PROCESSES OF EXTENSIONAL TECTONICS

by

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B.S., University of Southern California
(1978)

SUBMITTED TO THE DEPARTMENT OF
EARTH AND PLANETARY SCIENCES IN
FULFILLMENT OF THE REQUIREMENTS
FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

August 1982

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June 2, 1982

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L00429

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Submitted to the Department of Earth and Planetary Sciences on June 2, 1982 in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Geology

ABSTRACT

Traditional views of extensional tectonics postulate a faulted brittle layer directly overlying a plastically extended or intruded substratum. However, where lower bounds to normal fault systems are actually exposed, in situ accommodation of upper plate extension is not observed, which has led some to interpret the basal faults as soles to large surficial gravity slides.

The similar controversy over thin-skinned compressional belts (gravity slide vs. in situ accommodation) is geometrically resolved by the concept of a rigid underthrusting slab, the thrust belt being the surface accommodation of convergence between two rigid slabs separated by a shallow fault. Thin-skinned extension may be the result of divergence between such slabs, surficial extension occurring at the "feather edge" of the upper slab.

This hypothesis does not require the existence of a mid-crustal transition zone between "brittle" normal faulting and "ductile", taffy-like stretching. Indeed, taffy-like stretching may not occur at all within the lithosphere, surficial extension being transformed directly to convecting asthenosphere by a decollement between two lithospheric slabs.

Large, very low-angle normal faults dominate highly extended terranes, and both listric and planar normal faults are common components of their hanging walls. The very low-angle normal faults may have displacements from a few kilometers up to several tens of kilometers and I regard their hanging walls as extensional allochthons, analogous to (but with opposite sense of movement) thrust fault allochthons. Differential tilt between imbricate fault blocks suggests listric (curved) geometry at depth, whereas uniformly tilted blocks are more likely to be bounded by planar faults. Chaos structure, a deformational style widely recognized in the Basin and Range province, is an expression of large-scale simple shear on very low-angle normal faults.

Large scale extension in the Basin and Range province can be demonstrted independent of assumptions about cross-sectional normal fault geometry. Strike-slip faults in the southern Great Basin separate areas of Cenozoic upper crustal extension from relatively stable tectonic blocks. Linear geologic features, offset along the Garlock fault, Las Vegas Valley shear zone, and Lake Mead fault system, allow reconstruction of the southern Great Basin to a pre-extension configuration. The Sierra Nevada, Mojave Desert, Spring Mountains, and Colorado Plateau are treated as

stable, unextended blocks which have moved relative to each other in response to crustal extension, with the Spring Mountains held fixed to the Mojave block. Reconstruction of these faults indicates a minimum of 65% extension (140 km) between the southern Sierra Nevada and Colorado Plateau.

In the Mormon Mountains, southern Nevada, detailed geologic mapping of a west-directed extensional allochthon (Mormon Peak allochthon) with at least 15 km displacement provides insight into relationships between hanging wall tilt geometry, transport direction, and footwall structure of extensional allochthons. The strike of tilted strata is consistently north-northwest throughout the Mormon Peak allochthon, but the allochthon is divisible into domains of alternating tilt direction with boundaries trending both perpendicular and parallel to strike. Appoximately half of the area contains west-southwest tilts, i.e. tilts in the same direction as transport. Although the strike of tilted strata has repeatedly been shown to be pependicular to the orientation of transport of the allochthons, many workers have used dip direction as the sole indicator of their sense of transport, assuming that tilt is always directed away from transport. Relationships observed within the Mormon Peak allochthon, in which the sense of transport is known from its westward sense of downcutting in both hanging wall and footwall, indicate that this is not always true.

Favorable fenster-klippen distribution of the Mormon Peak allochthon also demonstrates that its complex pattern of tilt domains bears no relation to the tectonically inert footwall. Similar tilt patterns observed throughout the Basin and Range province may therefore have developed within single, unidirectionally-transported extensional allochthons, uninfluenced by rigid footwall blocks. Comparable geometric relationships are well-known from compressional tectonic regimes. In both cases, it appears that complex deformational systems in the upper crust are not guides to subjacent strain patterns in the middle and lower crust.

In addition to these extensional features, a newly discovered major thrust, the Mormon thrust, is exposed for a distance of 35 km parallel to its transport direction. Throughout this distance, the hanging wall is detached in strong dolomites of the Bonanza King Formation, just 100-150 m above the weak Bright Angel Shale. In the western 10 km of exposure, the footwall is detached in massive crinoidal limestone of the Monte Cristo Formation, only 50-100 m below relatively thinly bedded cherty limestones of the Bird Spring Formation, then ramps quickly to the east onto Mesozoic clastic rocks. Apparently both the hanging wall and footwall broke the "hard way", and define a 45 km-wide zone of initial breakage of the structurally lowest Sevier-age thrust in which ramp-flat goemetry is strikingly independent of stratigraphic anisotropy. Given the conditions at the end of Mesozoic time of a major west-dipping fracture, one might expect the formation of gently west-dipping extensional detachments to coincide with the older decollement. Instead, they ignore the Mormon thrust and cut into Precambrian crystalline rocks only 1-2 km beneath it.

Thesis Supervisor: Dr. B. Clark Burchfiel

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ACKNOWLEDGEMENTS

To adequately acknowledge the many people who made the completion of this thesis possible is a dubious task at best. To the extent that some people's contributions may have been overlooked, please forgive me.

Firstly I thank Clark Burchfiel, who originally suggested that the Mormon Mountains, and later, the field of extensional tectonics, might be a good thesis topic . Clark's example as both a geologist and a human being has been truly inspiring.

Bill Brace, in addition to teaching me a lot about rock mechanics, was able to inspire motivation in me during the attitudinal doldrums of my first summer in the field. Keep up the good work, tiger. Peter Molnar, in addition to footing half of my bills for the first three years of the study showed me that science was mainly a matter of delimiting ignorance. John Sclater, who accompanied me into the field on two occasions, provided financial support, gave me much to think about in terms of the physics of extension, and taught me what "F.R.S." meant.

Critical to my graduate career is a group of peers with whom I could confide questions like "What <u>is</u> a thrust fault, anyhow?". For keeping me from making too much of a fool out of myself outside the Building 54 womb, I thank my fellow graduate students, particularly P.L. Guth, J.M. Bartley, C.S. Cameron, J.E. Spencer, J. Sharry, J.H. Willemin, G.J. Axen, K.V. Hodges, P. Crowley, D.W. Klepacki, J.D. Walker, and L.H. Royden, the "clones", for their helpful scientific contributions and personal friendship.

Interactions with colleagues outside M.I.T. provided much food for thought, and I would particularly like to thank A.W. Bally, R.K. Dokka, G.P. Eaton, P.B. Gans, W.B. Hamilton, J. Helwig, F.G. Poole, S.M. Richard,

S. Snelson, and M.L. Zoback for stimulating discussions. J.L. Anderson, R.E. Anderson, R.G. Bohannon, G.A. Davis, G.H. Davis, E.G. Frost, W.S. Jefferson, E.L. Miller, S.J. Reynolds, T.J. Shackelford, and J.H. Stewart, and B.W. Troxel led me through areas they have mapped in the Basin and Range, in addition to providing invaluable discussion.

R.E. Anderson and R.G. Bohannon arranged helicopter tours for me through areas they have mapped adjacent to the Mormon Mountains, and have greatly influenced my ideas on strike-slip faulting, extension, and thrusting in the Utah-Nevada-Arizona corner.

Jeanne Harris, Mike Price, and Walter Neibuhr of Mobil Oil Corporation provided stimulating discussions concerning Mobil's deep test well near Overton, Nevada, and helped arrange partial financial support for my field work from Mobil. David Drowley of Amoco Production Company arranged for 10 days of helicopter time in which we mapped, in reconnaissance, much of the Tule Spring Hills area, and gave me an opportunity to get a glimpse of the poorly accessible terrane in the northern Mormon Mountains.

D.L. Weide and Walt Raywood of the University of Nevada at Las Vegas were particularly helpful during the course of the field work. Use of the faciities at UNLV made logistics infinitely simpler than they might have been otherwise.

Field assistants S.C. Semken, D.L. Olgaard, and J.M. Stock, M.S. Beaufait and J.D. Walker ably provided safety, good spirit, and good ideas during those hot summer afternoons, and their help is gratefully acknowledged. I hope they learned as much from me as I did from them. Few moments in the history of intellect rival Beaufait and I, a quarter-mile apart on either side of a resonant gorge, in a 45-minute shouting match over fault geometry.

Joann Stock, in addition to helping me draft all those figures in the final tense weeks before defending, was a source of comfort and personal example throughout the study.

Finally, my family in Whittier, California provided moral support throughout the study, and I cannot begin to express my thanks, especially for all those trips to the ill-located Los Angeles airport. Thank you all!

PREFACE

An important question facing structural geologists today is the magnitude and structural expression of extension within the continental lithosphere. Because extended areas eventually thermally subside and become covered with thick blankets of sediment rather than being uplifted to form mountain belts, our understanding of extensional processes has lagged far behind that of compressional processes. Where ancient rifts are incorporated into mountain belts via compression, structures related to the rifting process are extremely difficult to characterize because of the compressional overprint. However, a few highly extended regions, such as the Basin and Range province, have not thermally subsided, and thus offer ample opportunity to study extensional processes, which was the purpose of this dissertation. I have divided it into four chapters, each corresponding to a paper which either has already been published or will be published in the near future. For this reason, some of the material overlaps, and I sincerely apologize for any inconvenience this may cause the reader. Not all of the work that went into the papers is mine, and it is thus appropriate here to detail my contributiuons to them.

The first chapter is a short "brainstorm" paper I published in Nature in June of 1981. It explores an analogy between compressional and extensional tectonism in an attempt to explain the origin of a highly enigmatic set of Tertiary dislocational terranes in the Basin and Range. I accept the burden of responsibility for the concepts presented therein, in that at least one citation of them thus far has used the term "outrageous", and far less polite terms were used in review.

The second chapter represents an attempt by me and Clark Burchfiel to start from ground zero on cross-sectional normal fault geometry and establish a few rules of thumb for recognizing various types of normal faults in the field as well as dispel some of the ancient myths which shroud the field of extensional tectonics. As of this writing it is in press with the Journal of Structural Geology. I did most of the categorizing, calculating, drafting and writing but much of the kindling of ideas and the savoire faire of the Introduction and Conclusions are credited to Burchfiel.

The third paper addresses the question, How much has the Basin and Range extended? By means of province-wide palinspastic reconstruction of major strike-slip faults across the narrowest part of the province (only 360 km), a lower bound of about 140 km of extension may be deduced -- an estimate far in excess of that perceived by most workers. Although I was the one to conceive of the province-wide reconstruction, much of the geologic synthesis of the individual strike-slip faults is credited to my coauthors Peter Guth, Clark Burchfiel and Jon Spencer.

The fourth and largest chapter represents my attempt as a graduate student to generate some data rather than telling everyone else what their data actually means. It is a detailed structural and stratigraphic study of 200 square kilometers of Basin and Range geology which, as will be obvious to the reader, had a great of influence on the first three chapters, but as might be expected for a paper containing real data, will be the last reach press. Most of the mapping was done by me, but substantial amounts were done by Mark Beaufait and Doug Walker, who will share authorship on the manuscript when and if it ever gets out. The stratigraphic sections were measured and sampled for conodonts in conjunction with Forrest G. Poole of the U.S. Geological Survey, and the descriptions of the rocks presented here are based on his carefully taken field notes. All of the writing, cross-section work, data analysis and interpretation are mine, notwithstanding endless critical input by Walker and Beaufait.

Reprinted from Nature, Vol. 291, No 5817, pp 645-648, June 25 1981

Low-angle normal faults in the Basin and Range Province: nappe tectonics in an extending orogen

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Cenozoic normal fault mosaics bounded beneath by a basal fault in the Basin and Range Province (BRP) have traditionally been described either in terms of large-scale surficial gravity sliding or by some form of *in situ* lower plate accommodation. I suggest here that these areas may be an extensional analogue to thin-skinned compressional orogens, a process which may even dominate active BRP tectonics.

These terranes consist of a mosaic of rock sheets bounded by high- and low-angle normal faults[1,2] which resemble (and have been misinterpreted as) compressional nappe piles. In many areas, the normal faults 'bottom out' into a single, subhorizontal fault zone below which either minor or no tectonism has occurred[1-6], or a roughly synchronous, more ductile style of deformation is present[7,8]. The minimum areal extent of the basal faults is typically measured in thousands of square kilometres. For example, in east-central Nevada, normal fault systems developed in Palaeozoic miogeoclinal strata, Mesozoic and Tertiary intrusives, and Tertiary volcanics end abruptly at a basal fault (Snake Range decollement[1], Fig. 1) accompanied by a thin (0–100 m) zone of tectonite marble. Offsets on the basal and higher faults are difficult to contrain, but Nelson reports a 20-km minimum horizontal offset for one fault alone in the Deep Creek Range[9]. My palinspastic reconstruction of the Northern Egan and Cherry Creek ranges (Fig. 2) indicates a 16.5-km offset for one sheet in the mosaic, and the per cent increase in original width of the area is ~300%. As the basal and higher faults in Fig. 1 are commonly subparallel to bedding and often cut out thousands of metres of section, total offsets are probably at least in the range of several tens of kilometres. Most authors except one[10] who have mapped in the area outlined in Fig. 1 agree that the transport direction of the allochthonous sheets was eastwards. The faults also tend to cut consistently downsection to the east[11,12], (Fig. 2). This terrane apparently has every attribute of a thin-skinned compressional orogen (basal decollement, far-travelled allochthons, consistent direction of section transgression and transport) except (1) the low-angle faults habitually omit rather than repeat section, (2) associated high-angle faults are predominantly normal instead of reverse, and (3) folding is of subdued importance[13]. Though early workers attempted to correlate these faults with Mesozoic thrusting in the Sevier belt to the east[5], there is considerable evidence to support a Tertiary, extension-related origin for most of the deformation[13-16].

Models for these areas must account for three observations: (1) the basal and higher faults predominantly represent extension; (2) the rocks below the basal faults are tectonically inert during transport of the allochthonous sheets and high-angle fault blocks; and (3) the areal extent and transport distances of the allochthons are large. Current models fall into two categories. The first is a localized gravity-slide model in which domal upwarps shed their cover in a megalandslide fashion[16-18] (Fig. 3*a*). The second invokes penetrative ductile stretching and/or igneous intrusion to drive extension in the immediately overlying brittle fault mosaic[2,7,8,19,20] (Fig. 3*b*). Landslide models are weakened by the scarcity of geometrically required areas of shortening correlative in age, size and direction of transport with the extensional terranes. Thus, the model agrees with observations (2) and (3), but not with observation (1). *In situ* ductile stretching or intrusion models lose appeal when it is noted that most metamorphic fabrics and igneous bodies (if any) beneath the basal faults are of inappropriate age or geometry to be rigorously linked to normal faulting[4,6,21,22]. Thus, the model is in harmony with observations (1) and (3), but not with observation (2).

The key to the problem may lie in our understanding of compressional orogens, in which the horizontal movement of thin sheets of rock above an underthrusting slab may occur many tens (even hundreds) of kilometres from the nearest possible site of coeval lower plate shortening. Applying this concept to extensional tectonics to explain relationships observed in the BRP, I suggest that the basal faults 'root' into the crust (and perhaps the mantle) at a low angle, thereby serving as a narrow zone of decoupling between two thin sheets of rock (Fig. 3*c*). Towards the thick end of the upper plate, extensional faulting may be negligible or absent, but towards the thin end it loses its ability to remain coherent and becomes a thin-skinned extensional fault terrane. Lower plate extension may therefore be far removed horizontally (perhaps a distance many times the areal width of upper plate normal faulting) and at much greater depth than the thin-skinned extensional belt. This hypothesis allows for a rigid autochthon and eliminates the need for immense terranes of coeval shortening, thus satisfying all three observations.

