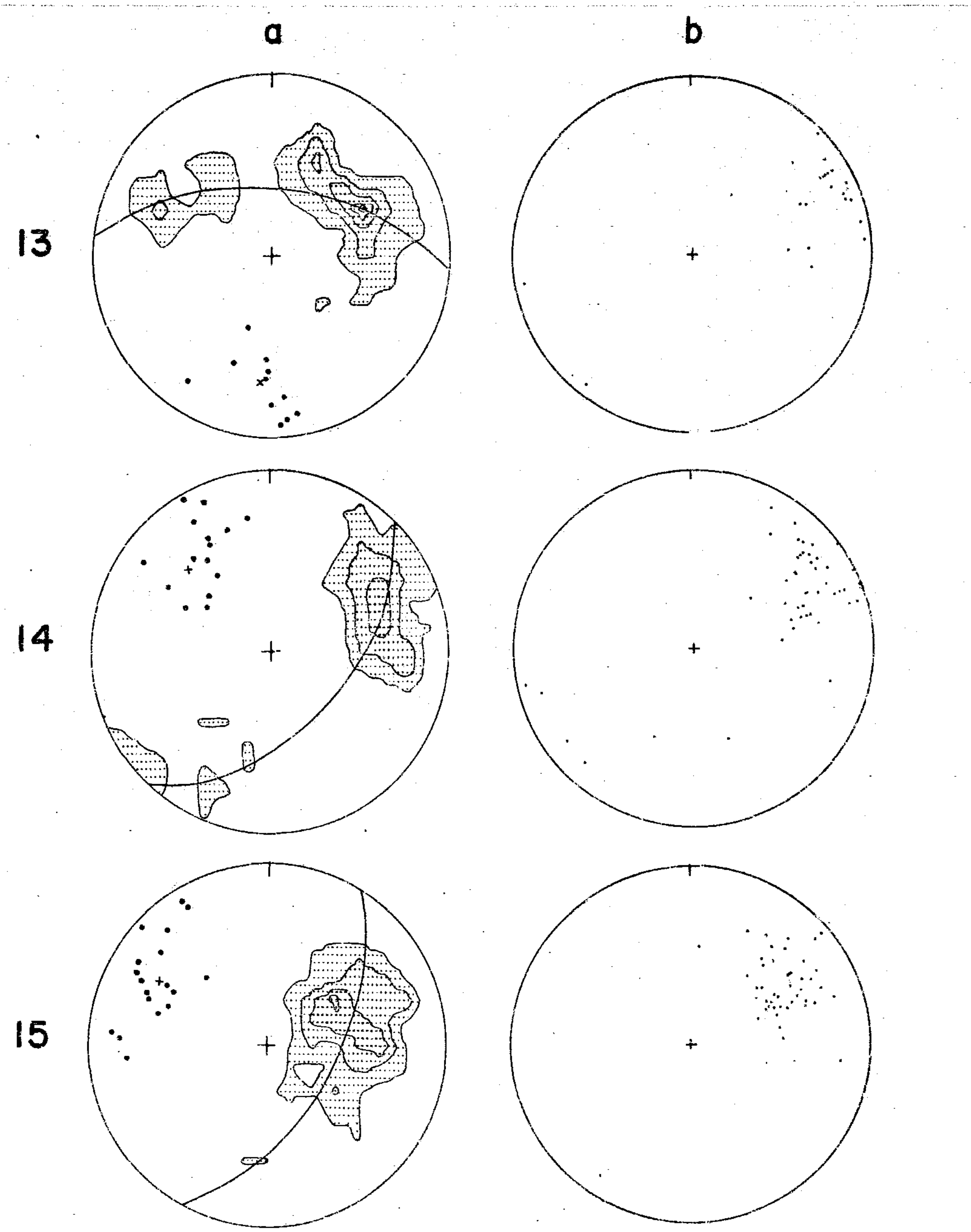
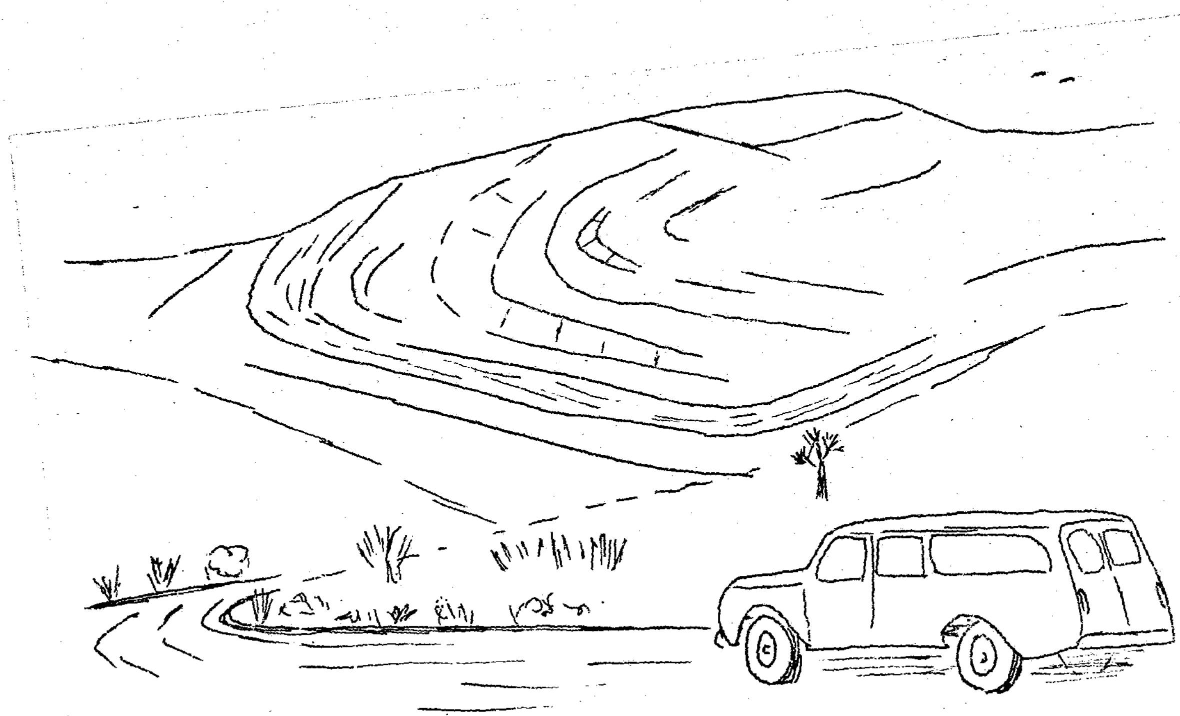

a
b
13
14
15

Figure 5. Sketches from photos of the San Lucas Canyon syncline and related minor structures. Hammer handle in 5b and 5c is 28 cm long.

5a) Core of the San Lucas Canyon syncline. Fold is viewed approximately parallel to trend of axis (N 15$^{o}$ E) from a distance of 400 m.

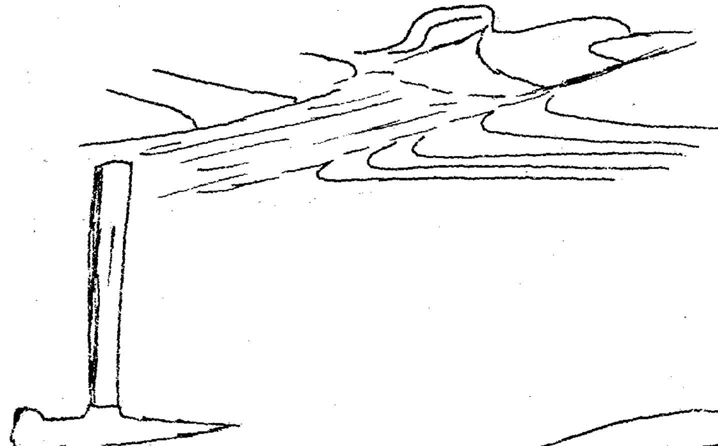
5b) Minor fold in domain 6 and related incipient transposition foliation. Fold axis orientation is 40$^{o}$, N25$^{o}$ E.

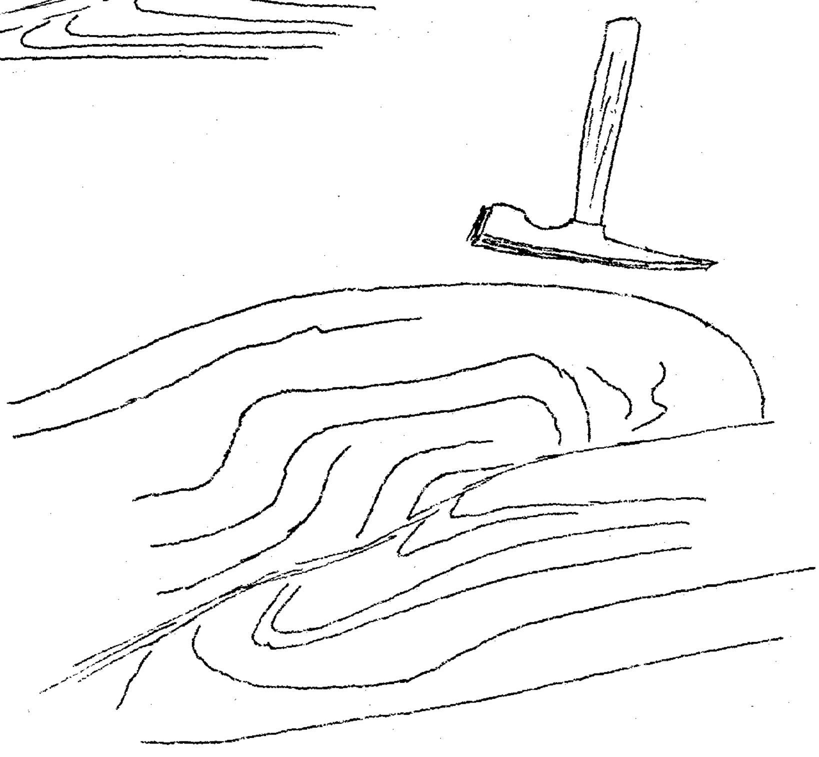
5c) Minor fold in domain 7. Fold axis orientation is 52$^{o}$, N 27$^{o}$ E. Slip surface (N 55$^{o}$ E, 25$^{o}$ SE) is roughly axial planar.



5a.

5b.





5c.

Hunter Mountain batholith at the crest of the Nelson Range.

Structural domains 4 and 6 and domains 5 and 7 (fig. 3) represent the upright and overturned limbs, respectively, of the San Lucas Canyon syncline. Plots of domains 4 and 6 (fig. 4-4a, 6a) show that the upright limb generally consists of strata with shallow northwest dips which are gently folded about a northwest-trending ($F_2$) axis. The plot of $S_0$ of domain 7 (fig. 4-7a) indicates that the western end of the overturned limb consists of steeply northwest and southwest dipping strata. The hinge of the fold in this domain is rounded and is close to tight in profile. The plot of $S_0$ in domain 5, the eastern projection of the overturned limb, shows that bedding dips steeply to the west (fig. 4-5a). In the southern part of this domain the core of the San Lucas Canyon syncline is complexly faulted (see STRUCTURAL GEOLOGY: Reverse faults). Neither this fault nor evidence of synformal closure of upright with overturned beds could be discerned to the north, presumably due to the increased effects of thermal metamorphism and poor exposure.

In and near the core of the San Lucas Canyon syncline in domains 6 and 7 are minor southeast overturned folds (fig. 5b, c) which parallel the trend of the main fold. These plunge variously northeast and southwest, possibly a response to refolding about northwest-trending $F_2$ axes (fig. 4-6a, 6b). Within some beds there are surfaces of slip or incipient transposition and flow which are axial planar to these overturned folds (fig. 5b, c). These surfaces, designated $S_1$, may represent a weakly developed or poorly preserved axial plane foliation to the San Lucas Canyon syncline. Plots of poles to these surfaces (fig. 4-6b, 7b) show that they generally dip at low angles to the east and northeast.

The San Lucas Canyon syncline presents a structural paradox in that there is no readily apparent anticline paired with its extensive overturned limb, yet a large area of upright strata of the Keeler Canyon Formation is present not far to the north. The overturned limb is separated from these upright Keeler Canyon strata by a discontinuous band of Rest Spring Shale. The major part of this terrain of Rest Spring Shale occurs in domain 10 (fig. 3). Here bedding in part defines a gentle, upright antiform which trends N $60^{\circ}$E and plunges $25^{\circ}$ (fig. 4-10a), but very poor exposures here and in limited outcrops to the east preclude the delineation of any major anticline. It is possible that this band of shale represents the core of an overturned anticline as shown in Plate 2 (cross section A-A', B-B', C-C'). Alternatively, a substantial reverse fault within the shale could explain the observed relations; however, no distinct fault surface nor stratigraphic throw is evident.

The only other folds in the Nelson Range that belong to the $F_1$ fold set are a series of upright folds in the Keeler Canyon Formation which occur in domain 6 along the uppermost part of San Lucas Canyon Road (Plate 1). These folds trend between N. 15 E. and N. 65 E. and plunge between $20^{\circ}$ and $40^{\circ}$ which is parallel to the trend of the San Lucas Canyon syncline which lies 1.5 km to the north. These folds can be traced only a short distance, as they are covered by alluvium to the southwest and trend into poorly exposed terrain to the northeast. Their northeast extension may be reflected by the double plunge of of the $F_2$ Lee Flat anticline (Plate 1). The Lee Flat anticline seems to have been superposed on a pre-existing northeast trending anticlinal warp which lay immediately south of the hinge zone of the San Lucas Canyon syncline (Plate 2, cross section B-B'). This anticlinal warp may be a disharmonic down plunge expression of the smaller folds exposed in San Lucas Canyon.

The overall style of major and minor structures place certain constraints on the mechanism of $F_1$ folding. Where $F_1$ folds have developed in the more competent, silty limestone beds of the Keeler Canyon Formation, the beds commonly show constant or near constant orthogonal thickness with only minor thickening in the cores of folds. The profiles of such folds correspond to Class 1B and 1C folds of Ramsey (1967). This suggests flexural slip as a primary mechanism of folding; such a mechanism would be expected in multi-layered, thin-bedded units (Donath and Parker, 1964). Slip parallel to the axial plane and incipient transposition observed in some intrafolial folds suggest a component of flexural flow within less competent beds. Therefore, $F_1$ folds probably formed by flexural slip and flow.

$F_2$ - Northwest-trending Folds

The trends of $F_2$ folds in the Nelson Range span the entire northwest quadrant, although the majority of folds range between N 15$^{\circ}$ W and N 65$^{\circ}$ W and have an average trend of N 50$^{\circ}$ W. The plunges of $F_2$ folds vary from horizontal to steep to the northwest and southeast. Variations in plunge amount and direction in some areas are the result of their superposition on pre-existing $F_1$ folds. Associated with $F_2$ folds is a prominant axial plan cleavage ($S_2$). $S_2$ is poorly expressed or absent in domains which are close to the intrusive contact. However, the sporadic occurrence of axial plane foliation in some folds near the intrusive contact suggests that the foliation was once widespread in the carbonate rocks and subsequently has been obliterated in these areas by recrystallization due to thermal metamorphism. Along portions of the intrusive contact the style and trend of $F_2$ folds have been modified by emplacement of the Hunter Mountain batholith. The areas most severely affected by the batholith will be discussed separately.

Structural domains 6 and 7 are the only areas in which refolding of $F_1$ by $F_2$ folds can be clearly demonstrated. The dominant $F_2$ fold in these domains is an upright, doubly plunging anticline herein named the Lee Flat anticline. On the basis of field relationships and as seen in aerial photographs this $F_2$ fold refolds the $F_1$ San Lucas Canyon syncline. This refolding relationship is illustrated diagrammatically in figure 6. As can be seen the Lee Flat anticline changes from a normal anticline to an antiformal syncline where its hinge trace crosses the hinge trace of the San Lucas Canyon syncline. In domain 6, the double plunge of the Lee Flat anticline is the result of superposition of this anticline on the antiformally warped, upright limb of the San Lucas Canyon syncline (Plate 2, cross sections A-A', B-B'). This has locally produced a dome and basin structure visible in aerial photographs.

In domain 6 (fig. 4-6a) the plot of $S_0$ defines a pi point that trends S 45° E and plunges 10°. In domain 7, where $F_2$ folds have been superimposed upon the steeply north-dipping overturned limb of the San Lucas Canyon syncline, the pi point plunges 45°, N 55° W (fig. 4-7a).