As shown in Fig. 3*c*, rock sheets (nappes[23]) may be carried along the basal fault and accreted to the autochthon, kinematically similar to nappe emplacement along the soles of overthrust masses in compressional belts. Note that the nappes in Fig. 3*c* have converged on one another, a relationship that superficially resembles shortening. However, the amount of movement between them is actually a measure of crustal divergence. If the faults had developed in a pristine sedimentary sequence, the fault between the nappes would place younger rocks on top of older rocks, a direct consequence of the fault

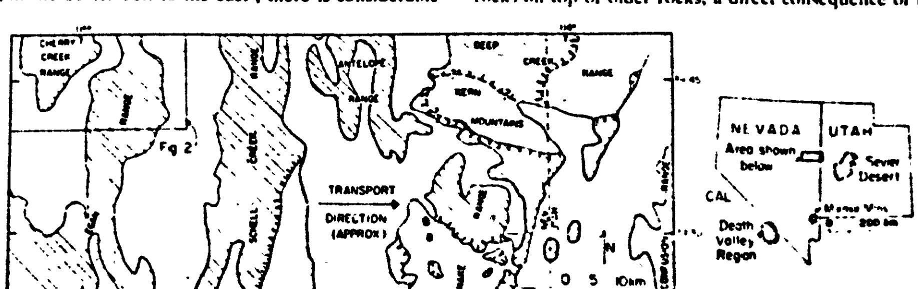

Fig. 1 Location map and tectonic interpretation of east-central Nevada, showing autochthon (stippled), basal fault zone (line with tick marks, queried where uncertain) and normal fault-mosaic (hatched). Basal fault in the Schell Creek Range is locally tilted steeply westwards, implying continued deformation after it ceased moving, possibly on a lower basal fault. Tilting and faulting in the Northern Egan Range may be related to this second event rather than to movement on the basal fault shown here.

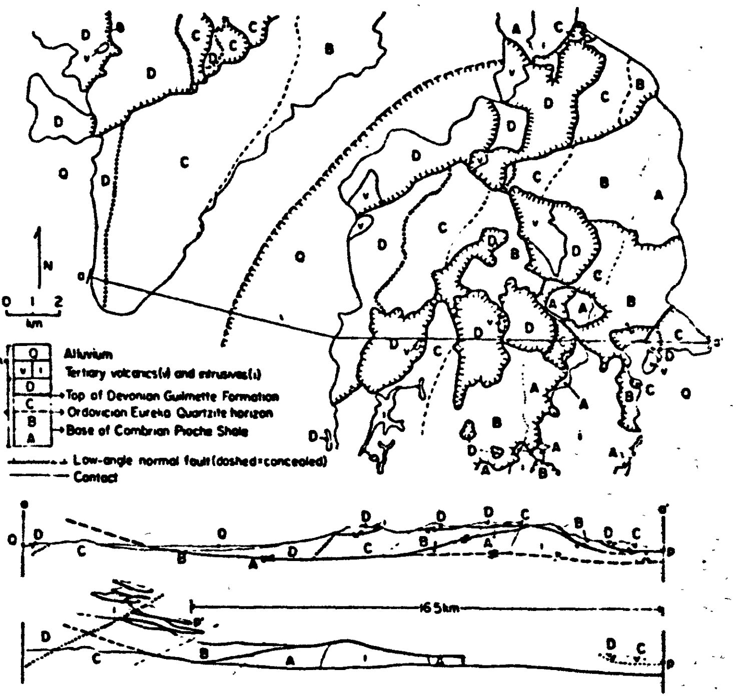

Fig. 2 Highly simplified map, cross-section and palinspastic reconstruction of the area indexed in Fig. 1. Reconstruction does not remove tilting, which was probably synchronous with faulting. Armstrong[15] interpreted the sheets as having formed exclusively by high-angle normal faulting, then subsequently rotated to the horizontal. Geology by Fritz[11].

rooting in the same direction as transport. Had the same juxtaposition occurred along a fault which rooted opposite its transport direction, movement between the nappes would be an expression of true crustal convergence. Large, rooted low-angle normal faults may form zones of ductile shear along deeper segments. If the displacement is large enough, these zones may shallow and be tectonically reworked in brittle conditions[21]. This scheme of nappe emplacement, combined with superjacent 'domino-style'[24] and listric normal faulting, are major mechanisms of extension above the basal fault. Although some aspects of this model are not new[24,26], its application to low-angle normal fault terranes in the BRP has been neglected in favour of gravity and *in situ* models.

Distinction between low-angle normal fault terranes and active BRP tectonics has been strongly emphasized by some workers[27], and others have related the apparent difference in style (low-angle compared with high-angle) to plate tectonics[28]. However, the neotectonics of the BRP are apparently sufficiently ill-characterized at depth to entertain the hypothesis that rooted, low-angle normal faults are an important mode of Plio-Quaternary extension, perhaps the fundamental control of its famous topography. For example, the Amargosa Chaos[4], a striking example of low-angle normal faulting[1], involves mid- to late-Pliocene volcanics, and hence was active well within what some consider to be a province-wide period of high-angle faulting. Further evidence suggesting that the present

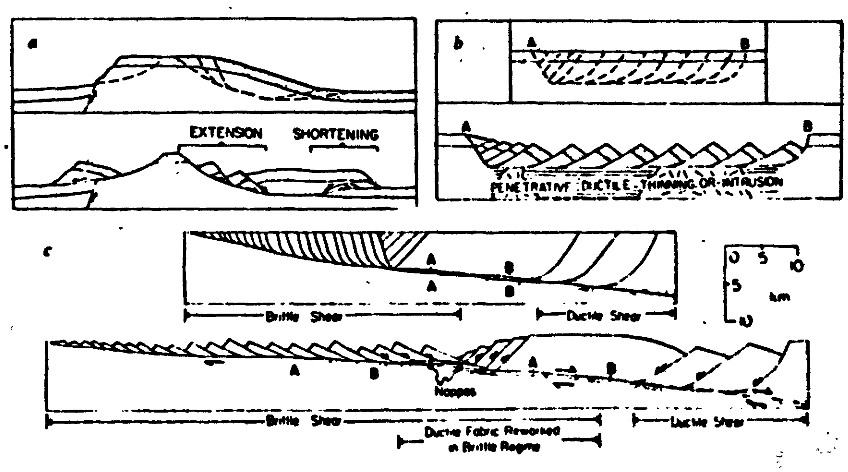

Fig. 3 First-order kinematic schemes to generate low-angle normal fault mosaics above a basal fault. *a*, Megalandslide model; *b*, *in situ* ductile stretching or intrusion model; *c*, rooted, low-angle normal fault model.

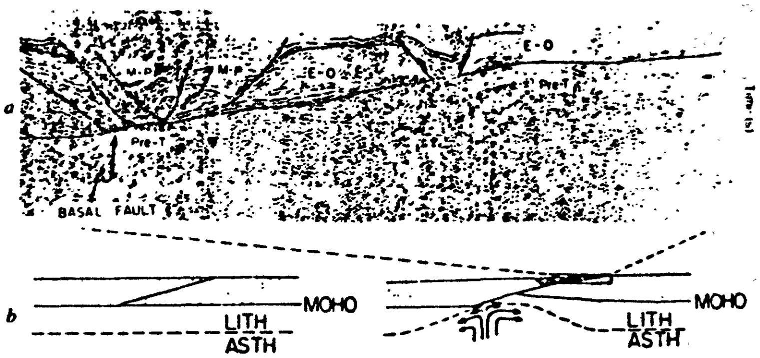

Fig. 4 *a*, North-east trending, 40-km long two-way time section through the Sevier Desert area[29]. Pre-T, pre-Tertiary. E-O, Eocene-Oligocene. M-P, Miocene-Pliocene. *b*, Lower crust-mantle extrapolation of faults such as that in *a*. Note that the geometry of the Moho is indistinguishable from how it would look if the crust had penetratively stretched. (Seismic line reproduced with permission of the Rocky Mountain Association of Geologists.)

mechanism may control active BRP tectonics comes from seismic reflection profiles of the Sevier Desert area[29]. Figure 4*a* and three other profiles reveal a minimum area of $\sim$5,000 km^2 underlain by a gently west-dipping low-angle normal fault, above which Miocene-Pliocene basin fill and Quaternary basalt flows are offset by normal faults[29-31].

The basal fault in Fig. 4*a* has been interpreted as a reactivated thrust fault[29], a common interpretation of low-angle normal faults situated in older thrust terranes[25,32,33]. However, recent studies of some exhumed thin-skinned extensional areas suggest that pre-existing anisotropies are not a prerequisite for low-angle normal faults[4], and thrust faults, if present, need not control their geometry. In the Mormon Mountains of southern Nevada (Fig. 1), a Miocene low-angle normal fault event is superposed on a system of Mesozoic thrusts, a situation geometrically identical to that in the Sevier Desert[34]. The normal faults there truncate the thrusts at high angle, and clearly involve Precambrian crystalline rocks of the thrust autochthon. Based on this study, the normal faults in the Sevier Desert may be unrelated to Mesozoic thrusts prominent in the surface geology of the area.

Figures 3*c* and 4*a* suggest that conservative estimates of province-wide extension (10–30%) based on its proportionality to regional tilts of Tertiary strata may have no meaning[18], because large, horizontal translations of rock masses may occur without appreciable rotation of strata (as well as opposite tilts within the same allochthon). The present model therefore compliments larger estimates (50–100%) deduced by strike-slip fault reconstruction (work in preparation) and crustal thickness arguments[20].

Another critical aspect of the model is that it presents an alternative to penetrative ductile thinning or intrusion to accommodate extension at crustal (and mantle) levels not exposed in the BRP. Whereas igneous intrusion may increase the width of the crust, it cannot thin it; at best, it can retain its original thickness, if the intrusions are vertical dykes. Alternatively, it can actually thicken, if the intrusions are sill-like, a property of the large early- to mid-Tertiary batholiths of southern Arizona[22]. Because intrusion cannot therefore thin deeper crustal levels, it is difficult to envisage it as a major mechanism of BRP extension. Although taffy-like stretching of crustal dimensions does not encounter these geometrical difficulties, there is no compelling evidence to suggest that it is a major process at depth. An attractive alternative, consistent with surface processes, is that large, rooted normal faults extend at a low angle deep into the lithosphere, and that extension at the surface is due to discrete shear between large coherent sheets (Fig. 4*b*). Pull-apart may then ultimately be accommodated by asthenospheric convection rather than any form of stretching.

I thank colleagues at MIT for discussions. This work was supported by NSF grant EAR 7713637 awarded to B. C. Burchfiel.

Received 18 December 1980; accepted 17 March 1981

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CHAPTER II. MODES OF EXTENSIONAL TECTONICS

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Abstract - Although hundreds of papers have been devoted to the geometric and kinematic analysis of compressional tectonic regimes, surprisingly little has been written about the details of large-scale strain in extended areas. We attempt, by means of quantitative theoretical analysis guided by real geological examples, to establish some ground rules for interpreting extensional phenomena. We have found that large, very low-angle normal faults dominate highly extended terranes, and that both listric and planar normal faults are common components of their hanging walls. The very low-angle normal faults may have displacements from a few kilometers up to several tens of kilometers and we regard their hanging walls as extensional allochthons, analogous to (but with opposite sense of movement) thrust fault allochthons. Differential tilt between imbricate fault blocks suggests listric geometry at depth, whereas uniformly tilted blocks are more likely to be bounded by planar faults. The tilt direction of imbricate normal fault blocks within large extensional allochthons is commonly away from the transport direction of these sheets, but in many cases tilts are in the same direction as transport, thus limiting the usefulness of the direction of tilting as a transport indicator. The presence of Chaos structure, a structural style widely recognized in the Basin and Range Province, implies large-scale simple shear on very low-angle normal faults, and does not necessarily form as a result of listric faulting.

INTRODUCTION

Work in extensional terranes has shown that they contain a great variety of faults active during deformation. Extensional faults can be grouped into two broad categories: (1) those which produce extension accompanied by rotation of beds, and in a subgroup are additionally accompanied by rotation of the faults; and (2) those which produce extension without rotation of faults or beds. In this paper we present an analysis of the geometric and kinematic properties of these fault types and examples of them from extensional terrains, principally the Basin and Range Province of the western United States. The Basin and Range Province is important in the study of the rifting process because it has not thermally subsided and been covered by post-rift sediments, it is well exposed and locally well mapped, it contains a great variety of extensional fault-types, and finally because vertical movements and erosion have been substantial a great variety of structural levels are exposed within the extensional complexes. We believe that the analysis of faults and extensional features in this region can be applied to other extensional terranes, particularly to passive continental margins.

Whereas in most extensional terranes the amount of extension is poorly constrained, there are regions within the Basin and Range Province where a minimum amount of extension can be measured without making any assumptions about cross-sectional fault geometry. Such analyses indicate province-wide extension of 64-100% and local areas which have extended well over 100% (e.g. see Hamilton & Myers 1966, Davis & Burchfiel 1973, Hamilton 1978, Wernicke _et al_. 1981). Extension of this order has been suggested for continental margins, but their fault geometries are less well-known. Our analysis indicates that certain fault types (or a combination of them) can easily produce extension of this magnitude, but these are not the ones commonly assumed to have produced the extension. The conclusions we reach are that widespread imbricate normal-fault blocks (Anderson 1971) and subjacent, large low-angle normal faults provide an explanation for the large magnitudes of extension observed. Listric normal faults function to relieve space problems between families of planar fault blocks, but are not the only contributor to the extension. The geometric and kinematic models we present are testable in the field and by geophysical techniques, especially siesmic reflection profiling.

EXTENSIONAL MECHANISMS

We divide extensional structures into two broad classes: rotational and non-rotational (Table 1).

Rotational extension

Rotational extension is defined as a kinematic mechanism in which extension occurs predominantly by progressive rotation of geological features (hereafter called beds). There are two possible fault geometries in this class: planar and listric.

Planar geometry is depicted in its simplest form in Fig. 1 (all models presented here are highly simplified and assume no penetrative deformation, pressure solution or bedding plane slip). As the rock mass is extended, both the faults and the beds rotate (Emmons & Garrey 1910) (Table 1). The simple calculation the basis of which is presented in Fig. 2 (Thompson 1960) allows us to determine the amount of extension if the attitudes of the faults and beds are known. Fig. 3 is a convenient graphical representation of the relationship derived in Fig. 2. Because field examples of rotational extension are invariably more complex than the geometry shown in Fig. 1, it is possible to use Fig. 3 only for approximate determinations. Accurate determination of extension in any given area can be done only by palinspastic restoration of geological maps and cross-sections.

A simple rotational listric fault (Table 1) is shown in Fig. 4. Extension occurs by the separation of two blocks on a curved surface. Since hanging wall strata move down a curved surface convex toward the footwall strata and maintain a constant orientation relative to that surface, listric faults produce differential tilt between hanging wall and footwall, that is, bedding dips are steeper in the hanging wall than in the footwall. Thus, a series of imbricate listric fault blocks should display successively steeper tilts as one traverses the blocks in the direction of downthrow (Fig. 5), whereas a row of planar fault blocks should all be tilted the same amount. Another distinction between rotational planar and listric faults is that planar faults must rotate with bedding whereas listric faults may remain fixed, rotation occurring only if their footwalls are rotated by structurally lower faults. Therefore, two groups of rotational faults may be distinguished, one in which both faults and beds rotate and includes both listric and planar fault geometry, and a second in which the faults remain fixed while the beds rotate, a situation restricted to listric fault geometry (Table 1).

A geometrically possible situation in which both faults and beds rotate but which is difficult to classify in terms of planar or listric faulting is shown in Fig. 6. The extreme displacement on a series of imbricate listric fault blocks (fault displacement being about the same as the length of the block) may result in a series of uniformly-tilted planar fault blocks. Even though the faults were initially listric, calculation of extension can be done assuming a planar geometry (Figs. 2 and 3), because the final geometry is that of a series of sub-planar blocks.

Consider a situation similar to that illustrated in Fig. 5 in which the lowest block is rotated on a listric fault but the remainder of the blocks rotate on planar faults above a basal detachment surface (Fig. 7). It is important to note that even though the planar faults may curve into the detachment surface, rotation and thinning of the mass above it is not a result of listric faulting. Curvature of the faults near the base of the blocks and the 'gap' (Fig. 1) created by rotation of the planar fault blocks above a detachment surface are space problems which in real geologic situations are accommodated by small-scale faulting, pervasive brecciation, and possibly plastic flow. The key diagnostic feature of fault geometry is differential tilt between blocks, not necessarily the geometry of the fault near a basal detachment surface.

Syntectonic sedimentary deposits which show increasing dip with age, commonly referred to as growth fault deposits, may develop in any setting involving rotational normal faulting. These deposits always dip toward the fault, except near the fault at the depositional surface where they may dip away from the fault.

A hypothetical situation shown on the geological map and section of Fig. 8 demonstrates the importance of diagnosing fault geometry. A low-angle normal fault dipping about 5° east offsets a section of 41 Ma volcanic rocks (TV1) and is overlain unconformably by a near-horizontal section of volcanics (TV2) dated at 39 Ma. Because the dip of bedding, and hence net rotation, of the two blocks is equal, the fault is best interpreted as planar, and its projection above the cross-section would be a straight line. However, if one were to assume that the fault was the flat portion of a listric fault and assumed it to steepen above the section, the estimate of extension represented by the fault would be considerably less than that predicted by a planar fault geometry (for an application see Le Pichon & Sibouet 1981). To emphasize this point, we have constructed a geometric model of a listric fault, shown in Fig. 9. The model assumes that the angle between bedding and the fault surface remains constant during deformation, and that curvilinear segments of the diagram are circular. The percent extension is calculated by comparing the distance between A and B with the length of a. By expressing the extension in terms of the dip of the beds next to the fault and the dip of the fault, we may compare the listric model with the planar model derived in Fig. 2. The result, shown in Fig. 10, demonstrates that for a given maximum stratal rotation and fault dip, the listric geometry yields far less extension than the planar geometry.