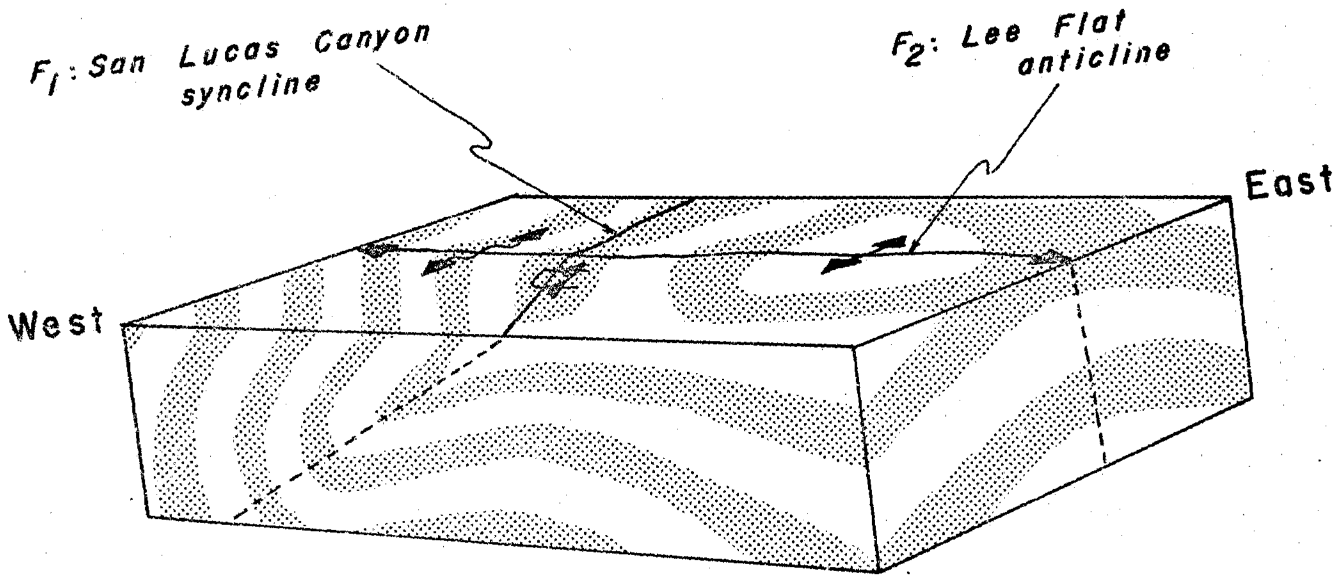
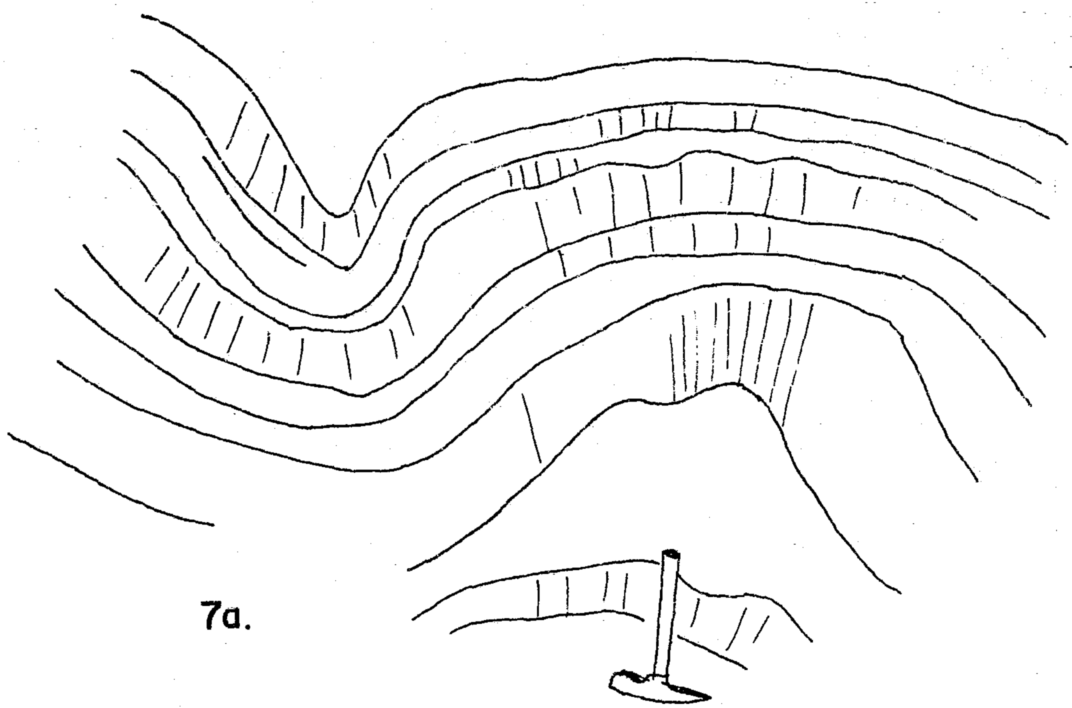



Figure 6. Schematic block diagram which shows the refolding of the San Lucas Canyon syncline by the Lee Flat anticline. View is to the north.

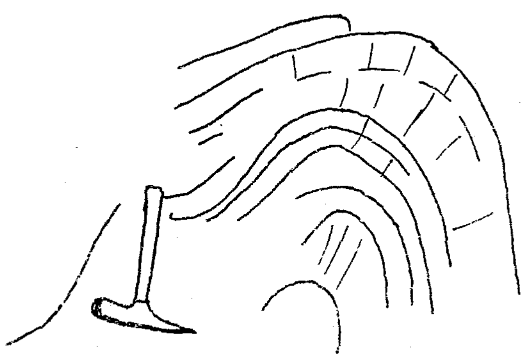
Rather than several folds of large amplitude and wavelength, as seen in domain 6, $F_2$ folds in domain 7 are more numerous and of small amplitude and wavelength. Minor northwest-trending folds (fig. 7a) are abundant in both these domains and $S_2$ axial plane cleavage is well developed. In clean limestones and shales this cleavage is essentially continuous and dips steeply to the southwest, whereas in silty limestones this cleavage is more widely spaced (1-5 cm) and commonly is fanned or refracted such that it dips steeply southwest and northeast. A step-like bedding-cleavage intersection which locally is associated with cleavage in more competent layers in domain 6 indicates small amounts of movement (less than 2 mm) along cleavage surfaces.

Domains 4 and 5 (fig. 3), which encompass the eastern half of the San Lucas Canyon syncline, do not display the degree of superposed folding as seen in domains 6 and 7. Domain 4 (fig. 4-4a) consists of several open, upright folds which trend N. 15$^{o}$ W and are horizontal or plunge very gently southeast. Small scale folds and distinct axial plane foliation are lacking. Domain 5 is characterized by gentle to open upright folds which plunge 35$^{o}$, S. 82$^{o}$ W (fig. 4-5a). This approximately due west trend is believed to be the result of deflection or counter-clockwise rotation of northwest trending folds in a localized zone adjacent to the Hunter Mountain batholith. These folds refold the projected overturned limb of the San Lucas Canyon syncline and are represented as synformal anticlines and antiformal synclines on Plate 1.

Domain 9 contains a series of upright and overturned folds which trend N. 25$^{o}$ W and plunge approximately 35$^{o}$ (fig. 4-9a). Near the



7a.



7b.

Figure 7. Sketches from photos of minor northwest-trending folds related to $F_2$. Views are parallel to axial trends; hammer handle is 28 cm long.

7a) Minor folds in core of Lee Flat anticline in domain 6. Fold axes orientations are $9^{\circ}$, N $37^{\circ}$ W and $4^{\circ}$, S$37^{\circ}$ E. Note refracted axial plane cleavage.

7b) Minor fold in domain 8 which is oriented $50^{\circ}$, N $45^{\circ}$ W. Rocks here are metamorphosed to calcsilicate hornfels, yet style of fold is similar to that in unmetamorphosed areas (fig. 7a).

axis of San Lucas Canyon, $F_2$ folds are upright and open, but eastward these folds become progressively tighter and locally are overturned. Hinge zones of folds here commonly are fractured. The steeply west dipping limb of an overturned syncline along the east margin of domain 9 forms most of the upper east wall of San Lucas Canyon. Along trend this fold changes from overturned and isoclinal to upright and open in profile. Locally the core is characterized by minor folds, boudins and reverse faults.

Domain 8 (fig. 3) consists of open to tight, upright folds along the crest of the Nelson Range. Rocks here are thoroughly metamorphosed to calcsilicate hornfels and represent a portion of the roof rocks of the Hunter Mountain batholith. Due to metamorphism, some fold hinge traces could be followed only approximately. Poles to bedding indicate that folds trend N. $55^{\circ}$ W and plunge approximately $30^{\circ}$ (fig. 4-8a). Some of these folds trend into the intrusive contact and apparently are truncated.

Lithology of the Keeler Canyon Formation has strongly influenced the style of $F_2$ folds at all scales. Minor folds (fig. 7a, b), as well as major folds, exhibit disharmonic style in profile due to the thin bedded nature of the rock. Profiles of folds which involve thick limestone beds and more competent silty limestone beds show constant or near constant orthoganal thickness of these beds such as Class 1B and Class 1C folds of Ramsey (1967). Less competent interbeds of limestone and shale thin and thicken irregularly in the hinges and on the limbs of these folds apparently in response to folding of the more competent beds. These style characteristics of $F_2$ folds suggest that flexural slip and flow were the primary mechanisms of folding (Donath and Parker, 1964). Closely spaced cleavage in some beds and minute offset along spaced cleavage in other beds suggest that a small component of passive folding may also have occurred.

Folds adjacent to the Hunter Mountain Batholith

Several lines of evidence indicate that the major folds in the Nelson Range predate emplacement of the Hunter Mountain batholith. This evidence contradicts McAllister's (1956) interpretation that plastic deformation and shouldering aside of Paleozoic rocks during emplacement of the batholith produced many of the folds in the Nelson Range.

First, rocks of the Keeler Canyon Formation and Rest Spring Shale within approximately 1 to 2 km of the intrusive contact have been metamorphosed to calcsilicate hornfels and andalusite hornfels, respectively. In spite of the silicification associated with this contact metamorphism, the style of $F_2$ folds within most of the calcsilicate terrain is similar to that of $F_2$ folds in unmetamorphosed rocks (fig. 7a, b). If metamorphosim had occurred prior to folding, then the mechanism and hence style of folding in such brittle, silicified rock would be expected to contrast markedly with that in relatively incompetent limestone and shale.

Second, truncation of $F_2$ folds and bedding along portions of the contact (Plate 1) suggest that the batholith intruded rocks which were already deformed.

Third, apparent tightening or appression of fold hinges and deflection of axial trends of pre-existing folds has occurred within a zone approximately 1 km wide immediately adjacent to portions of the intrusive contact. Style and geologic environment of these modified folds are similar to those of the "boudinaged flow folds" as described by Gulliver (1976a) in the Talc City Hills to the south. Gulliver (1976a) proposed that these structures are the result of appression and tightening of the cores of pre-existing folds due to renewed contractile strain during emplacement of a nearby pluton. This renewed strain caused axial plane cleavage within the cores of the pre-existing folds to act as a plane of separation along which silty and sandy layers were boudinaged and rotated within a matrix of more ductile limestone. It is believed that an analogous mechanism is responsible for similar structures in domains 1, 2 and 3 (fig. 3) in the Nelson Range.

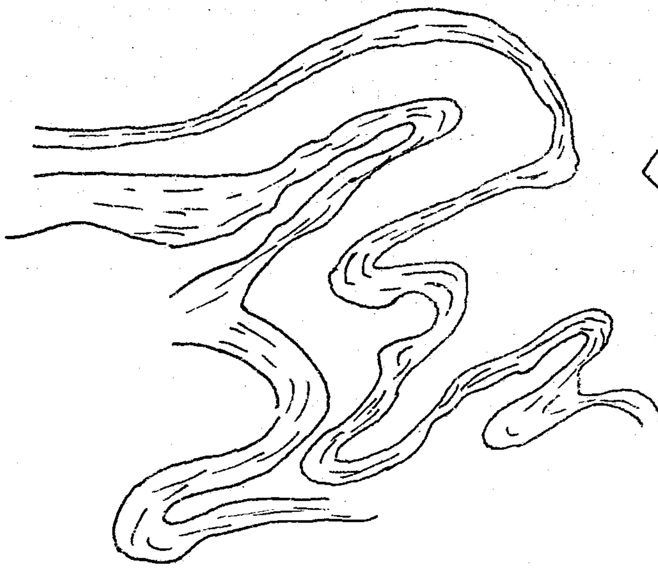
Effects of the batholith are most extreme in domain 1 which represents a segment of wall rock adjacent to the batholith. Figure 4-1a shows that bedding dips very steeply to the northeast or southwest along the intrusive contact. Numerous minor folds which plunge to the northwest are parallel with boudins in silty and shaly layers. This, in conjunction with closures seen in aerial photos, strongly suggests that rocks in domain 1 comprise upright, isoclinal folds, the axes of which are oriented $45^{\circ}$, N $20^{\circ}$ W. For reasons previously cited it is believed that these are pre-intrusive folds which have been appressed or tightened during emplacement of the batholith. Evidence of tightening of fold hinges in domain 1 consists of zones of boudinage which range up to 1 m in thickness (fig. 8c). Boudins consist of rectangular blocks of calcsilicate hornfels up to 15 cm in length in a marble matrix. The blocks generally show evidence of rotation and are aligned roughly parallel to the gross structural grain of this

Figure 8. Sketches from photos of structure modified by emplacement of the Hunter Mountain batholith.

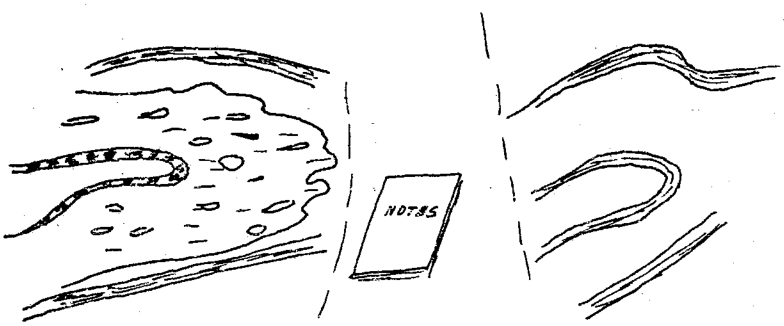
8a) Flattened minor fold in domain 2 ($51^{o}$, N $83^{o}$ W). Hammer is for scale (28 cm).

8b) "Boudinaged flow fold" in domain 2 ($35^{o}$, S $30^{o}$ W). Notebook is 13 cm x 20 cm.

8c) Zone of boudinage in suspected fold core in domain 1. Hammer is for scale (28 cm).



8a.

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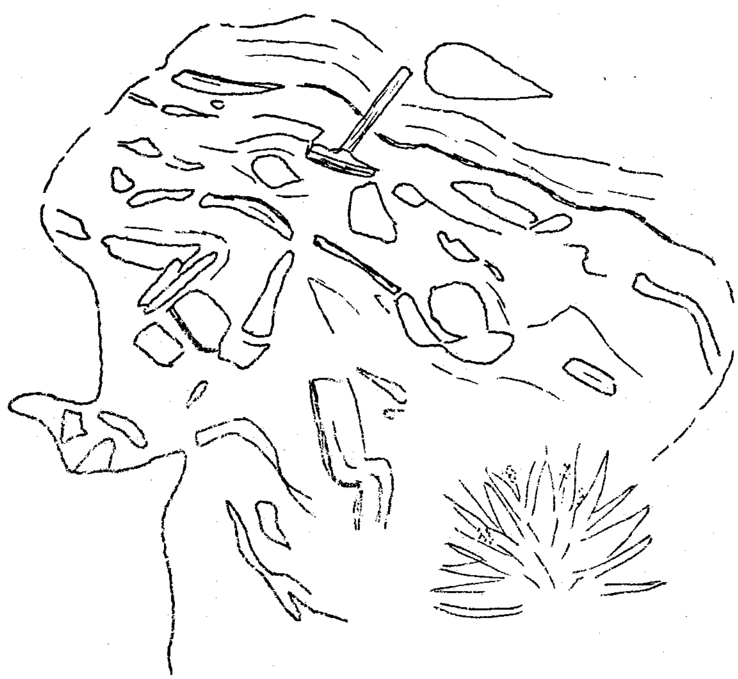


8c.

domain. The marble matrix contains minor isoclinal folds. A locally well developed cleavage or transposition foliation dips steeply northeast. If this represents a foliation which is axial planar to $F_2$ folds, then the batholith has rotated this formerly southwest dipping planar element to the west about an axis roughly parallel to the original fold axes.

Domain 2 is composed of generally west-dipping strata (fig. 4-2a). Field evidence indicates that bedding is folded into northwest trending, northeast overturned isoclinal folds of the $F_2$ set. Evidence includes local reversal in dip, upright and overturned beds and northwest-trending zones of minor folds which are believed to represent cores of major folds. The style of some of these minor folds (fig. 8a, b) are similar to the "boudinaged flow folds" of Gulliver (1976a, fig. 14) which suggests that major folds in this domain have been tightened by the intrusion of the batholith. The orientations of minor fold axes range from northwest to southwest (fig. 4-2a) and generally lie within the plane of bedding.

In domain 3, field evidence of folding can be seen only locally on the east face of the Nelson Range were a series of tight, northeast-overturned folds are exposed (fig. 9). The cores of these folds are disharmonic and locally are characterized by bedding plane faults. Tightness of the fold hinges is attributed to the adjacent and subjacent batholith. The plot of poles to bedding (fig. 4-3a) indicate that fold axes are horizontal and trend N $45^{\circ}$ W.

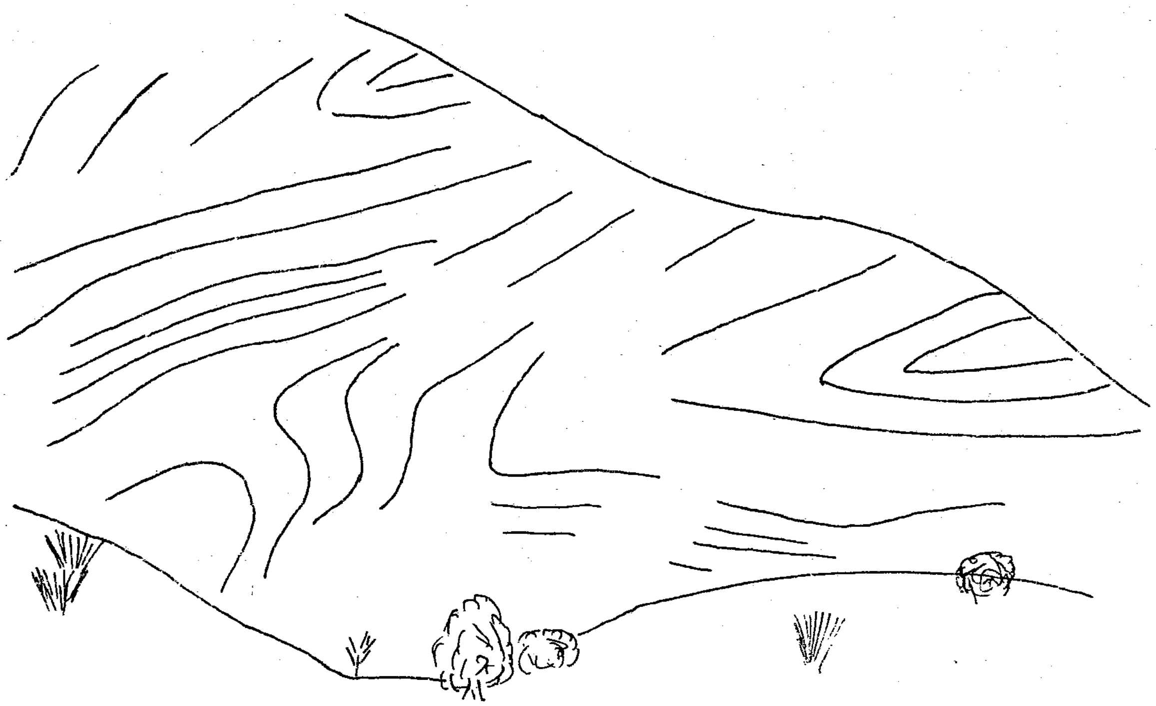


Figure 9. Sketch from photo of form surfaces in folds adjacent to the batholith in domain 3. Fold axes are horizontal and trend N 45$^{o}$ W. View is N 60$^{o}$ W from a distance of 600 m.