Non-rotational extension

Non-rotational extension is defined as a kinematic mechanism in which extension takes place without rotation of geological features, for example the high-angle normal fault in Fig. 11 (Table 1). For convenience we have plotted the dip of the fault versus fault displacement for various values of net extension. For non-rotational faults which have a very gentle dip, determination of displacement is difficult in most geological situations. Consider for example an undeformed sequence of strata cut by a very low-angle normal fault. The relationships displayed in Fig. 11 show that displacement and extension on the fault are specified if the dip of the fault and the stratigraphic omission across it are known. Thus, a more convenient representation of the magnitudes of extension possible on these faults is a plot of stratigrahic omission versus fault dip, shown in Fig. 12. In real geological situations, low-angle normal faults, in common with thrust faults, may show a ramp-decollement geometry if developed in sedimentary sequences of variable competence (Fig. 13) (Dahlstrom 1970). The fault dip appropriate for use in Fig. 12 in this situation would be the fault-bed angle averaged along the direction of transport. Introduction of listric faulting by ramps will cause rotation of strata, but the gross picture of one large sheet moving over another is most easily visualized as non-rotational, since there is no net rotation of the hanging wall. The distinguishing characteristic of low-angle normal faults, as opposed to thrust faults, is the juxtaposition of younger rocks on older with omission rather than repetition of strata. Thus, the rules of interpretation are the inverse of thrust faulting.

Large-displacement, very low-angle normal faults may show a number of movement planes, just as large displacement thrusts do. For example, consider the situation depicted in Fig. 14(a) where a large, low-angle normal fault is initiated in an undeformed sedimentary sequence. After an offset of one stratigraphic unit, movement is initiated on a slightly higher plane (Fig. 14(b)). The thin sheet between the two faults is accreted to the footwall of the first fault, and movement on the second of one more stratigraphic unit creates the configuration shown in Fig. 14(c), an attenuated stratigraphic section. The total displacement on the fault system is the sum of that across the two faults, and thus no matter how complex the system of faults, the total stratigraphic omission may be used in Fig. 12 to determine how much extension the stack represents, provided the average fault-bed angle is reasonably well-known.

EXAMPLES

We believe that the modes of extension discussed above can be found in the geological record, and present here some examples of each type.

Perhaps the most completely documented type of extensional fault is the simple listric normal fault, shown here in a reflection profile from a continental margin setting (Fig. 15). Although Fig. 15 is a time section only and thus cannot be regarded as a true geologic section, the differential tilt between hanging wall and footwall and growth fault character are clear. We interpret the hanging wall as a series of uniformly tilted fault blocks bound-ed by planar faults, which serve to extend it without creating differential tilt between blocks, analogous to the geometry shown in Fig. 7. It is important to note that in moving downward from the steeply dipping to the more gently dipping portions of the listric fault, differential rotation should gradually decreases until the fault has the characteristrics of a very low-angle non-rotational fault.

A series of imbricate normal- fault blocks showing successively steeper tilts was mapped by Anderson (1971) (Fig. 16). The configuration requires initially curviplanar fault blocks, although some have apparently been 'straightened out' by the extension (cf. Fig. 6).

Figure 17 shows a small-scale example of rotational planar faulting from the Rawhide Mountains of west-central Arizona, where measured and theoretical determinations of percent extension using Fig. 2 agree at about 25-30%. The first documented planar rotational normal faulting was recognized by Emmons & Garrey (1910) (Fig. 18). They mapped an impressive sequence of evenly tilted fault blocks in Tertiary volcanic rocks in the Bullfrog Hills of southern Nevada. They interpreted the structures there as having formed like a row of tilted dominos in which both faults and beds rotate simultaneously and concluded that they could be most easily explained by extension of the crust. Fifty years later the relations between fault dip, stratal dip and extension, were derived by Thompson (1960), and were again derived by Morton & Black (1975). Despite three separate conceptualizations of this mechanism, it has been almost completely ignored in the Basin and Range literature, where beginning with Longwell (1945) virtually all small-scale imbricate normal faults were described as listric (except Thompson 1971, who speculated that some of Anderson's (1971) normal faults may be planar) to the point that low-angle normal fault and listric normal fault came to be used interchangeably by many authors. We believe that true listric normal faults as envisioned by most workers form only a portion extended terranes, and that much (if not most) extensional strain in the earth's crust is accomodated by both rotational and non-rotational planar normal faults.

Non-rotational, high-angle extensional faults are described abundantly by many geologists (e.g., Stewart 1971), but much less attention has been given to their low-angle counterparts. One of us (Wernicke 1981) has emphasized the potential importance of these faults in accommodating lithospheric extension, and we suspect that structures produced by this type of fault are extremely common in the Basin and Range Province. For example, Dechert (1967) (Fig. 19) mapped a stack of fault slices in Paleozoic miogeoclinal and Tertiary volcanic strata in the Schell Creek Range of east-central Nevada across which about 5 km of strata are missing. Because the faults are nearly parallel with bedding, the structure is most logically thought of as having been produced by the mechanism shown in Fig. 14. Fig. 12 suggests that these sheets record tens of kilometres of extension. Noble (1941) coined the term Chaos for this type of structure after an example he mapped in the Death Valley region (Amargosa Chaos), and attributed its formation to a large, regional thrust sheet. Wright &

Troxel (1973) recognized the extensional nature of the Amargosa Chaos, and proposed that it developed in a zone of coalescing listric normal faults. Although their interpretation is reasonable for portions of the Amargosa Chaos, we would like to emphasize that a comparable structural assemblage may form without rotational faulting simply by peeling sheets off the base of the hanging wall, that is the allochthon, of large displacement, non-rotational normal faults.

Another example of a large-scale extensional allochthon can be seen on a seismic reflection profile from the Sevier Desert area of central Utah (McDonald 1976) (Fig. 20). The section reproduced convincingly demonstrates that parts of a given hanging wall may remain unrotated while other parts may rotate in opposite directions. We conclude from this that (1) the sense of rotation in a given hanging wall does not indicate its transport direction, and (2) the boundaries between tilted and non-tilted parts of extensional allochthons give no information as to the extent of their basal faults beyond those boundaries. In other words, large areas of seemingly intact rock at the surface may be underlain by large, low-angle non-rotational normal faults.

DISCUSSION

The extensional mechanisms described and discussed above are not mutually exclusive and can operate contemporaneously within a given extensional system. Listric normal faults, because of their geometry, can separate areas undergoing differential extension and merge at depth with low angle normal faults. Above a low-angle normal fault further extension can be accomodated by imbricate normal faulting in which both faults and beds rotate (Fig. 19). In areas where large magnitude extension has occurred, the low-angle normal faults and superjacent imbricate rotational faults may be the main contributors to the overall extension. This simple scheme can be modified and become more complex where differential extension has occured in the rotated block sequence, in which case listric faults should be present. The low-angle fault or faults at the base of the faulted sequence anastomose leading to the development of Chaos-type structure. Low-angle faults may also be present at different structural levels, thus dividing the crust into an imbricate stack of allochthonous slices each with both rotational and non-rotational fault blocks in their hanging walls.

One of the important problems is the geometry and character of the low-angle normal fault at the base of a series of faulted blocks. Wernicke (1981) has suggested such faults may involve the entire lithosphere. If this is the situation, the lower crust may be extended simply by divergence of two rigid slabs separated by a gently dipping shear zone. Brittle shear would occur at shallow levels and grade downward along the low-angle fault to a zone of ductile shear (Fig. 21). The ductile shear would be restricted to the shear zone rather than distributed uniformly throughout the footwall crustal block, as envisioned by Eaton (1979) and Le Pichon & Sibouet (1981). With such a geometry, continental crust could be attenuated to any thickness, and unmetamorphosed sedimentary rocks could be juxtaposed by large-displacement low-angle normal faulting with any part of the crust. In this type of extensional system, zones of ductiley sheared rock that formed at deep structural levels along the shear zone early in the history of deformation may be reworked in the brittle regime as shown in Fig. 19.

It is possible, for example, that the prominent reflector at the base of the imbricate normal fault blocks in the Bay of Biscay (reflector 's',

Fig. 22) is simply a crustal-scale low-angle normal fault, rather than a boundary between brittle extension and penetrative ductile stretching as suggested by de Charpal et al. (1978) and Le Pichon & Sibouet (1981). Such an interpretation is consistent with geometrically identical examples in the Basin and Range Province where the rocks below the basal detachments behaved as rigid plates during emplacement of the extensional allochthons (e.g. Misch 1960, Davis et al. 1980, Wernicke 1981). Thus, although Le Pichon & Sibouet (1981) demonstrated that the degree of extension by imbricate normal faulting was as large as a factor of two or three, their assumption that the lower crust stretched penetratively by that amount may be incorrect; it is geometrically possible that the crust beneath the basal reflector and mantle lithosphere has not extended at all! This view has rather dramatic implications for geophysical models of passive margin rifting using subsidence history because it violates their fundamental assumption that the lithosphere stretches like a large elastic band (e.g. McKenzie 1978). It implies that a certain amount of crust which originally lay above reflector 's' is now incorporated in the complex southern margin of the Bay of Biscay and in the western Pyrenees. Furthermore, it implies the lower crust can be very heterogeneous.

Large displacement low-angle normal faults which serve as boundary faults to both rotational and non-rotational extensional fault mosaics have been termed detachment faults by Davis et al. (1980), and Dokka (1981) and Davis et al. (1981) followed this usage when describing a number of terranes throughout southern California, Arizona, northern Sonora, Mexico. We believe the overall geometry and kinematics of many extensional terranes have much in common, but since their three-dimensional geometries are poorly known the use of such terminology should remain informal.

Non-rotational high-angle normal faults are present in extensional terranes, but contribute only a small amount to the overall extension. They may be superimposed on older, large-scale extensional fault systems like those described above (Eberly & Stanley, 1978, Zoback et al. 1981) and/or be the dominant fault type in areas that have undergone lesser amounts of extension in an inhomogeneously extended terrane.

CONCLUSIONS

The geometry and kinematics of faults in extensional regions can be grouped into two broad categories: (1) those which produce extension accompanied by rotation of layers and in a subgroup are accompanied by rotation of the faults as well; and (2) those which produce extension without rotation of faults or layers (Table 1). Our analysis suggests that large-scale extension is accomplished by large displacements on low-angle normal faults of the second group and rotated faults and fault blocks (both listric and planar) of the first group. Non-rotated listric normal faults are geometrically important as 'space fillers' but may not be as significant as the other fault types in producing large-scale extension. Several fault types are related and form contemporaneously: listric normal faults, rotated faults and fault blocks, and large-displacement, non-rotational low-angle normal faults may all form a single fault system with the individual fault types unequally developed from place to place.

While many of the examples presented here are from the Basin and Range Province of the United States, the ideas are probably valid for any terrane which has undergone large magnitude extension. Passive continental margins are probably regions of large magnitude extension, and similar complex fault systems may have developed during their formation.

Our ideas are ultimately testable. Greater detailed mapping coupled with geophysics and drill hole information should provide the necessary three-dimensional control. From such data we should be able to palinspastically reconstruct the extended terrane, just as we do thrust fault terranes.

Acknowledgements -This paper benefitted from the critical reviews of R.E. Anderson, A.W. Bally, P.L. Hancock and James Helwig. We thank M.S. Beaufait, R.G. Bohannon, G.A. Davis, R.K. Dokka, E.G. Frost, P.L. Guth, W.B. Hamilton, L. Royden and T.J. Shackleford for fruitful discussions. This work was supported by NSF Grant 7713637 awarded to B.C. Burchfiel and NSF Grant EAR7926346 awarded to B.C. Burchfiel and P. Molnar.

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TABLE 1

TYPES OF NORMAL FAULTS

GROUP	ROTATED	FAULT GEOMETRY
Non-Rotational	Nothing	Planar
Rotational	Beds	Listric
"	Beds and Faults	Planar or Listric

FIGURE CAPTIONS

Fig. 1. Extension and attenuation of a rock mass by rotational planar normal faulting.

Fig. 2. Geometrical derivation of the relationship between fault dip (ϕ), bed dip (θ), and percent extension of the rock mass, as done by Thompson (1960).

Fig. 3. Graphical representation of relations derived in Fig. 2.

Fig. 4. Listric normal fault with "reverse drag" (Hamblin, 1965).

Fig. 5. Imbricate listric normal faults.

Fig. 6. Thin, imbricate listric fault blocks "unflexed" by extreme extension, the displacement on each fault roughly the same as the length of the blocks.

Fig. 7. Listric normal fault bounding a family of planar fault blocks from a stable area.

Fig. 8. Hypothetical geologic map and section depicting the difference in amount of extension determined palinspastically between planar and listric rotational normal faulting.

Fig. 9. Listric fault model. See text for explanation.

Fig. 10. Comparison of extension for listric and planar rotational normal faults.

Fig. 11. Non-rotational normal fault, with equations and graphical representation of the relations between displacement (d), extension (e), stratigraphic omission (S), and fault dip (ϕ).

Fig. 12. Plot of stratigraphic omission (S) versus fault dip or average fault-bed angle (ϕ) using non-rotational fault model and equations from Fig. 9.

Fig. 13. Ramp-decollement geometry for large normal faults. Introducing a component of listric faulting forces some rotation of bedding (after Dahlstrom, 1970).

Fig. 14. Formation of chaos structure. a) Unfaulted sedimentary sequence, with reference points marked x and o. b) Fault displaced one stratigraphic unit. c) New fault with an additional offset of one stratigraphic unit. See text for discussion.

Fig. 15. Listric normal fault revealed by seismic reflection profiling. Reproduced with permission of Wintershall, A.G.

Fig. 16. Cross-section of imbricate listric faulting in the Eldorado Mountains of southern Nevada (after Anderson, 1971).

Fig. 17. Outcrop-scale example of rotational planar normal faulting from the Rawhide Mountains of west-central Arizona, studied by Shackelford (1980). The area shown is about 2 meters high.

Fig. 18. Cross-section of probable imbricate planar (to the left) and listric (to the right) normal faults in the Bullfrog Hills of southern Nevada (after Emmons and Garrey, 1910).

Fig. 19. Cross-section from the Schell Creek Range of east-central Nevada, showing the development of chaos structure (after Dechert, 1967).

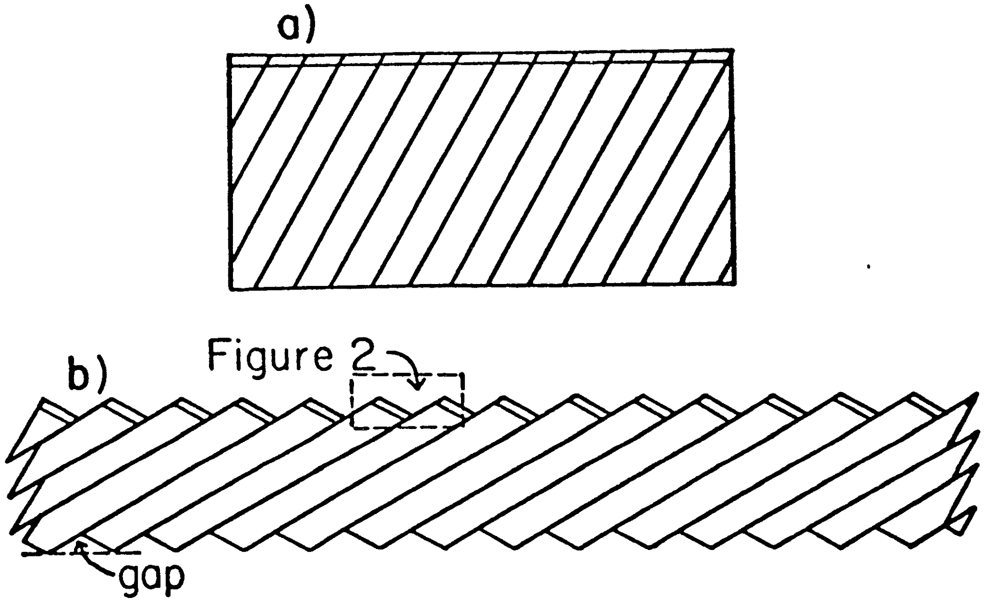
Fig. 20. Two-way time section of the Sevier Desert area, central Utah (after McDonald, 1976). Reproduced with permission of the Rocky Mountain Association of Geologists.