## Folds in the Southeastern Inyo Mountains

### Southwest-trending folds

Southwest-trending folds and warps characterize the foothills of the southeastern Inyo Mountains immediately west of Lee Flat. This area corresponds to structural domains 11, 12 and 13 (fig. 3). Folds here plunge to the south at gentle to moderate angles and range in trend between south and S $24^{o}$ W.

Domain 11 contains an upright, open syncline. A plot of poles to bedding (fig. 4-11a) shows that the fold axis is oriented $30^{o}$, S $15^{o}$ W. This domain lacks cleavage and minor folds. As can be seen in map and cross sectional views (Plate 1 and Plate 2, E'E') the northwest limb of this fold is of small areal extent compared to its southeast limb. This southeast limb blends into a southwest dipping homocline of Keeler Canyon Formation and lower Owens Valley Formation which extends almost 2 km to the southern border of the map. If projected south beneath an extensive cover of Cenozoic basalt, this homocline matches the stratigraphic and structural trends exposed in the Santa Rosa Hills.

Domain 12 comprises a recumbent, eastward overturned syncline (fig. 10) the axis of which plunges $30^{o}$ due south. This syncline has an overturned limb composed of the basal Keeler Canyon Formation and Rest Spring Shale. Faults complicate the relationship between this overturned limb and upright beds of the Keeler Canyon to the west which are warped into gentle southwest-plunging, upright, open folds. McAllister (1956) shows an overturned anticline in the intervening area between upright and overturned beds, however, the hinge trace of this proposed fold lies entirely within a poorly exposed terrain of Rest

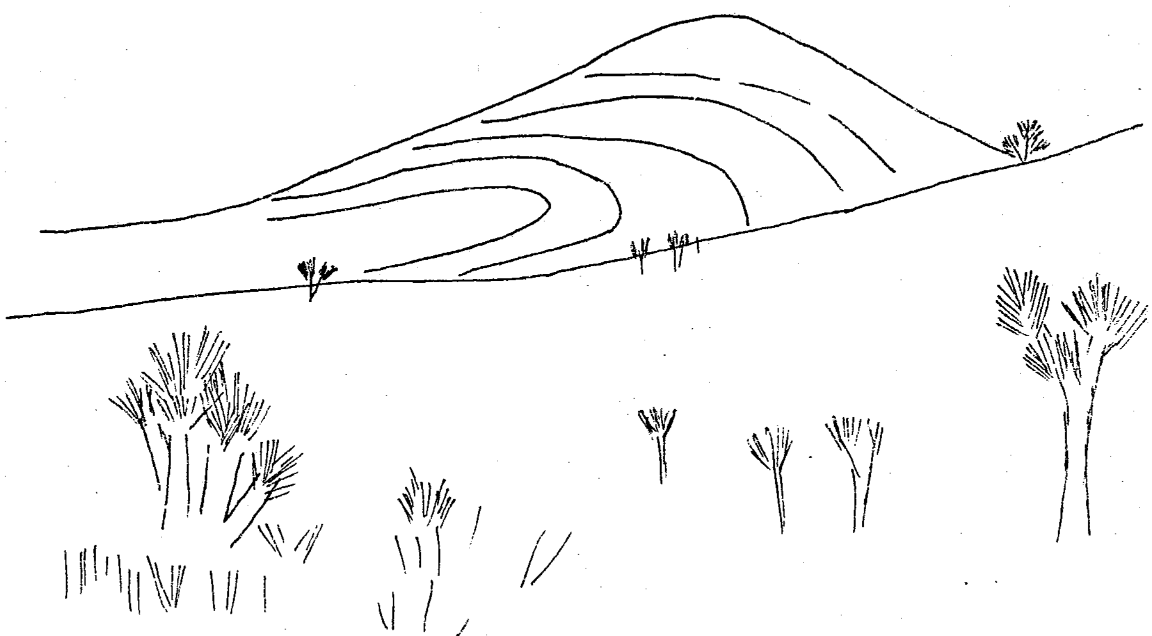


Figure 10. Sketch from photo of recumbent syncline in domain 12. Fold axis orientation is $30^{\circ}$, S. View is due south from a distance of 500 m.

Spring Shale and no such antiformal closure was found during the present study.

A stereonet plot of poles to bedding (fig. 4-12a) for domain 12 shows that folds have an average orientation of 25$^{o}$, S 5$^{o}$ W. This orientation is paralleled by minor folds. Cleavage which is axial planar to these minor folds is poorly developed (fig. 4-12b). Near the lower contact of the Keeler Canyon Formation is an incipient transposition foliation which is parallel to bedding. Associated with this foliation is a mineral lineation which lies in the plane of the foliation. Boudins are present in silty layers. Both parallel the trends of major and minor folds.

In the southern part of domain 13 (fig. 3) the basal conglomerate of the Owens Valley Formation is folded into an open synclinal warp which plunges to the west and is similar to open warps in domain 12 to the south. In the west central portion of this domain, folds are upright, plunge southwest, and range from close to tight in profile. To the east these folds become more open and poorly defined. The northernmost part of this domain is complexly faulted, and folds are poorly exposed. A stereonet plot of domain 13 shows that, in general, folds plunge 30$^{o}$, S 5$^{o}$ W (fig. 4-13a). This trend is parallel to several types of lineations including minor folds, mullions and stretched "golf balls" in the lower part of the Keeler Canyon Formation. Southwest-dipping foliation (spaced 1-10 mm) which is axial planar to major and minor folds is well developed (fig. 4-13b).

Although folds in these domains have a narrow range of trend, they exhibit a diversity of styles which in part may be attributed to the effects of differing lithologies or the dominant mechanisms of folding.

The recumbent syncline in domain 12 (fig. 10) involves the basal part of the Keeler Canyon Formation which consists of relatively competent, massive conglomerates with interbeds of less competent limestone and shale. This fold shows thinning on its limbs such as Class 2 folds of Ramsey (1967). Incipient transposition and layer parallel lineations suggest that the transfer of material away from the limbs was by flexural flow (Donath and Parker, 1964). Therefore, flexural slip and flow were probably the dominant mechanisms involved in formation of this fold.

To the north in domain 13, the lower Keeler Canyon Formation has substanitally more interbeds of shale and less conglomerate. This apparently has resulted in a lower competency contrast which favored a more passive mechanism of folding and the formation of well developed axial plane cleavage.

The open profile of the syncline in domain 11 is possibly a result of the competent siltstone member of the Keeler Canyon Formation which occurs in the core of the fold.

Northwest-trending folds

A series of northwest-trending folds occur in domains 14 and 15 which consist of the low, rolling hills east of Conglomerate Mesa (fig. 3). Most of these folds trend between N 30° W and N 65° W, and have an average trend of N 50° W.

Domain 14 consists of open to isoclinal, upright to northeast overturned folds. Fold axes trend approximately N. 45° W. and most plunge approximately $35^{o}$ to the northwest (fig. 4-14a), but at the southern margin of the domain some folds plunge gently to the southeast. In cross section folds are asymmetrical with short northeast-dipping or overturned limbs and long southwest-dipping limbs. Axial plane foliation (spaced 1-10 mm) is very prominent and parallels bedding on southwest dipping limbs (fig. 4-14b). Minor folds and bedding-cleavage intersection lineations parallel the trend of the major folds.

Domain 15 contains upright, close to tight folds with angular hinges. Figure 4-15a shows that on the average, these folds trend N $60^{o}$ W and plunge approximately $30^{o}$ northwest. As in domain 14, these folds are asymmetric and have prominent southwest dipping axial plane cleavage. These folds are clearly seen only in the lower member of the Owens Valley Formation. Although the middle shale member is strongly foliated, no macroscopic folds were observed. The upper conglomerate member of the Owens Valley Formation was not mapped, however, McAllister (1956) shows northwest-trending folds in this unit.

The profiles of major folds in domains 14 and 15 are poorly exposed, but anastimosing and en echelon hinge traces in map view (Plate 1) indicate that the folds are strongly disharmonic in profile. This is probably due to the thin-bedded nature of the lower member of the Owens Valley Formation and suggests that flexural slip between layers may be the primary mechanism of folding in the more competent silty limestone and conglomerate beds. Abundant interbeds of less competent shale and shaly limestone have favored the formation of closely spaced axial plane cleavage.

## FAULTS

Faults in the study area have been divided into three groups: 1) reverse faults; 2) northwest-trending high angle faults; and 3) northeast-trending high angle faults. Reverse faults are interpreted to be the result of Mesozoic compressional deformation. Most high angle faults in the study area are thought to be related to Cenozoic basin and range deformation. However, the occurrence of andesite porphyry dikes of probable Late Mesozoic age along some of the northwest-trending high angle faults suggests that some of these faults are pre-Cenozoic.

### Reverse Faults

Reverse faults in the study area usually are rooted in and shear out the limbs and cores of major folds and are thought to be related to Mesozoic compressional deformation.

A reverse fault is recognized in the northern part of domain 13 (fig. 3) based on the repetition of a sequence of Rest Spring Shale and the "golf ball" horizon of the basal Keeler Canyon Formation. Rocks of the Keeler Canyon Formation immediately below the fault surface contain numerous northwest-trending,east-vergent drag folds and a southwest-dipping axial plane foliation. To the east, an isolated patch of Keeler Canyon Formation overlies and truncates strata of the Rest Spring Shale. This outlier is believed to represent a klippe of the reverse fault to the west. Strata in this klippe are folded into an east-overturned anticline. On the basis of this and the minor structures below the main fault surface to the west, it seems likely that this fault is genetically related to folds in this area and that the slip direction was to the east. Since the outlying klippe of this fault superposes younger over older strata, an estimate of slip amount is not possible.

A reverse fault 0.5 km long occurs in the core of the overturned syncline along the eastern margin of domain 9 (fig. 3). This steeply west dipping fault superimposes overturned beds eastward over upright beds. Beds on either side of the fault are characterized by transposition foliation which is parallel to the fault surface and minor isoclinal drag folds which are parallel to the trend of the syncline. Throw on this fault is very small, and the fault dies out along trend to the north where the fold becomes upright. Farther north another portion of the core of this syncline is faulted in a similar manner.

Along the northern boundary of domain 11 (fig. 3), a south-dipping fault occurs along the lower contact of the siltstone member of the Keeler Canyon Formation. The fault is characterized by a breccia zone 1 m thick which consists of angular fragments of quartzite and siltstone. This fault generally is parallel to bedding above the fault, but locally truncates underlying strata. On the basis of its orientation this fault is interpreted to be a reverse fault which is rooted in the core of the syncline immediately to the south. There is no stratigraphic evidence of significant movement on this fault and the fault probably dies out at shallow depth (Plate 2, cross section E-E').

In domain 5, beds of the Keeler Canyon Formation in the projected overturned limb of the San Lucas Canyon syncline are juxtaposed over upright strata to the east without a well defined synformal closure. A structurally complex zone, which is 10 to 30 m wide and oriented parallel to bedding, separates upright and overturned beds in the southern part of this domain. This zone is characterized by irregular pods of limestone breccia, boudins, incipient transposition of bedding and complex folds (fig. 11). These folds have diverse orientation (fig. 4-5a) and style; axial plane cleavage is conspicuously absent.

It is believed that this zone represents the core of the San Lucas Canyon syncline which has been disrupted by faulting. This zone has been mapped as a reverse fault (Plate 1) because of the apparent eastward juxtaposition of the fold limbs along this west-dipping structural discontinuity. The fault surface depicted on the map represents the base of a strongly deformed "golf ball" horizon in the Keeler Canyon Formation near the top of the zone. This horizon is folded by northwest-trending minor folds. This suggests that the fault zone may have existed prior to $F_2$ folding and was probably an original feature of the core of the San Lucas Canyon syncline. Subsequent refolding by $F_2$ folds and possible flattening during intrusion of the batholith have further deformed this structure. An estimate of slip direction or amount is not possible.

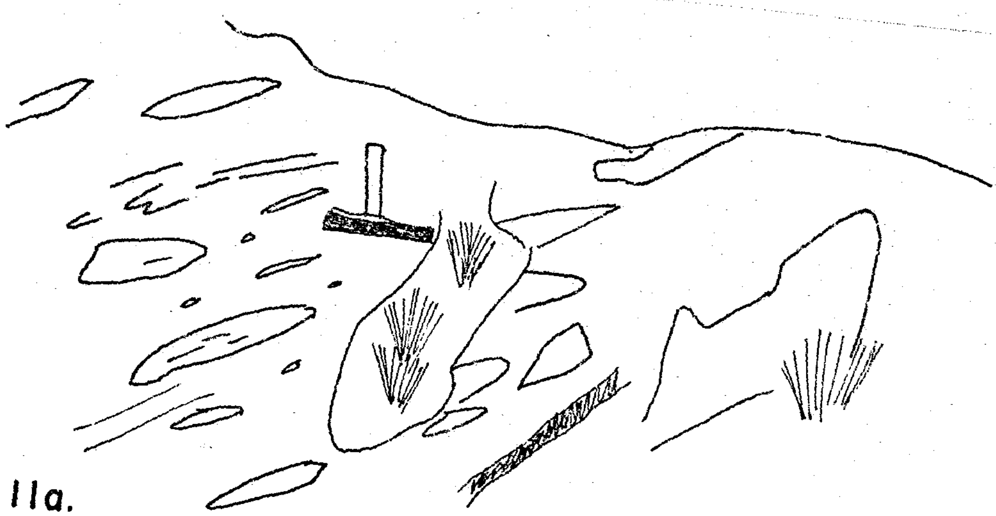
The existence of a major low angle thrust fault, the Conglomerate Mesa thrust fault (McAllister, 1956; Elayer, 1974), along the contact between the middle and upper members of the Owens Valley Formation at Conglomerate Mesa could not be substantiated. Inspection of the contact showed that some shearing is evident, but preservation of stratigraphic order and lack of substantial cutting out or repetition of the section suggests that only minimal slip is likely to have occurred. Shearing

Figure 11. Sketches from photos of structures related to reverse fault in core of San Lucas Canyon syncline in domain 5. Hammer handle is 28 cm.

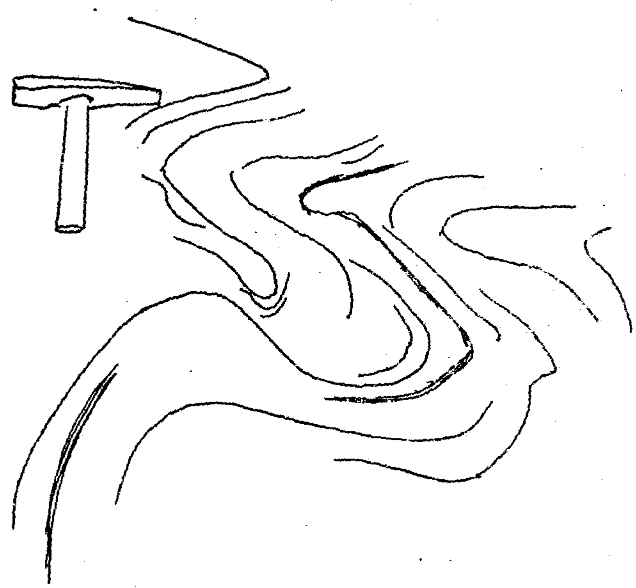
11a) Pod of brecciated limestone within fault zone. Clasts and matrix are stretched and boudinaged parallel to bedding.

11b) Disharmonic folds in "golf ball" bed of Keeler Canyon Formation. Folds are oriented $22^{\circ}$, S$55^{\circ}$ W.

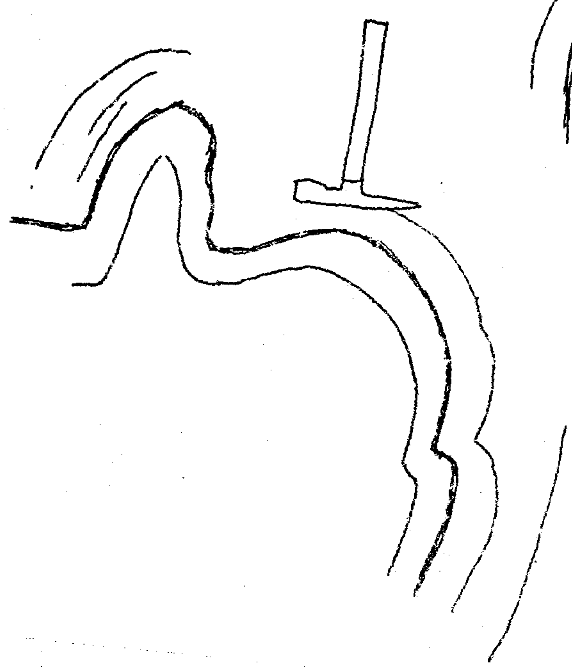
11c) Minor folds with axes oriented $25^{\circ}$, S $50^{\circ}$ E. View is to the north.



11a.



11b.



11c.

along the contact could well be the result of the significant contrast in competence between the middle shale and upper conglomerate members, such that stratal shortening during folding was accompanied by shearing within and along the top of the shale member. As a result the broadly folded upper member became detached, but no significant displacement occurred.