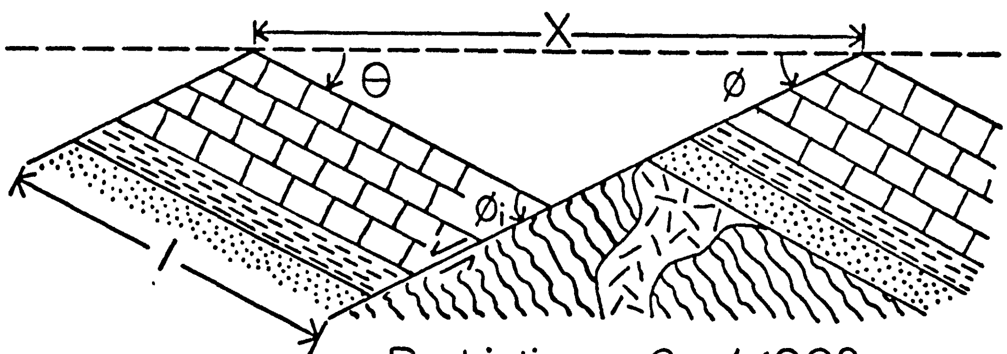
Fig. 21. Large, low-angle normal fault bounding an extensional fault mosaic comprised of listric and planar rotational faults and an example of how chaos might form (two imbricate nappes) beneath an imbricate pile of rotational normal faults. Also shown is a means by which a brittle fault mosaic may be juxtaposed upon a

slightly earlier formed penetrative ductile fabric (e.g., Snoke, 1980). From Wernicke (1981).

Fig. 22. Seismic reflection profile from the Bay of Biscay (from de Charpal *et al.*, 1978).

FIG. 1





Restriction: $\Theta + \phi < 90°$

$$\%\text{ext.} = (X-l)100 = \left[\frac{\sin(\phi+\Theta)}{\sin\phi} - 1\right]100$$

alternatively, $$\%\text{ext.} = \left[\frac{\sin\phi_i}{\sin(\phi_i-\Theta)} - 1\right]100$$

FIG. 2

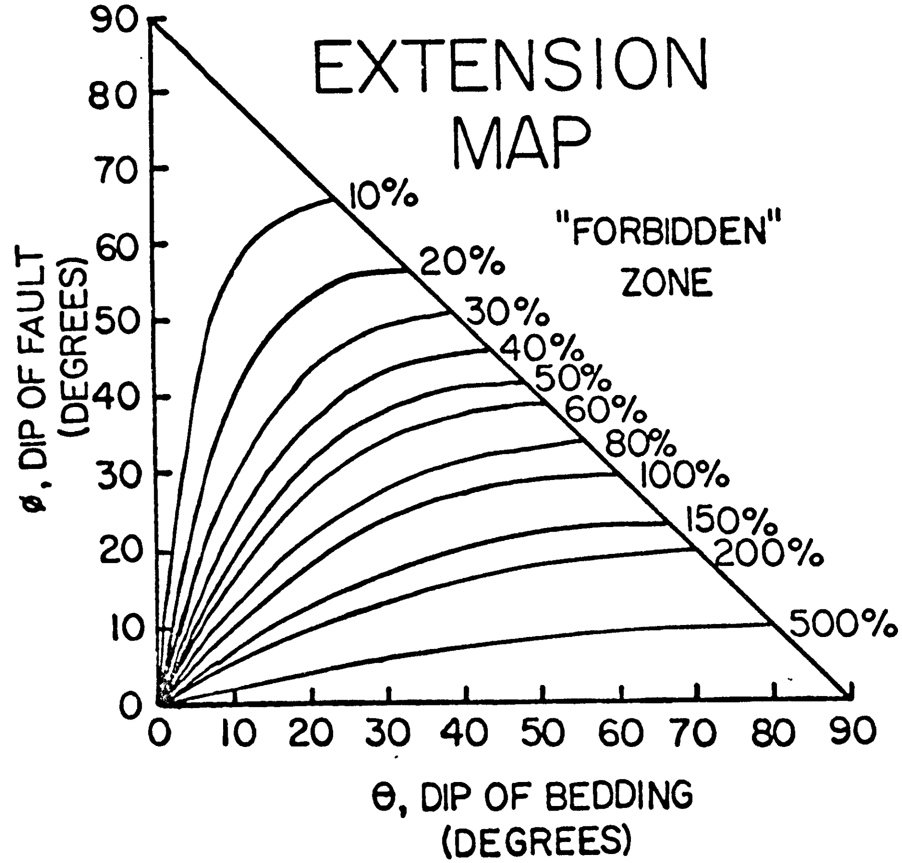

EXTENSION
MAP
"FORBIDDEN"
ZONE
10%
20%
30%
40%
50%
60%
80%
100%
150%
200%
500%
φ, DIP OF FAULT
(DEGREES)
θ, DIP OF BEDDING
(DEGREES)
0
10
20
30
40
50
60
70
80
90

FIG. 3

FIG. 4

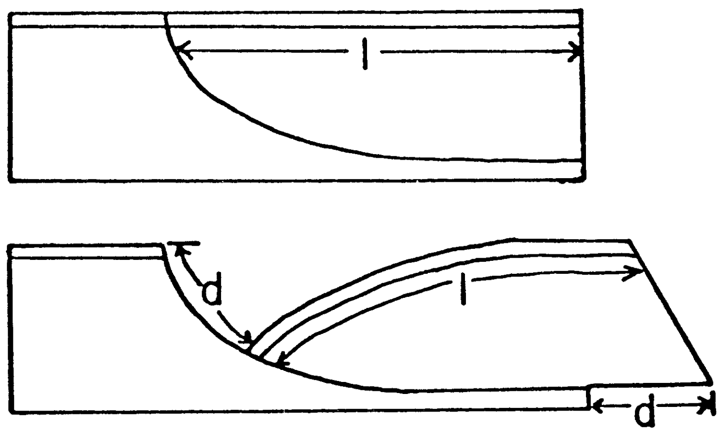


FIG. 5

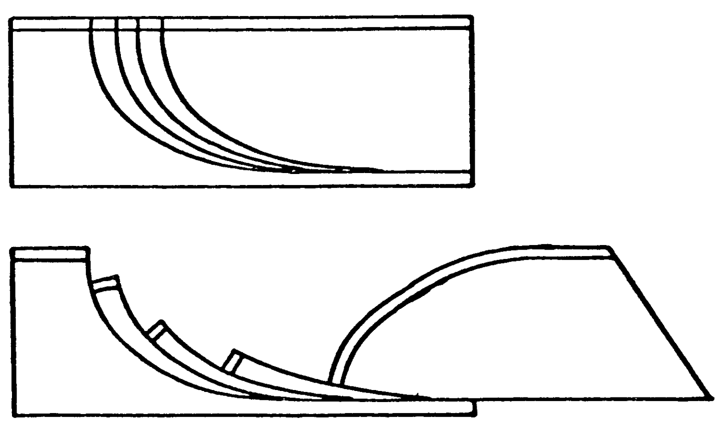


FIG. 6

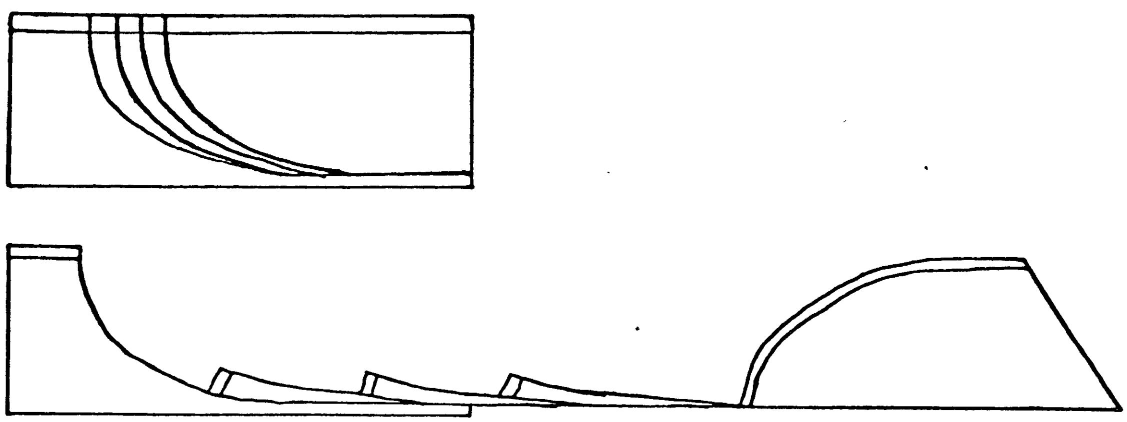


FIG. 7

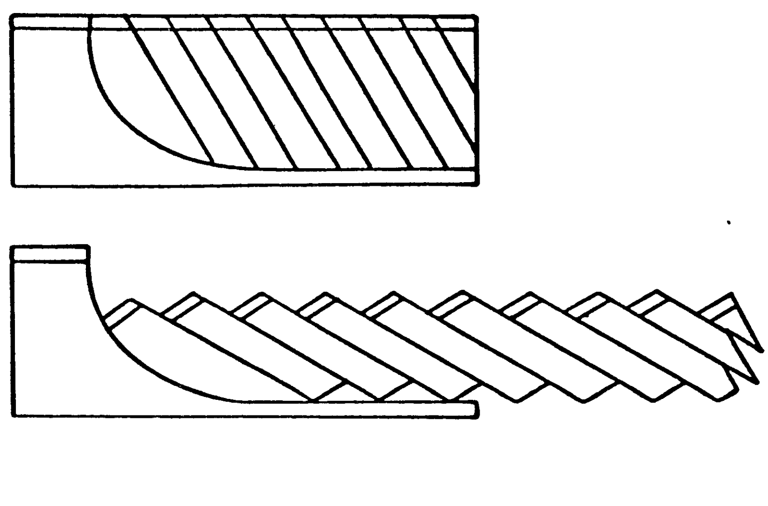


FIG. 8

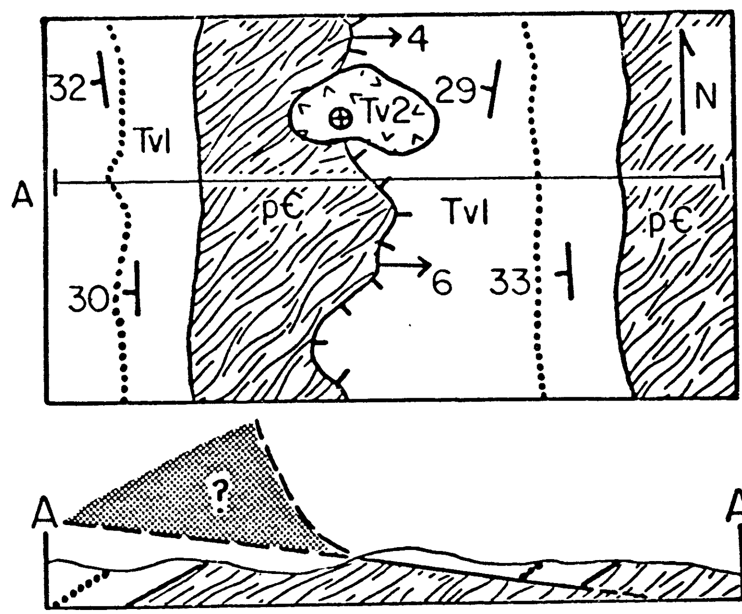


FIG. 9

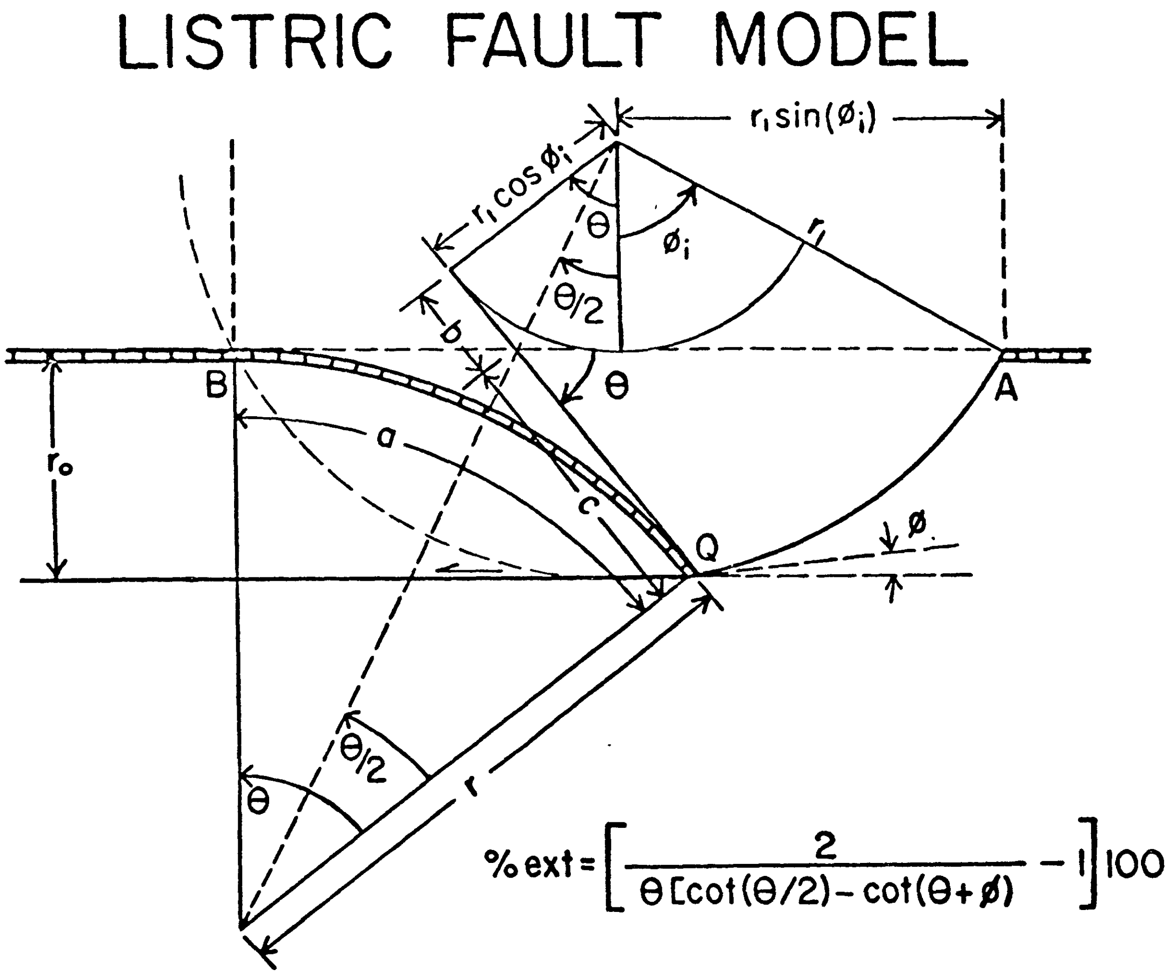

LISTRIC FAULT MODEL
r₁sin(Ø_i)
r₁cosØ_i
Θ
Ø_i
r₁
Θ/2
b
Θ
B
A
a
r₀
c
Q
Ø
Θ/2
Θ
r
%ext = [2 / Θ[cot(Θ/2) − cot(Θ+Ø)] − 1]100