Northwest-trending high angle faults

Northwest-trending faults are common throughout the study area. These faults have an average trend of N. 45$^{o}$ W, ranging between N. 15$^{o}$ W and N. 70$^{o}$ W. Where these faults are exposed in Paleozoic limestone they are vertical or dip very steeply and are characterized by breccia zones 1 to 2 m wide. Stratigraphic separation on these faults, based on estimates from cross sections, is generally less than 200 m.

Aligned scarps in alluvium near the west margin of Lee Flat (Plate 1) indicates Holocene movement on some members of this fault set. Since these faults also control the northwest trend of the Nelson Range and the southern Inyo Mountains it is presumed that some are normal faults related to Cenozoic basin and range deformation. However, some northwest-trending faults are parallel to and locally include fractured and altered andesite porphyry dikes which are believed to be members of the Late Jurassic Independence dike swarm. Since these faults are not affected by pre-Middle Jurassic folds, a Middle to Late Jurassic age is indicated for at least part of the northwest-trending fault set. It is possible that there are two sets of northwest-trending faults of very different ages, or perhaps one set that formed in the Jurassic and were partly reactivated in the late Cenozoic.

Northeast-trending high angle faults

Faults which have trends between N $40^{o}$ W and N $50^{o}$ W occur predominantly in the Nelson Range. Where they are exposed along the east wall of San Lucas Canyon, these faults are vertical and characterized by breccia zones 2 to 5 m thick. The northern-most fault of this set in domains 8 and 9 shows 100 m of right-lateral strike separation of a fold hinge line and the intrusive contact. Elsewhere, a vertical component of movement is indicated on several of these faults based on stratigraphic separation in cross sections.

These faults truncate andesite porphyry dikes, and they both truncate and are truncated by northwest-trending faults. However, since the age(s) of the northwest-trending fault set is uncertain, the relative age of the northeast-trending faults could be as old as Late Jurassic, or as young as Holocene.

# DISCUSSION

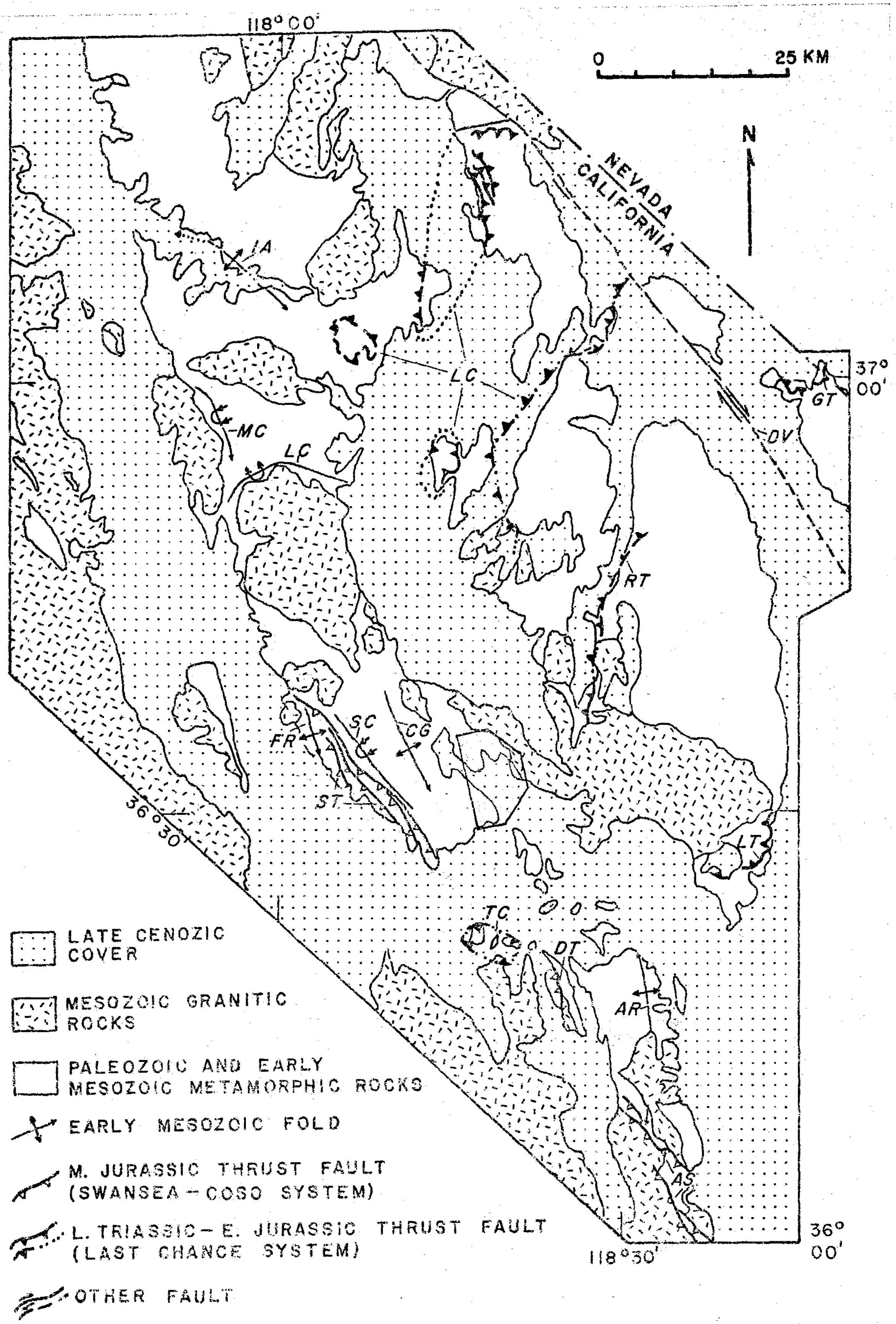
## REGIONAL MESOZOIC STRUCTURES

An overview of the regional geology is presented to illustrate the relationship of the study area to the major structures of Mesozoic age in adjacent areas of California. These structures are discussed in chronologic order.

In the ranges immediately east of the study area (fig. 12) are exposed imbricate, low angle faults of the Last Chance thrust system (Stewart and others, 1966; Burchfiel and others, 1970). Major faults in this system are the Last Chance fault in the Last Chance and Saline ranges (Stewart and others, 1966), the Racetrack fault in the northern Panamint Range and Quartz Spring area (McAllister, 1952, 1956), the Lemoigne fault in the Panamint Range (Hall, 1971) and the Grapevine fault in the Cottonwood Mountains (Reynolds, 1971). Also included in this system, based on style and relative age (Gulliver, 1976a) is the Talc City thrust fault (Hall and MacKevett, 1962) exposed south of the study area.

Faults of the Last Chance system are characterized by general northeast trends, initial shallow dips to the west and eastward tectonic transport measured in tens of kilometers. The locus of faulting commonly is along incompetent shaly Mississippian strata such as the Perdido Formation and Rest Spring Shale (Stewart and others, 1966). Eastward overturned folds are common beneath many of these fault surfaces. In Lead Canyon in the central Inyo Mountains (fig. 12), the overturned limb of a major northeast-trending, recumbent anticline (Nelson, 1967) is continuous with rocks in a portion of the upper plate

Figure 12. Index map to major Mesozoic structures in eastern California (modified after Gulliver, 1976a; Dunne and others, 1978). Study area shown in gray. Abbreviations are as follows: AR, anticline in Argus Range; CG, Cerro Gordo anticline; DT, Davis thrust fault; DV, Death Valley-Furnace Creek fault; FR, Front Ridge anticline; GV, Grapevine thrust fault; IA, Inyo anticline; LA, anticline in Lead Canyon; LC, Last Chance thrust fault; LT, Legmoine thrust fault; MC, Mazourka Canyon syncline; RT, Racetrack thrust fault; SC, Slate Canyon syncline; ST, Swansea thrust fault; and TC, Talc City thrust fault.



118° 00'
0
25 KM
N
NEVADA
CALIFORNIA
IA
LC
MC
LC
37° 00'
GT
DV
RT
SC
CG
FR
ST
36° 30'
LT
TC
DT
AR
AS
36° 00'
118° 30'
LATE CENOZIC COVER
MESOZOIC GRANITIC ROCKS
PALEOZOIC AND EARLY MESOZOIC METAMORPHIC ROCKS
EARLY MESOZOIC FOLD
M. JURASSIC THRUST FAULT (SWANSEA – COSO SYSTEM)
L. TRIASSIC – E. JURASSIC THRUST FAULT (LAST CHANCE SYSTEM)
OTHER FAULT

of the Last Chance thrust fault (Stevens, 1969).

These faults involve rocks as young as Permian and are truncated and intruded by the Hunter Mountain batholith. Burchfiel and others (1970) propose a Late Triassic to Early Jurassic age; however, Dunne and others (1978) believe that faulting may have begun by Middle Triassic time based on the timing of deformations in adjacent areas.