FIG. 10

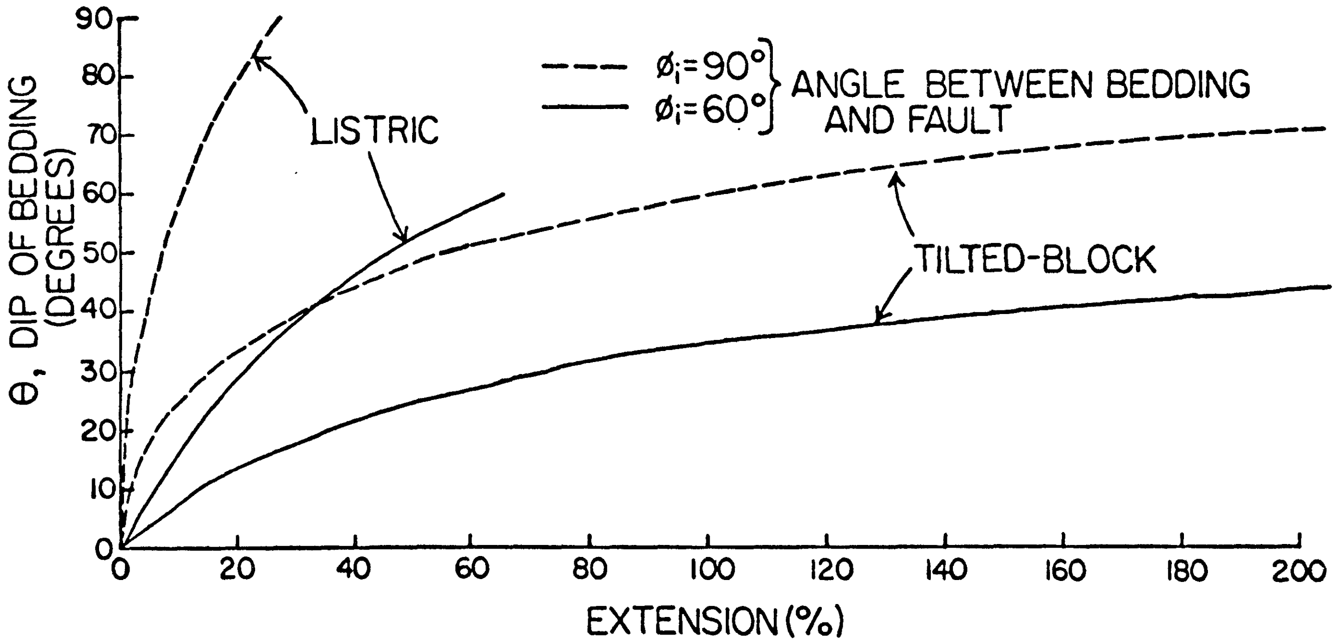

θ, DIP OF BEDDING (DEGREES)
EXTENSION (%)
LISTRIC
TILTED-BLOCK
ø_i=90°
ø_i=60°
ANGLE BETWEEN BEDDING AND FAULT
0
10
20
30
40
50
60
70
80
90
100
120
140
160
180
200

FIG. 11

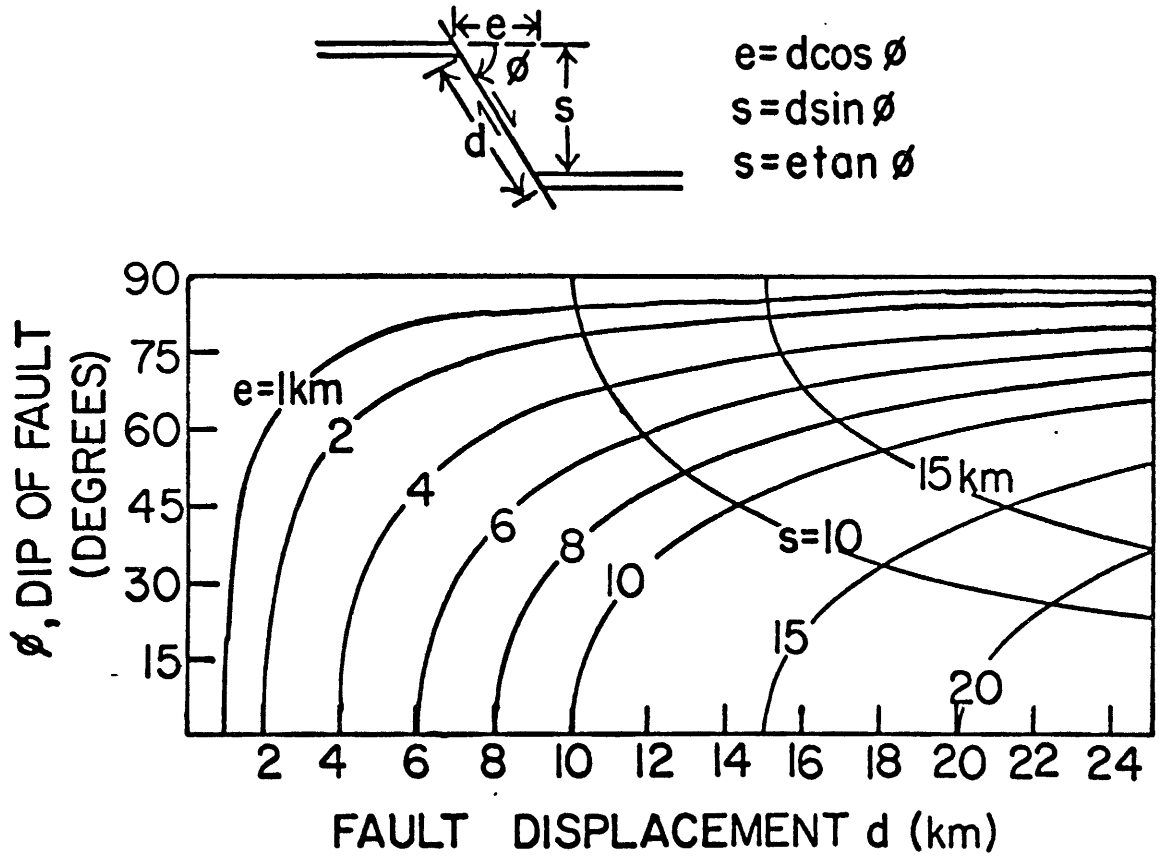
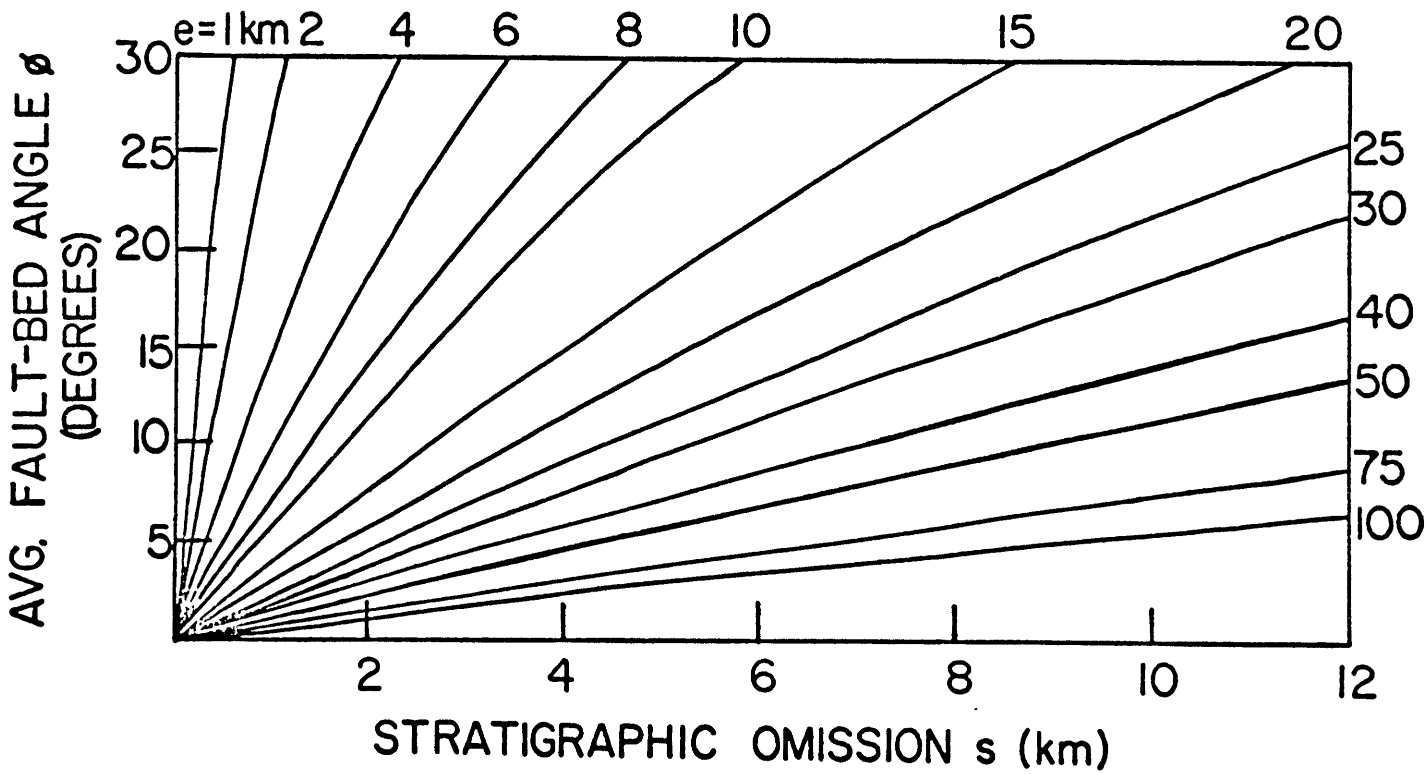

e
ø
s
d
e=dcos ø
s=dsin ø
s=etan ø
90
75
60
45
30
15
ø, DIP OF FAULT (DEGREES)
e=1km
2
4
6
8
10
15
20
15km
s=10
2 4 6 8 10 12 14 16 18 20 22 24
FAULT DISPLACEMENT d (km)

FIG. 12



e=1km 2
4
6
8
10
15
20
25
30
40
50
75
100
30
25
20
15
10
5
AVG. FAULT-BED ANGLE ø
(DEGREES)
2
4
6
8
10
12
STRATIGRAPHIC OMISSION s (km)

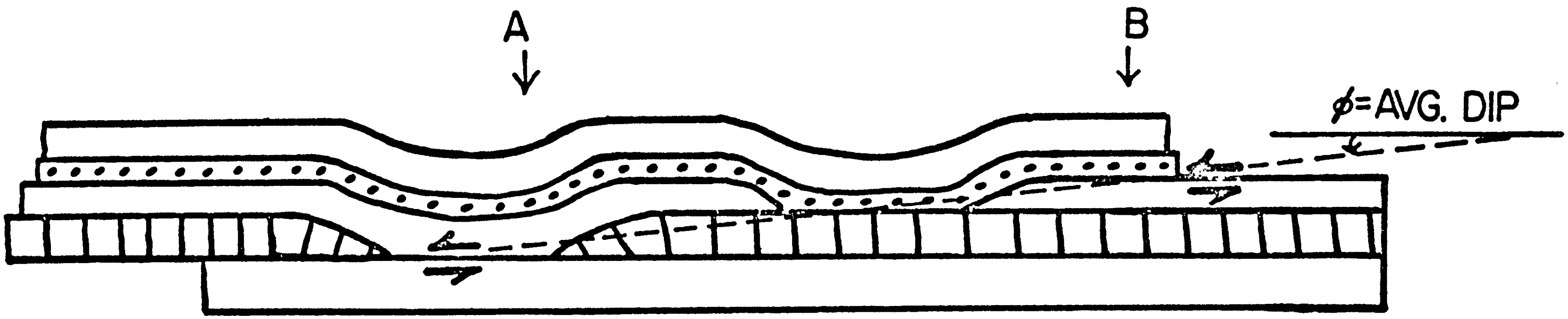


FIG. 13

FIG. 14

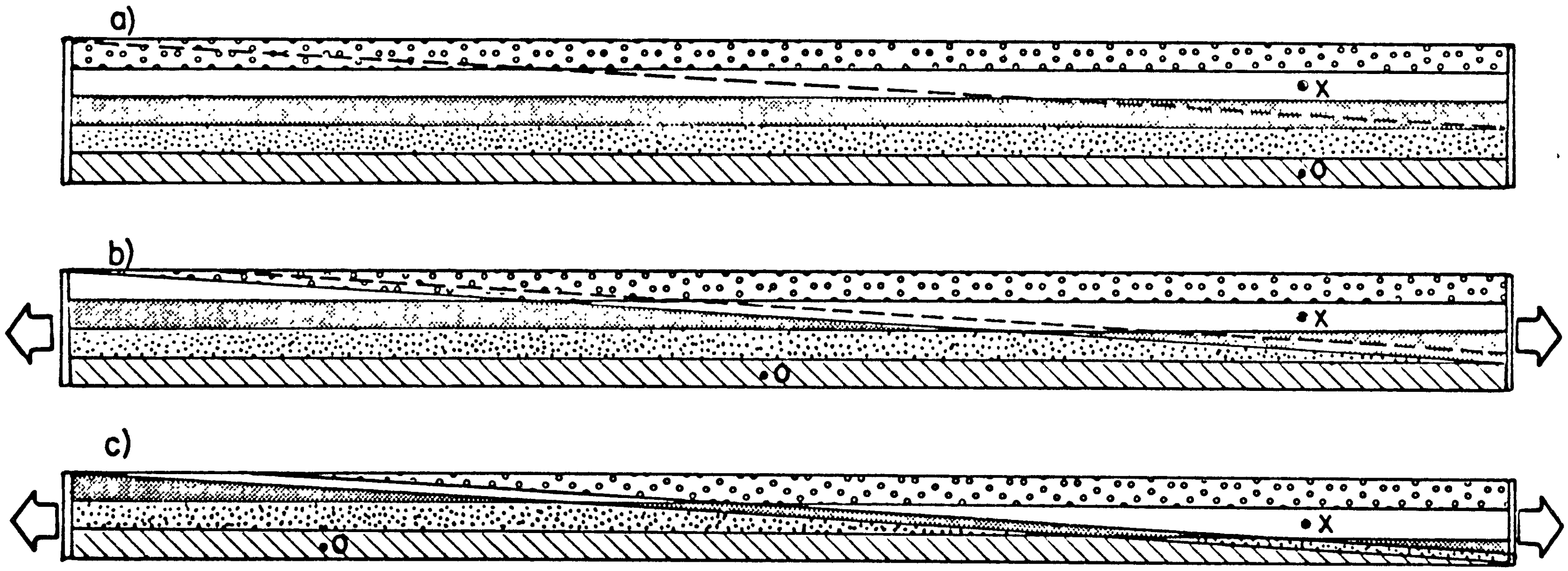
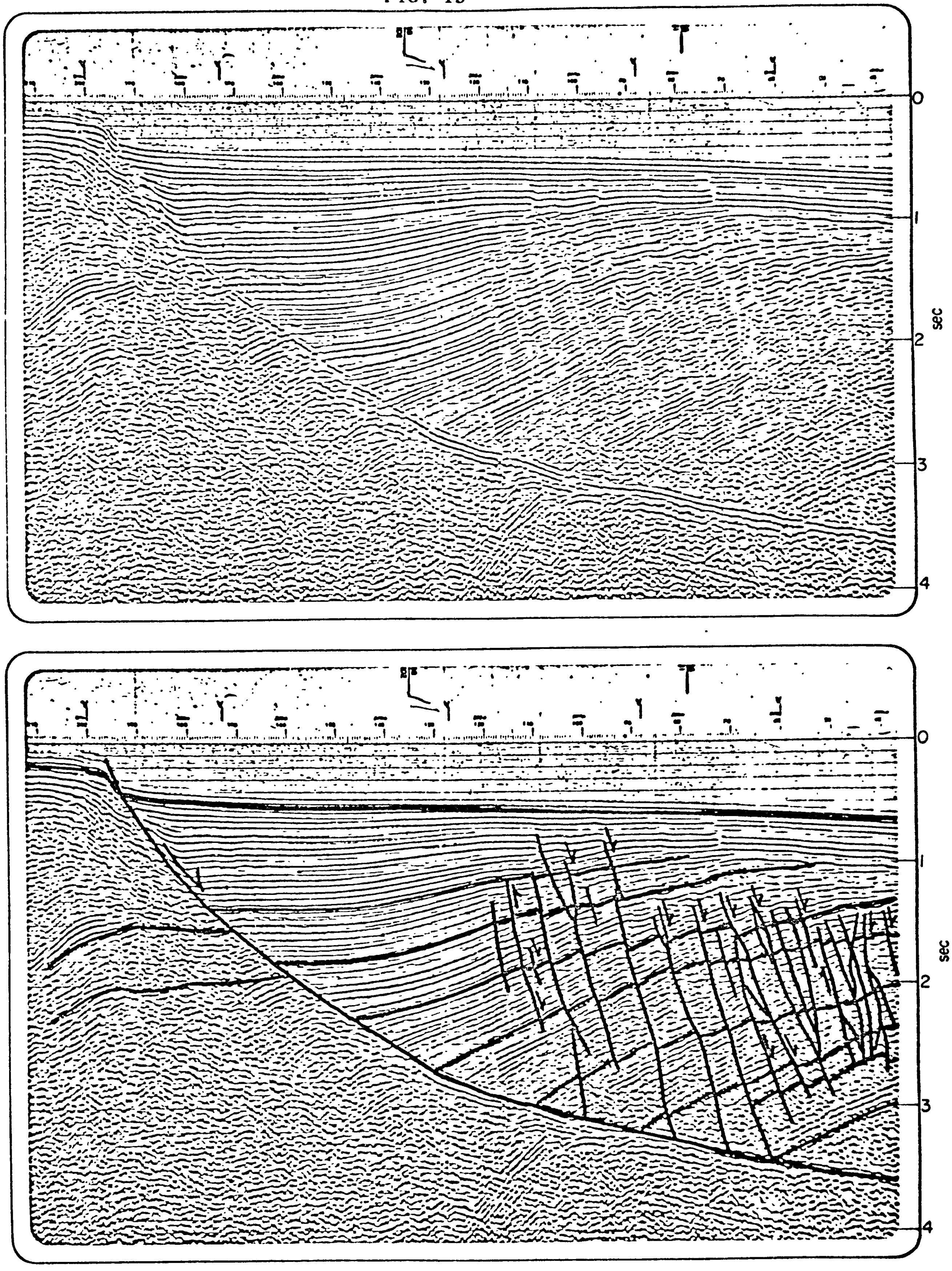


FIG. 15



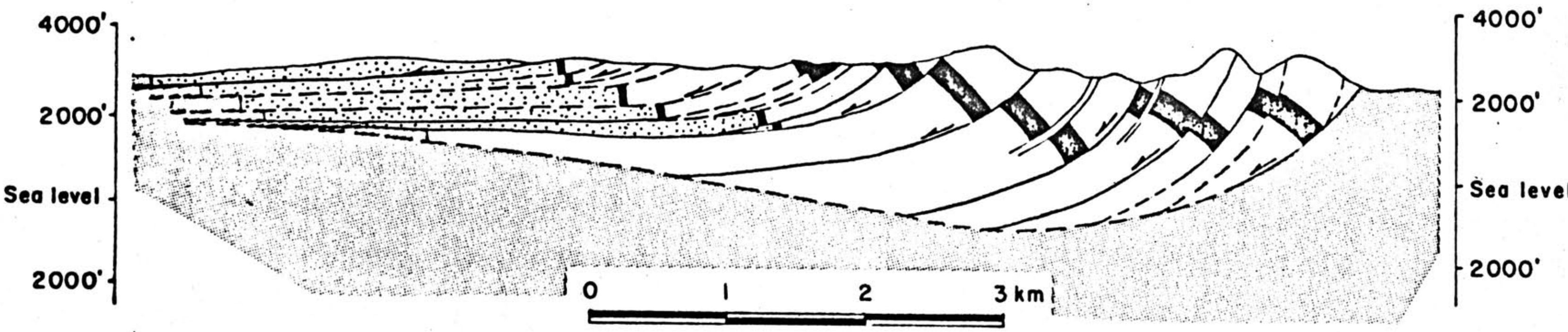


FIG. 16

FIG. 17

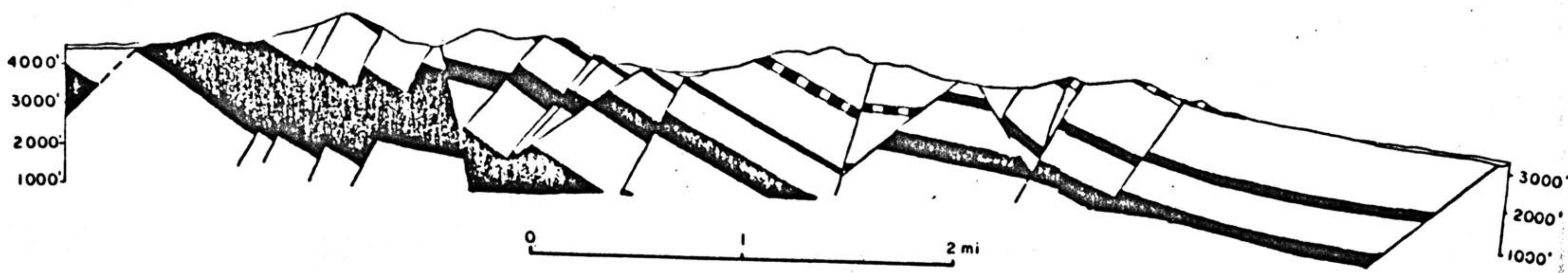


FIG. 18

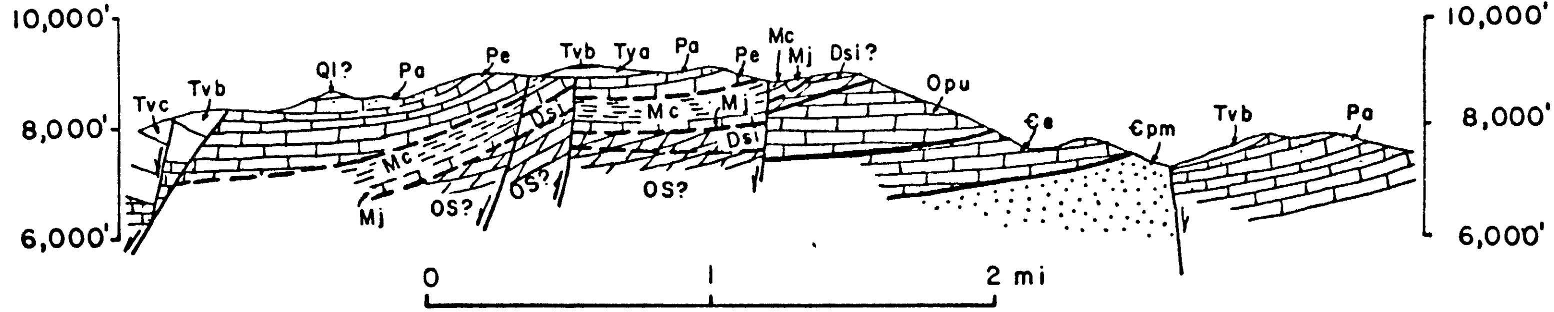

10,000'
8,000'
6,000'
Tvc
Tvb
Ql?
Pa
Pe
Tvb
Tya
Pa
Pe
Mc
Mj
Dsi?
Opu
€e
€pm
Tvb
Pa
Dsi
Mc
Mj
Dsi
Mc
Mj
OS?
OS?
OS?
10,000'
8,000'
6,000'
0
1
2 mi

FIG. 19

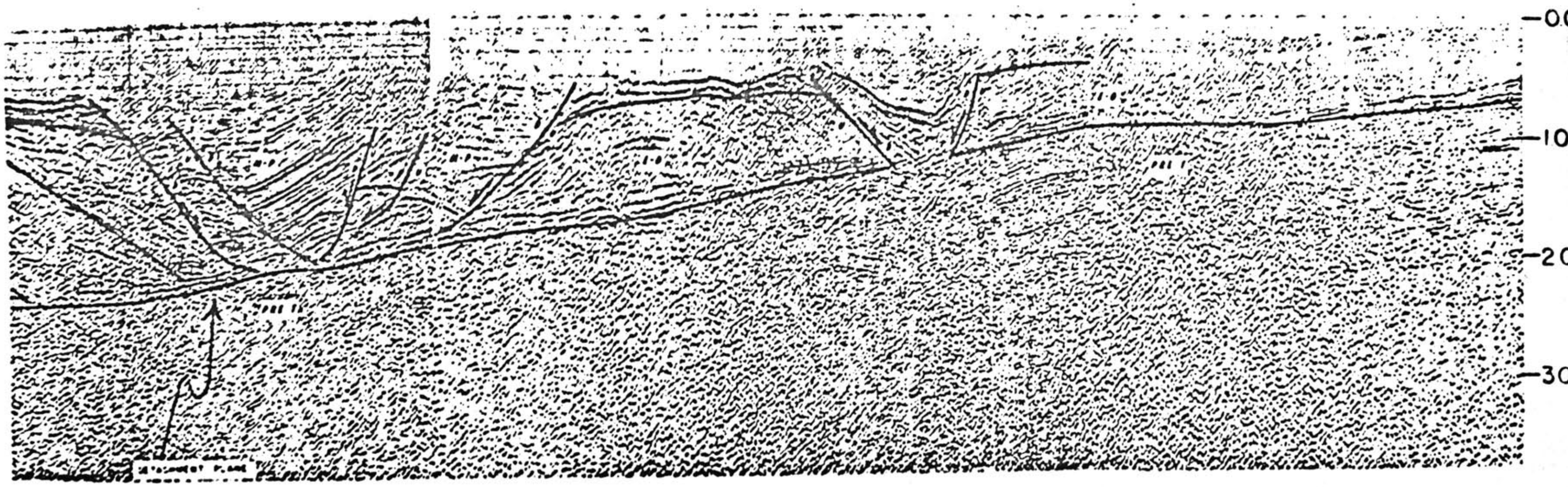


FIG. 20

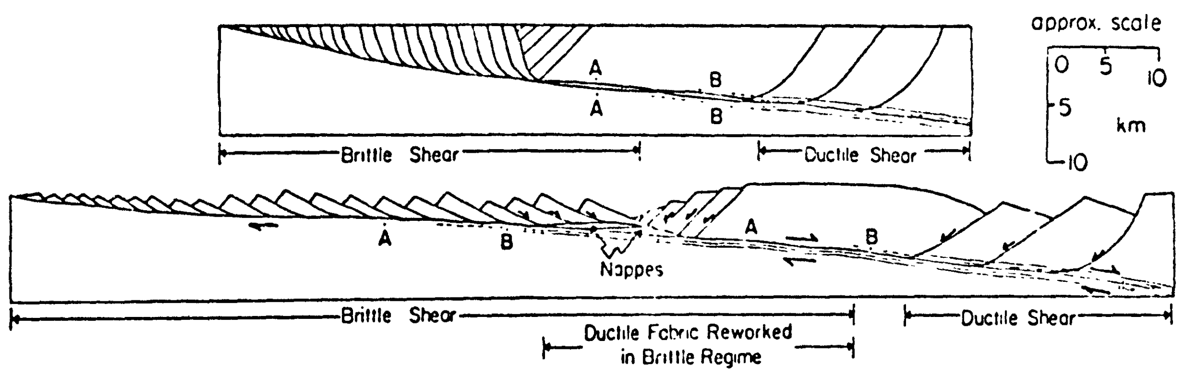

approx. scale
0 5 10
5
km
10
A
B
A
B
Brittle Shear
Ductile Shear
A
B
A
B
Nappes
Brittle Shear
Ductile Fabric Reworked in Brittle Regime
Ductile Shear

FIG. 21

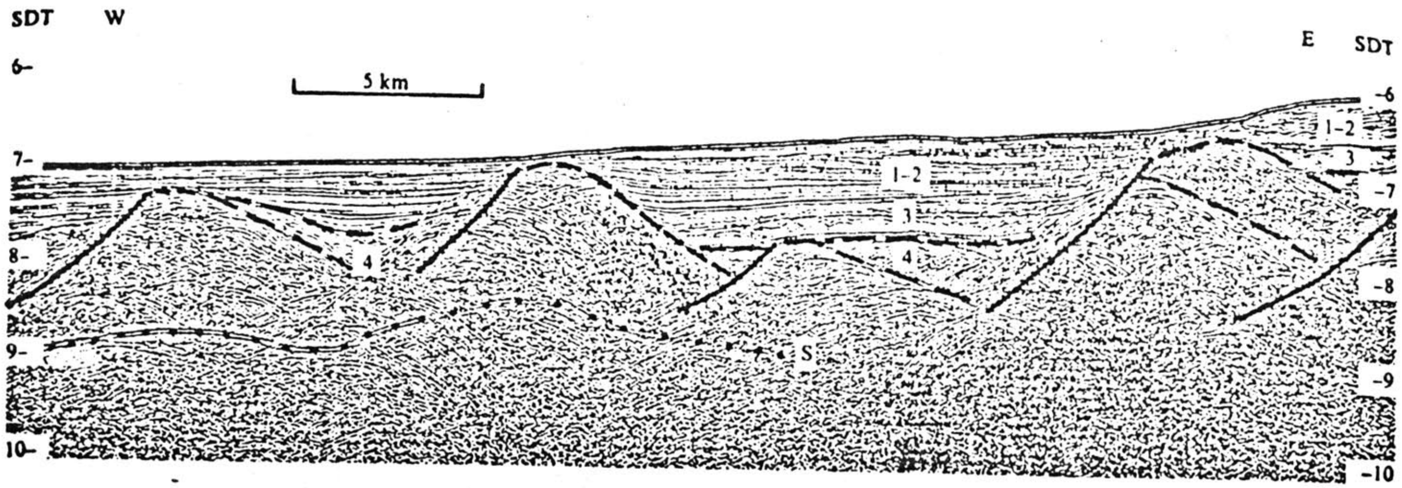


FIG. 22

CHAPTER III. MAGNITUDE OF CRUSTAL EXTENSION IN THE SOUTHERN GREAT BASIN

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Abstract

Strike-slip faults in the southern Great Basin separate areas of Cenozoic upper crustal extension from relatively stable tectonic blocks. Linear geologic features, offset along the Garlock fault, Las Vegas Valley shear zone, and Lake Mead fault system, allow reconstruction of the southern Great Basin to a pre-extension configuration. The Sierra Nevada, Mojave Desert, Spring Mountains, and Colorado Plateau are treated as stable, unextended blocks which have moved relative to each other in response to crustal extension, with the Spring Mountains held fixed to the Mojave block. Our reconstruction indicates a minimum of 65% extension (140 km) between the southern Sierra Nevada and Colorado Plateau.

INTRODUCTION

The amount that continental lithosphere may extend without (or prior to) the formation of oceanic lithosphere is of central importance to geodynamics, yet accurate determinations of large-scale intracontinental extension, constrained by several independent lines of evidence, are sparse. For example, the amount of Cenozoic extension in the Basin

Range Province (BRP) has traditionally been an extremely difficult quantity to constrain. Estimates of province-wide extension presented thus far in the literature range from 10-100% increase over original width, which corresponds to about 70-400 km of pull-apart in the northern BRP. Conservative estimates (10-30%) are based on assumptions of normal fault geometry in which stratal rotation is proportional to the amount of extension (Thompson, 1960). Assuming an average angle of 60° between faults and beds, the observed average tilt of 15-20° for late Cenozoic Basin and Range fault blocks led Stewart (1980) to deduce 20-30% extension for the entire province. It should be noted, however, that Stewart intended this estimate to apply only to extension related to the modern basins and ranges, and not to events which predate their formation. Liberal estimates are based on three arguments. One is the relative crustal thicknesses between the BRP and the Sierran and Colorado Plateau provinces (Anderson, 1971b; Hamilton, 1978). Estimates ranging from 30-100% can be made if one assumes the current 20-35 km BRP crust was thinned from a crust as thick as the 40-50 km Colorado Plateau and Sierran crust. Unfortunately, actual crustal thickness in many areas of the BRP is highly uncertain, and the pre-extension configuration of the BRP Moho can only be assumed. Another estimate is based on palinspastic restoration of Mesozoic paleotectonic elements of the Cordillera (Hamilton and Myers, 1966; Hamilton, 1969) which realigns the Sierran batholith and associated oceanic terranes with the Idaho batholith and oceanic terranes in western Idaho and eastern Oregon. This method yields roughly 50-100% extension for the northern BRP. A third estimate is based on the Cenozoic clockwise rotation of the western Cascades bracketed at 27±7° by Magill and others (1981). They used a combination of paleomagnetic and geologic constraints to estimate 340 km (74%) extension in the nothern BRP and 210 km (140%) at the latitude of Las Vegas, Nevada. We feel that their analysis provides support in favor of a large amount of extension, but within the uncertainties of the constraints from which these figures were deduced, extension could have been as little as 210 km (36%) in the northern BRP and 80 km (33%) at the latitude of Las Vegas. Problematically, Mankinen and Irwin (1982) and Craig (1981) have presented data which suggest smaller clockwise rotations of the Cascades (12° ± 11°) and Klamaths (12° ± 16°).

Perhaps the most accurate way of estimating extension is by restoring offset of linear geologic features across strike-slip faults which represent transform-like faults between areas of differential extension (Hamilton and Myers, 1966; Wright and Troxel, 1968) (Fig. 1). This method has been effectively used to quantitatively constrain minimum amounts of extension on a subregional scale (e.g., Davis and Burchfiel, 1973; Guth, 1981). The favorable distribution of these structures across the southern Great Basin enables us to make an accurate minimum determination of extension across the entire province at the latitude of Las Vegas (Fig. 1a). Since this method is independent of assumptions implicit in the estimates based on normal fault geometry, large-scale paleogeography, crustal thickness, and paleomagnetic data, we feel that it provides an independent test of those estimates.