Northwest-trending folds of varied style and an associated southwest-dipping axial plane cleavage pervade this part of eastern California (fig. 12) and impart the predominant structural grain to the region (Dunne and others, 1978). Northwest-trending folds involve rocks as young as Middle and even Late (?) Triassic age and are intruded by rocks of the Hunter Mountain batholith. In the Talc City Hills, Gulliver (1976) has reported that the Talc City thrust and related folds are folded by northwest-trending folds. Based on these relationships, these folds postdate the Last Chance fault system and range in age from Late Triassic to Early Jurassic. Younger northwest-trending folds have been recognized in a few areas (Dunne and others, 1978), but they are not regionally significant.

To the west and south of the study area is situated the Swansea-Coso thrust system (Moore, 1974, 1976; Dunne and others, 1978). This system is comprised of an aligned series of northwest-trending, moderate to steeply southwest-dipping reverse faults. The system includes the Swansea thrust system in the southern Inyo Mountains (Kelley, 1973; Kelley and Stevens, 1975), the Davis and related thrust faults in the Darwin quadrangle (Hall and MacKevett, 1962; Gulliver, 1976a, b), and the Argus Sterling thrust fault in the Argus and Slate ranges (Moore, 1974, 1976).

These faults generally have a northeast slip direction, and slip amounts are estimated to be 1 to 3 km. The Argus Sterling thrust fault truncates northwest trending folds and its movement is bracketed by plutons having minimum ages (K/Ar) of 168 m.y. and 140 m.y. (Moore, 1976). Therefore this fault system is of Middle Jurassic age.

Northwest-striking faults and fractures extend diagonally across the southern Inyo Mountains, Argus Range and southern Slate Range. Dunne and others (1978) have proposed that these structures represent a broad shear zone which has accommodated a few kilometers of left-slip. These structures truncate the Swansea-Coso thrust system and are locally intruded by dikes of the Independence swarm. Therefore, these faults and fractures are of Middle to Late Jurassic age (Dunne and others, 1978).

The effects of Cretaceous deformation which have been recognized in the eastern Sierra Nevada to the west (Nokleberg and Kistler, 1977) and the eastern Mojave Desert and Nevada to the east (Burchfiel and Davis, 1971) are very poorly expressed in this part of eastern California. Sets of minor conjugate strike-slip faults which occur in the Inyo Mountains are known or inferred to be Late Cretaceous in age (Dunne and others, 1978).

The regional tectonic forces which produced most of the structures discussed above were the result of contractile strain behind the Andean-type arc which characterized the western margin of North America during the Mesozoic (Burchfiel and Davis, 1975). Heating and concurrent thermal weakening of the lithosphere associated with the evolving Sierra

Nevada batholith to the west made this region a preferred site for compressional failure in response to this contractile strain (Dunne and Gulliver, 1976).

RELATIONSHIP TO REGIONAL STRUCTURES

As previously discussed, the relative ages of folds and faults within the study area are only partially understood. Relating these structures to regional Mesozoic deformations presents two major problems.

First, the style of folds in the study area is largely dependent on lithology. Park (1969) has pointed out that fold style is commonly used as a criteria for the recognition and correlation of fold generations in fold belts, but that the main elements of fold style such as concentricity, angularity, disharmony, tightness and associated foliation are strongly dependent on lithology, and this method of fold correlation cannot be used by itself.

Second, the axial trends of superposed $F_1$ and $F_2$ folds in the Nelson Range are oriented at a high angle to one another, but the trends of folds in the study area in general display a wide range of orientations. Any attempt to group folds in the south half of the study area with $F_1$ or $F_2$ folds based solely on their trends would be arbitrary. In addition, fold orientation as a means of regional correlation of fold generations has little significance unless it can be established that the stress regime responsible for a particular generation of structures affected the entire region contemporaneously (Park, 1969).

As indicated in the preceding section, however, eastern California is characterized by several sets of Mesozoic faults and folds which have distinctive style and geometry. Whether or not these sets of structures represent distinct events or simply high points in more or less continuous deformation, available cross-cutting relationships indicate that recognized sets of structures formed throughout the region during a limited period of time.

Roberts (1977) points out that the absence of absolute or universal methods for structural correlation introduces a subjective element into regional structural synthesis. As a result, any proposed structural model should only offer the simplest account of all available evidence, including style and orientation, in a manner which is internally consistent. This philosophy is followed in the ensuing discussion.

$F_1$ folds involve rocks as young as Pennsylvanian and Permian (?) and are older than pre-Middle Jurassic $F_2$ folds. Although $F_1$ and $F_2$ folds could be as old as Permian, folds of pre-Mesozoic age have not yet been proven to exist south of the northern Inyo Mountains (Dunne and others, 1978). Thus, $F_1$ folds are herein proposed to be temporally and genetically related to the Last Chance deformation on the basis of their older age relative to northwest-trending $F_2$ folds.

The trend and vergence of $F_1$ folds is consistent with the generally eastward direction of tectonic transport inferred for Last Chance thrust faults. Refolding of $F_1$ by $F_2$ folds in the Nelson Range is similar to the sequence of events described by Gulliver (1976a) in the Talc City Hills where the Talc City thrust fault and related folds are refolded by younger northwest-trending folds. No faults of the Last

Chance system are recognized in the study area, but rocks in the Santa Rosa Hills and southeastern Inyo Mountains are interpreted to be continuous with the lower plate of the Talc City thrust fault (Gulliver, 1976a). In the central Slate Range and southern White Mountains northeast-trending, southeast-vergent folds are overprinted or truncated by northwest-striking structures (Moore, 1976; Dunne and others, 1978) and may also represent a structural sequence analogous to that found in the study area.

Dunne and others (1978) suggest that some of the northeast-trending folds in this region may be related to a zone of right-lateral faulting and oroflexure of Mesozoic age near the California-Nevada border (Albers, 1967; Buckley, 1974). However, structures within this zone are probably Late Jurassic (Buckley, 1974) in age whereas $F_1$ folds in the Nelson Range pre-date the Early to Middle Jurassic Hunter Mountain batholith.

Other folds in the study area which may be cogenetic with $F_1$ folds are the south-and southeast-plunging folds in domain 11 and 12. The style and orientation of the recumbent syncline in domain 12 are similar to that of the San Lucas Canyon syncline (compare figures 5a and 9), although it is of much smaller areal extent.

The relationship of $F_1$ folds to the "incongruous" northeast-trending folds mapped by Elayer (1974) west of Conglomerate Mesa is unclear. These "incongruous" folds are described as superposed on northwest-trending folds and may represent a minor, younger fold event.

$F_2$ folds in the Nelson Range, as well as most of the folds in domains 13, 14 and 15 in the southeastern Inyo Mountains, are correlated with other northwest-trending folds of Late Triassic to Early Jurassic age in this part of eastern California. This correlation is based on: 1) their general northwest trend, 2) their locally well-developed southwest-dipping axial plane cleavage and 3) their pre-Hunter Mountain age.

The diversity of northwest trends exhibited by these folds might be interpreted as suggesting the presence of several sets of subparallel northwest-trending folds in the study area such as proposed by Elayer (1974) in the Inyo Mountains to the west. However, there is no evidence of refolding or cross-cutting relationships involving northwest-trending folds, and regional age relationships indicate that most northwest-trending folds formed during a discrete deformational phase of limited duration. The diversity of trends may reflect gradual changes in stress patterns during this deformational phase as predicted by the progressive strain model of Ramsey (1967).

Faults of the Middle Jurassic Swansea-Coso thrust system have not been recognized in the study area. The possibility exists that some of the northwest-trending folds may be temporally related to this fault system, but this is not considered likely for the reasons discussed above.

Some northwest-striking faults in the study area may be of Middle to Late Jurassic age. If so, they may be correlative with other faults and fractures of similar age and orientation which have been reported in this region (Dunne and others, 1978).

Andesite porphyry dikes of presumed Late Jurassic age are seen to cut all major Mesozoic structures discussed above, and they are cut only by Cenozoic basin and range faults. Therefore, all Mesozoic structures in the Nelson Range and southeastern Inyo Mountains apparently predate the Late Jurassic.

## SUMMARY OF MESOZOIC HISTORY

The structural history recorded in the Nelson Range and southeastern Inyo Mountains indicates that this area shares a similar history of superposed Mesozoic deformations as reported in adjoining areas of eastern California.

The oldest deformation produced a set of folds ($F_1$) which are northeast-trending and either upright or overturned to the southeast. These folds are interpreted to be temporally and genetically related to the Last Chance thrust fault system of Middle Triassic (?) to Early Jurassic age.

$F_1$ folds are locally refolded by a younger set of folds ($F_2$). $F_2$ folds consist of north- to northwest-trending, upright and northeast-overturned folds with southwest-dipping axial plane cleavage. $F_2$ folds predate the Early to Middle Jurassic Hunter Mountain batholith and they are correlated with other northwest-trending folds of pre-Middle Jurassic age in eastern California.

Emplacement of the Hunter Mountain batholith locally has appressed folds and altered their trend near the intrusive contact in the Nelson Range, but intrusion did not produce any major folds in this area as suggested by previous workers.

Some of the northwest-striking faults in the area may be of Middle to Late Jurassic age and could be related to a northwest-trending zone of faults and fractures of similar age which extends across the region.

These structures were intruded by the Independence dike swarm in the Late Jurassic.

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