GARLOCK FAULT

Reconstruction of the western part of the transect is possible by matching correlative features across the Garlock fault. Left-lateral displacement on the Garlock fault, due to east-west extension in the BRP, appears to increase westward to a zone of maximum offset between the southern Sierra Nevada and the Mojave block (Hamilton and Myers, 1966; Troxel and others, 1972; Davis and Burchfiel, 1973). Correlation of the southern part of the Independence dike swarm north of the Garlock fault with dike swarms on the south side of the fault was first proposed by Smith (1962), who concluded that approximately 64 km of left-lateral offset had occurred since emplacement of the dikes in Mesozoic time (Fig. 1a). Smith and Ketner (1970) later proposed 48 to 64 km of left-lateral offset of the Paleozoic Garlock Formation. Finally, Davis and Burchfiel (1973) proposed that the Layton Well thrust, mapped by Smith and others (1968), is the offset equivalent of a thrust fault in the Granite Mountains 56 to 64 km to the east. These and other possible correlative features also discussed by Davis and Burchfiel (1973) constitute strong evidence for major left-lateral displacement.

Davis and Burchfiel (1973) proposed that the Garlock fault once continued at least 30 km to the east of its current surface termination at the southern Death Valley fault zone to a point of zero offset between the Kingston Range and Silurian Hills (Fig. 1a). According to this hypothesis, approximately 60 km of offset on the eastern Garlock fault was accommodated by extensional faulting across a terrane now 115 km wide. This terrane, encompassing the Owlshead Mountains, Kingston

Range, and a broad, structurally complex intermediate area must have undergone at least 100% extension. Recent unpublished mapping by Burchfiel and others (1981 M.I.T. field camp) has shown that the eastern part of the Kingston Range contains numerous low-angle normal faults and steeply tilted fault blocks, similar to the intermediate area (Wright and Troxel, 1973). Interestingly, the extended terrane ends to the south roughly along the eastward projection of the Garlock fault.

In Fig. 1b, we have essentially followed the reconstruction of Davis and Burchfiel (1973), and note that as long as crustal shortening south of the Garlock fault can be ruled out, our estimate of 60 km net crustal extension north of the Garlock fault is minimum.

LAS VEGAS VALLEY SHEAR ZONE AND LAKE MEAD FAULT SYSTEM

Reconstruction of the eastern part of the transect is made possible by matching features across the Las Vegas Valley shear zone and Lake Mead fault system (Fig. 1a). Right-lateral offset on the Las Vegas Valley shear zone was first postulated by Longwell (1960) on the basis of apparent eastward displacement of thrust faults which place Cambrian carbonates over Jurassic sandstones (Keystone-Muddy Mountain system). Burchfiel (1965) showed that the shear zone did not exist as a surface rupture west of the northernmost part of the Spring Mountains block, but was instead expressed by large-scale bending (termed an "oroflex" by Albers, 1967), which gradually gave way eastward to a discrete fault zone. He strengthened Longwell's estimate of offset by correlating the Wheeler Pass thrust in the Spring Mountains with the Gass Peak thrust north of the shear zone, suggesting an offset of about

23 km by surface breaking beneath Las Vegas Valley and an additional 21 km by bending. To this figure, Longwell (1974) postulated an additional 25 km of right-slip due to bending of the Wheeler Pass thrust from a north-south to northeast strike, as observed in the structural grain of two lower thrusts (Lee Canyon and Keystone) in the Spring Mountains. These data, in addition to stratigraphic data summarized in Stewart and others (1968), indicate total displacement in the range 44-69 km.

The transform character of the Las Vegas Valley shear zone envisioned by Davis and Burchfiel (1973) and Liggett and Childs (1974) was examined in detail by Guth (1981). He observed that oroflexural bending in the Specter Range area was not as great as the offset of the Gass Peak-Wheeler Pass thrust, and suggested this difference in offset could be accommodated by extension north of the shear zone. His hypothesis is strongly supported by the surface geology, in which few extensional structures can be found south of the shear zone in the Spring Mountains block, but a broad terrane of steeply tilted Paleozoic and Tertiary strata are offset along low-angle normal faults north of it (Longwell, 1945; Guth, 1981). Guth further noted that the spacing between the Gass Peak and Glendale thrusts north of the shear zone is nearly identical to that between the Wheeler Pass and Keystone thrust to the south. The areas between the faults both north and south of the shear zone are devoid of structures suggestive of significant Tertiary extension, thus supporting the transform concept and indicating that these areas behaved as stable blocks during extension.

Offset on the complex Lake Mead fault system was first noticed by Anderson (1973), who documented a 20 km left-lateral offset of the

12.7 m.y. Hamblin-Cleopatra stratvolcano by one fault of the system, and speculated on roughly 65 km of total displacement. In a study of Tertiary sediments in the Lake Mead region, Bohannon (1979) provided firm support of about 65 km of left-slip on the system as a whole based on the remarkable similarity between Frenchman Mountain (location F, Fig. 1a) and the South Virgin Mountains, including (1) a distinctive stratigraphic sequence at the base of the Tertiary section (locations a and a' on Fig. 1a), (2) gradual southward pinchout of identical Mesozoic formations beneath the basal Tertiary unconformity, and (3) the presence of distinctive monolithologic breccia sheets of Gold Butte rapakivi granite (Anderson, 1973; Longwell, 1974; now exposed in bedrock only in the South Virgin Mountains, location GB, Fig. 1a) in the Tertiary section at Frenchman Mountain. Consistent with Bohannon's figure, Smith (1981) provided data in support of a 40 km offset for one strand of the Lake Mead fault system by correlating the River Mountains stock just southeast of Frenchman Mountain with a compositonally similar volcano-plutonic assemblage in the northern Black Mountains (b' and b, respectively, Fig. 1a).

DISCUSSION

Conservative reconstruction of offset features along the strike-slip faults yields a total pull-apart of 140 km for the transect (Fig. 1b), which corresponds to a 65% increase in original width. This figure is based on the assumption that areas on the "stable" side of the strike-slip faults have not also extended. However, Dokka (1981) has mapped an area in the Mojave block which shows extreme northeast-southwest upper crustal extension between 21 and 16 m.y., although the

Mojave block immediately south of the Garlock has been tectonically stable relative to Pliocene and Quaternary extension in the Death Valley area. The Silurian Hills (Kupfer, 1960; Fig. 1a), just south of the projected trace of the Garlock east of the Death Valley fault zone inferred from geophysical data (Plescia and Henyey, 1982), contains extensional structures similar to those in the highly extended Death Valley region, and may possibly represent large magnitude extension, although the terrane south of the Silurian Hills between the southern Avawatz Mountains and the Spring Mountains block does not contain any significant extensional structures (Burchfiel and Davis, 1971; DeWitt, 1980). In any event, there is no evidence of significant crustal shortening anywhere south of the Garlock during extension.

Other areas of possible extension not accounted for in the reconstruction include the region between the Independence dike swarm and the Sierra Nevada, possible minor extension in the northern Spring Mountains block, almost certain extension between the South Virgin Mountains and the Colorado Plateau (see cross-section on p. 115 in Longwell, 1945).

Since the sources of uncertainty seem to have the effect of increasing the estimate, we believe that 140 km is a reasonable minimum figure, corresponding to an increase in original width of about 65%. In light of the uncertainties, it is possible that the total extension is in the neighborhood of 80-100%. We believe this result lends considerable credibility to the other lines of evidence disussed above that suggest Cenozoic extension of the BRP on the order of a factor of two as first deduced by Hamilton and Myers (1966). According to our analysis, the roughly 30(+) km crust characteristic of our transect

(Smith, 1979) was at least 45-50 km thick following the Mesozoic Sevier orogeny. The configuration of the Moho at that time may have been similar to the modern-day Andes, in which the crust thickens as one moves from the Brazilian Shield into the back-arc thrust belt (James, 1971).

Since the crust in the northern BRP is currently as thin or thinner than the crust in our transect, and since large areas of the northen BRP have been shown to exhibit the same structural style as highly extended areas in our transect (e.g., Armstrong, 1972) we believe that it has experienced a similar percent extension and regard 60-80% (300-500) km) increase in width as likely.

The data discussed here also support geophysical studies of continental rifting which have recently found that extension of the continental lithosphere by a factor of two without forming oceanic crust may be quite common (McKenzie, 1978). It is now the task of geologists studying well exposed exampies of severe continental pull-apart to develop models of the cross-section geometry of upper crustal extension which are consistent with that deduced by other means. The widespread occurence of Tertiary younger-over-older low-angle fault terranes throughout the BRP (Young, 1960; Wright and Troxel, 1973; Moores and others, 1968; Anderson, 1971a; Armstrong, 1972, among the early workers) is highly significant toward this development. The interpretation of these terranes as large, rooted low-angle normal fault complexes (Wernicke, 1981; Wernicke, 1982; Wernicke and Burchfiel, in press) geometrically analogous to (but not necessarily a reactivation of) thin-skinned compressional belts is one means by which large-scale extension may be expressed in the surface geology. The existence of such complexes suggests that large horizontal translations of rock masses may occur without significant stratal rotation, i.e., terranes regarded as "stable" in our reconstruction may be underlain by large low-angle normal faults.

Although our purpose here has been to place a lower limit on total extension in a straightforward manner using reliable data on the amount and direction of strike-slip offsets, we remind the reader that the less straightforward task of relating the timing and kinematics of individual extended terranes to the strike-slip faults still lies ahead, and will provide a further test of the arguments presented here.

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Acknowledgements

We thank R. E. Anderson, R. G. Bohannon, G. A. Davis, Dietrich Roeder, and L. A. Wright for critical reviews which greatly improved the manuscript. This work was supported by NSF grants EAR 7913637 awarded to B. C. Burchfiel and EAR 7926346 awarded to B. C. Burchfiel and Peter Molnar.

FIGURE CAPTION

Fig. 1(a) Map of strike-slip faults and offset geologic features in a transect extending from the Colorado Plateau to the Sierra Nevada. Shaded areas are assumed not to have undergone major extensional faulting while non-shaded areas include both extended and stable terranes. Fault movements can generally be modeled as the result of movement between the Colorado Plateau, Sierra Nevada, and the Mojave-Spring Mountains block. GB = Gold Butte, F = Frenchman Mountain. See text for discussion. (b) Fault reconstruction based on offset features shown in (1), indicating about 140 km of net pull-apart with a component of southward movement of the Sierras with respect to the Colorado Plateau. See text for discussion.

Fig 1a

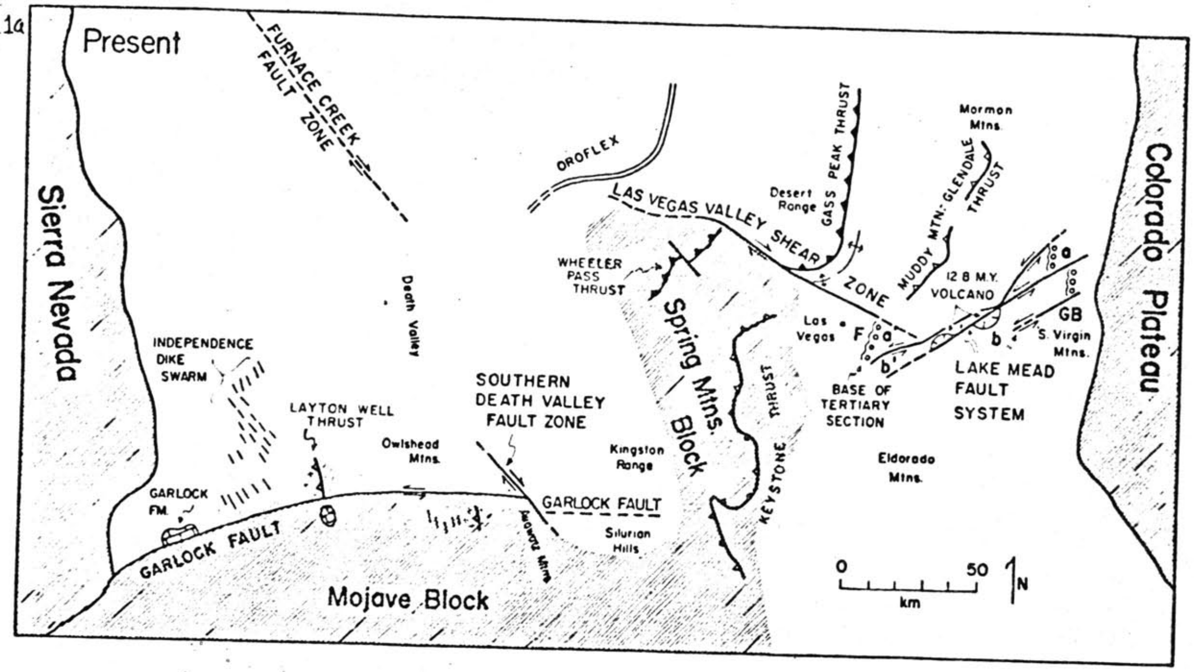
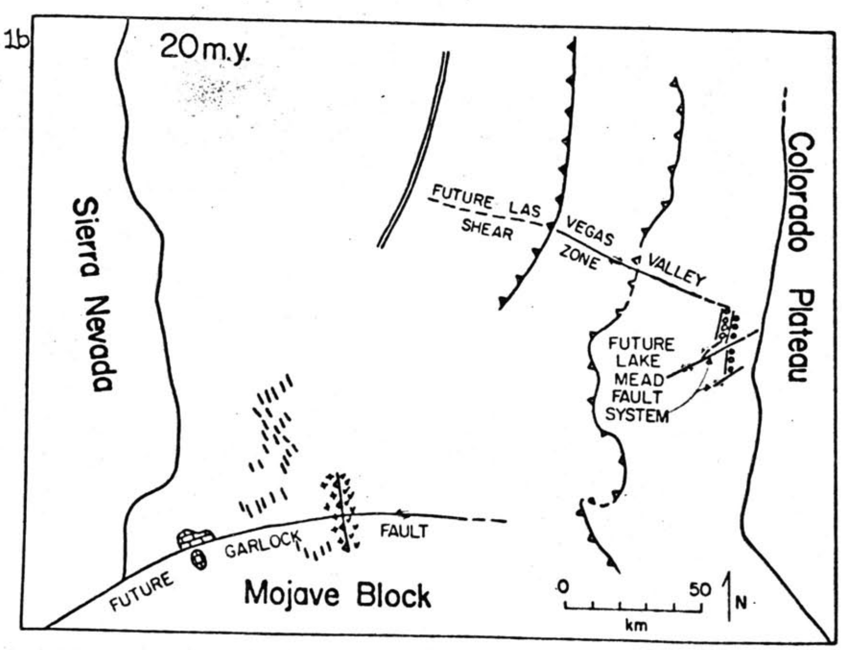

Present
FURNACE CREEK FAULT ZONE
OROFLEX
LAS VEGAS VALLEY SHEAR ZONE
Desert Range
GASS PEAK THRUST
Mormon Mtns.
MUDDY MTN.-GLENDALE THRUST
Sierra Nevada
Colorado Plateau
WHEELER PASS THRUST
Spring Mtns. Block
Death Valley
12.8 M.Y. VOLCANO
Las Vegas
GB
S. Virgin Mtns.
INDEPENDENCE DIKE SWARM
LAKE MEAD FAULT SYSTEM
BASE OF TERTIARY SECTION
SOUTHERN DEATH VALLEY FAULT ZONE
LAYTON WELL THRUST
Owlshead Mtns.
KEYSTONE THRUST
Kingston Range
Eldorado Mtns.
GARLOCK FM.
GARLOCK FAULT
Silurian Hills
Mojave Block
0 50 km
N

Fig 1b


20 m.y.
Sierra Nevada
Colorado Plateau
FUTURE LAS VEGAS VALLEY SHEAR ZONE
FUTURE LAKE MEAD FAULT SYSTEM
FUTURE GARLOCK FAULT
Mojave Block
0 50 km
N

CHAPTER IV

GEOLOGY OF THE CENTRAL MORMON MOUNTAINS, LINCOLN COUNTY, NEVADA

INTRODUCTION

In Bally and others' (1966) classic account of the Cordilleran foreland fold and thrust belt in Canada, convincing evidence was presented suggesting that a period of extension followed eastward thrusting, and was characterized by west-dipping, downward-flattening low-angle (listric) normal faults, which apparently had re-activated the older thrust faults. "Back-slippage" on thrust faults has subsequently been well documented by seismic reflection profiling throughout parts of the thrust belt not severely overprinted by extensional tectonics (e.g., Royse and others, 1975). The normal faults commonly have up to several kilometers of displacement, and form basins in which older strata are antithetically rotated into the fault surface, with older strata dipping more steeply than younger. In some parts of the Cordilleran thrust belt (particularly the Nevada-Utah sector or Sevier orogenic belt of Armstrong, 1968) and its "hinterland" to the west, a group of complex, enigmatic terranes are characterized by large-scale younger-over-older faulting much more complex than the indisputably "back-slipped" portions of the thrust belt in Canada and Wyoming. In these terranes the concept of thrust fault reactivation has been applied to faults which lack prima facie evidence of an earlier compressional phase of movement (e.g. McDonald, 1976; Sprinkle, 1979; Keith, 1982; Drewes, 1967, 1981).

The Mormon Mountains, located along the eastern margin of the Basin and Range physiographic province, are a well-exposed example of superposed thin-skinned compressional tectonics and low-angle normal faulting. Here, structures associated with the Sevier orogeny are easily distinguishable from the extensional faults because of their classical fold-thrust belt geometry and their detailed characterization in relatively undisturbed ranges along strike of the Sevier belt to the south (e.g., Burchfiel and others, 1974). Thus, this terrane offers an unusual opportunity to examine the influence of pre-existing compressional structures on large-magnitude extensional tectonics, as well as an opportunity to gain new insights on the extensional structures themselves.

Apart from the brief mentions of the Mormon Mountains by Spurr (1903) and Longwell (1926), the first to describe the geology of the map area of this report (Plate I, Fig. 1) was C.M. Tschanz (1959; Tschanz and Pampeyan, 1970). He mapped the area during the late 1950's and early 60's in reconnaissance as part of the geologic map of Lincoln County, in which he outlined the basic distribution of rock types. However, due to a lack of detailed stratigraphic information, limited field time and exceptional structural complexity, he was not able to identify the major structural features of the area. Olmore (1971) described and mapped in detail the structure of the East Mormon Mountains and a portion of the Tule Spring Hills. He interpreted the geology there in terms of large-scale, east-directed thrust faulting locally affected by "backslipping", but most of the faults he mapped are now known to be unrelated to the Sevier orogeny, and structures in the Tule Spring Hills that are related to it, but not thought to have experienced "backslippage" by Olmore, have been severely modified by Tertiary extensional deformation (Wernicke, in prep.). Olmore's interpretations, similar to those of many workers who interpreted low-angle normal faults in terms of compressional thrusting, predated major breakthroughs in the understanding of extensional tectonics in the Basin and Range made during the 1970's (for example, Anderson, 1971; Wright and

Troxel, 1973; Proffett, 1977). Nonetheless, his mapping has provided important constraints for interpreting the area considered in this report. Longwell and others (1965) mapped the southern portion of the Mormon Mountains in reconnaissance for the 1:250,000 geologic map of Clark County, and published several reports on the geology of the Muddy Mountains and North Muddy Mountains (Longwell, 1921, 1926, 1949, 1962). Ekren and others (1977) mapped part of the northernmost Mormon Mountains as part of a regional study of Tertiary rocks.

About nine months of fieldwork were required for this study, accomplished intermittently between June, 1979 and February, 1982. Mapping was done directly on three 1:24000 7.5-minute topographic maps published by the U.S. Geological Survey (Fig. 1), the boundaries of which are shown on Fig. 2 with respect to place localities in the map area mentioned in the text. Stratigraphic sections were measured with a Jacob's staff.

The Mormon Mountains can be reached from three gravel roads accessible from Interstate 15 (Fig. 1). The westernmost road branches east from U.S. Highway 93 just south of its intersection with the Union Pacific Railroad near Moapa and follows the railroad northward through Meadow Valley Wash. The next road to the east branches north from Interstate 15 at the Overton exit (four-wheel drive only), and the easternmost road branches north from Interstate 15 at the Carp exit. Access to the map area itself is only possible by four wheel-drive vehicle on the jeep trails shown on Figures 1 and 2, except for the easternmost part which can be reached directly from the Carp road.

The physiography of the Mormon Mountains is extremely rugged, with cliff faces hundreds of meters high present throughout the range. The overall aspect of the topography is that of a broad dome, ranging from about 900 m (3000 feet) elevation on the flanks to over 2250 m (7400 feet) at the crest. Due to the generally low elevation and weathering characteristics of carbonate rocks in arid environments, outcrop is virtually 100% along the flanks of the range and seldom below 50% at elevations above 1675 m (5500 feet), where pinyon, juniper, and locally Ponderosa forest flourish.

The climate of the map area ranges from intolerably hot (105°F+) at lower elevations during the summer to intolerably cold (usually below freezing) at higher elevations in the winter, but is essentially ideal for working year-round provided the appropriate elevation is selected (above 1500 m (5000 feet) in summer, below 1500 m (5000 feet) in winter). Rainfall is heaviest in the winter and spring, with frequent snows above 1500 m (5000 feet). Higher elevations are typically targets for spectacular late afternoon thundershowers during the month of August.

Wildlife, notably desert bighorn, deer, coyotes, jackrabbits, a variety of small rodents, bold, bluish-black wasps with bright orange wings, uncountable squadrons of bloodthirsty blackflies (early June only), and brownian herds of grasshoppers served to make life interesting during the study.

STRATIGRAPHY

The stratigraphy of the central Mormon Mountains consists of an approximately 2000 m thick Cambrian through Pennsylvanian transitional shelf-platform sequence which unconformably overlies Proterozoic crystalline rocks and is in turn overlain unconformably by Quaternary alluvial and landslide deposits (Plates I, II and III). The Paleozoic strata may be divided into three major sequences: a 284 m-thick lower terrigenous sequence, a 1205 m-thick middle dolomite sequence, and 511 m-thick upper limestone sequence. The lower terrigenous sequence consists of the Tapeats Sandstone and Bright Angel Shale (Lower to Middle Cambrian), the middle dolomite sequence is composed of the Bonanza King Formation, Nopah Formation, Pogonip Group, Ely Springs Dolomite, and the Ironside Dolomite Member of the Sultan Limestone (Middle Cambrian through Late Devonian), and the upper limestone sequence consists of the Valentine and Crystal Pass limestone members of the Sultan Limestone, the Monte Cristo Limestone, and the lower part of the Bird Spring Formation (Late Devonian through Pennsylvanian). The lower terrigenous sequence contains several thin carbonate units within the Bright Angel Shale, and the dolomite and limestone sequences are intercalated with several sandstone, siltstone and shale units.

As discussed below, the Paleozoic rocks occur within four major structural elements, which are, from bottom to top: the autochthon (relative to Mesozoic thrust faults), parauthochthonous slices beneath the Mormon thrust, rocks between the Mormon thrust and the Mormon Peak detachment, and rocks above the Mormon Peak detachment. Although the rocks in each of these elements have experienced sizeable tectonic transport with respect to one another, they are remarkably similar stratigraphically, with the exception of the Bonanza King Formation, as discussed below. A similar situation exists in thrusted transitional shelf-platform deposits to the south in the Spring Mountains (Burchfiel and others, 1974, Axen, 1980). To the west, in the Meadow Valley Mountains, Paleozoic rocks coeval with part of the Mormon Mountains section are much thicker and differ markedly in lithology (Tschanz and Pampeyan, 1970).

The Paleozoic strata of the Mormon Mountains are ideal for detailed structural analysis of a complexly deformed terrane, because the units are regionally laterally persistent in both thickness and lithology and exhibit extreme changes in lithologic character over very small stratigraphic intervals.

The thicknesses and lithologic descriptions below and in Plate III are taken from sections measured largely in the East Mormon Mountains in conjunction with F.G. Poole of the U.S. Geological Survey. The sections measured were chosen because of their lack of deformation and good exposure. The location of the measured sections is shown in Fig. 3. Although these sections are largely not in the area of Plate I, they are identical to those in the Mormon Mountains, except some of the units, particularly the Bonanza King Formation, appear to thicken westward. The measured sections are in the autochthon, and these same units may be slightly thicker above the Mormon thrust and Mormon Peak detachment. Unfortunately, most sections above the Mormon thrust and essentially all those above the Mormon Peak detachment are highly deformed, making thickness determinations for comparison with autochthonous sections difficult. In light of these considerations, the thicknesses reported below and in Plate III should be regarded as minimum.

The stratigraphic nomenclature used in this study was taken largely from the Paleozoic of the Spring Mountains region about 100 km southwest of the Mormon Mountains, with which the Mormon Mountains section is closely correlative. This nomenclature is only one of several possible, and correlation with sections other than in the Spring Mountains will be discussed below under the individual formation headings.

Precambrian metamorphic and igneous complex (p€m)

The stratigraphically and structurally lowest map unit in the central Mormon Mountains is a heterogeneous and complexly deformed assemblage of crystalline rocks including paragneiss, orthogneiss, pelitic schist, amphibolite, calc-silicate rock, granitic intrusions, lineated quartzites, and tourmaline-bearing, muscovite-rich pegmatite dikes. These rocks crop out in two main areas, a band along the west side of the range and a triangular-shaped area around the Whitmore Mine (Fig. 2, Plate I). No attempt was made to decipher the complex history of this rock assemblage. It is certain that all foliations and lineations in these rocks were formed prior to their peneplanation and deposition of the Lower Cambrian Tapeats Sandstone. Olmore (1971) reports a 1.7 Ga K-Ar date (mineral unspecified) on a muscovite pegmatite body in the East Mormon Mountains.

Tapeats Sandstone (€t; 145 m)

The Tapeats Sandstone, first described by Noble (1914) from exposures in the Grand Canyon, crops out in two bands in the map area, one along the west side of the range and another in the central part. It unconformably overlies Precambrian crystalline rocks in both areas, and is generally not detached from the basement as hypothesized by Olmore (1971). The Tapeats occurs only in the autochton. A sporadically occurring basal pebble conglomerate, never observed to be more than a few centimeters thick, contains clasts of granite, gneiss, and schist from the underlying crystalline complex. The Tapeats was measured to be 145 m thick in the East Mormon Mountains, but may be slightly thicker in the area of Plate I. Nowhere in the Mormon Mountains or East Mormon Mountains does the Tapeats approach a thickness of 244 m as reported by Wheeler (1943).

The lower two-thirds of the Tapeats consist of a grayish red, medium- to coarse-grained, thin- to thick-bedded arkosic sandstone commonly containing rip-up clasts, ripple marks, and cross-lamination. The upper third is similar to the lower third except the color is more commonly light brownish gray than reddish brown and is locally granular and conglomeratic. The uppermost 15 m commonly contains micaceous siltstone beds similar to the lower part of the Bright Angel Shale. Both the sandstones and siltstones are locally highly bioturbated and contain the trace fossil Scolithus. The name "Prospect Mountain Quartzite" was applied by Wheeler (1943), Tschanz and Pampeyan (1970) and Olmore (1971) to the unit described here. The term Tapeats Sandstone is preferred because the section exposed in the Mormon Mountains more closely resembles the type-section of the Tapeats in the Grand Canyon where it forms an 87 m thick veneer of sandstone overlying Precambrian crystalline rock than the type-section of the Prospect Mountain Quartzite defined by Hague (1892) in the Eureka District of central Nevada (minimum 500 m thick, that always overlies thick sequences of late Precambrian clastic rock; see also Stewart, 1970, p. 7).

The age of the Tapeats Sandstone is probably Early Cambrian in the Mormon Mountains, by analogy with exposures near the Arizona-Nevada border where it contains Early Cambrian fossils (McKee and Reeser, 1945). Specimens of linguloid brachiopods found in the upper third of the Tapeats Sandstone near Whitmore Mine indicate that at least this part is not

Precambrian. It is possible, however, that the base of the Tapeats extends into the Precambrian (Stewart, 1970).

Bright Angel Shale (Ꞓa; 139 m)

The Bright Angel Shale, first described by Noble (1914) for exposures along the Bright Angel trail in the Grand Canyon, crops out in two bands in the map area conformably overlying the Tapeats Sandstone, and does not occur in any structural level higher than the autochthon. The contact between the Tapeats and the Bright Angel is gradational over about 10 m, and in this study it is drawn at the horizon at which the amount of medium- to coarse-grained sandstone is exceeded by siltstone. This horizon occurs at the pronounced slope break between the Tapeats and Bright Angel. A thickness of 139 m was measured in the East Mormon Mountains, but the unit may be slightly thicker in the two bands which crop out in the area of Plate I.

The Bright Angel Shale may be divided into three units that probably correspond, from bottom to top, to the Pioche Shale, Lyndon Limestone and Chisholm Shale (Wheeler, 1943; McNair, 1951; Tschanz and Pampeyan, 1970), three formations originally defined in much thicker sections to the northwest (Walcott, 1908; Westgate and Knopf, 1932). Following Stewart's (1970) nomenclature for transitional shelf-platform facies rocks to the southwest, the name Bright Angle Shale is used here.

The lower unit (Pioche Shale) consists of 87 m of micaceous siltstone, shale and sandstone. The proportion of sandstone generally decreases upward, while that of shale increases. The sandstones are well-indurated, quartz-rich, locally glauconitic, yellowish brown weathering, medium- to very coarse-grained (locally granular), thin- to medium-bedded, locally cross-laminated, and at one locality contained calcarenite. The shales are argillaceous and are generally gradationally mixed with siltstones on a fine scale. Collectively they are light olive-gray to dark yellow-green, very thin to thin, wavy-bedded, micaceous, and form small recesses between slightly more resistant sandstone layers. The sandstones, siltstones and shales are all abundantly bioturbated.

The middle unit (Lyndon Limestone) consists of 14 m of medium- to dark-gray, mottled dark- to medium-gray weathering, fine- to medium-grained limestone laced with discontinuous, thin (0.5-1 cm) layers of orangish brown silty limestone. Trace fossils and *Girvanella* are abundant. The upper unit (Chisholm Shale) consists of 38 m of argillaceous, olive-gray shale and siltstone and medium brown, very thin interbeds of calcareous glauconitic quartzite. A one meter-thick bed of yellowish brown calcarenite is present near the top of the unit. In general, the upper unit contains a higher percentage of shale than the lower unit.

The age of the Bright Angel Shale in the Mormon Mountains is Early and Middle Cambrian. The Bright Angel Shale is wholly Middle Cambrian in the eastern Grand Canyon region, but to the west near the Arizona-Nevada border, it contains Early Cambrian (olenellid) faunas near the base (McKee and Reeser, 1945). Wheeler (1943) collected *Albertella* 12 m below the top of the lower unit (Pioche Shale) and *Glossopleura* from the upper unit (Chisholm Shale) that date the upper part of the Bright Angel in the East Mormon Mountains, as Middle Cambrian. Trilobites collected 15 m below the top of the Bright Angel in this study (81FP-159F, Fig. 3) were identified and assigned to the Middle Cambrian by M.E. Taylor.

Bonanza King Formation (Ꞓb; approx. 769 m)

The Bonanza King Formation was first named by Hazzard and Mason (1936) for exposures in the Providence Mountains, California (Fig. 1). It was subsequently divided into two members, the lower Papoose Lake Member and the upper Banded Mountain Member, in the Halfpint Range on the Nevada Test Site (Fig. 1) by Barnes and Palmer (1961). This subdivision proved to be quite useful on a regional scale, because the two members are divided by a distinctive 10 m thick bed of silty dolomite which is remarkably persistent over most of the southern part of the Cordilleran miogeocline, and is informally referred to here as the "silty unit." The Banded Mountain Member occurs at all structural levels, and is the stratigraphically oldest unit above the Mormon thrust and Mormon Peak detachment. The Papoose Lake Member occurs only in the autochthon, although at one locality not in the area of Plate I the silty unit is present immediately above the Mormon Thrust.

Lithologically, the Bonanza King Formation is characterized by interlayed bands of light and dark dolomite on a variety of scales (Fig. 4). The dark bands, which appear black from a distance, often contain a distinctive light and dark gray mottled texture caused by burrowing organisms (Fig. 5). The light bands, which appear white from a distance, are often either sugary-textured (caused by secondary dolomitization) or very finely laminated and, locally, cross-laminated.

Papoose Lake Member (€bp, approx. 200 m). The Papoose Lake Member may be divided into two units, a lower cliff-forming, massive-splitting limestone and dolomite (€bpl) and an upper unit consisting of alternating light and dark bands of dolomite (€bpd). The Papoose Lake Member crops out in two bands in the central Mormon Mountains above the Tapeats Sandstone and Bright Angel Shale. An accurate thickness of the Papoose Lake could not be obtained because it is without exception in both the Mormon Mountains and East Mormon Mountains highly faulted. Probably the least faulted section is one near the Whitmore Mine (D-D', Plate II) that has a structural thickness of about 200 m; the lower unit is 50 m thick and the upper unit is 150 m thick.

The lower unit (€bpl) consists of a dolomitic limestone at the base that grades upward into pure dolomite within 10-20 m of the top. The basal few meters consist of a mottled light- and dark-gray, fine-grained, laminated to very thin-bedded, massive-splitting limestone with small patches of dolomite. The limestones become gradually more dolomitized upsection, and the unit becomes locally recrystallized and coarse-grained but maintains its massive splitting character. The unit is pervaded, especially near the base, with the small (0.5-1 cm diameter), round algal form Girvanella. At the base of the unit these are usually replaced by dolomite or light brown-weathering silty dolomite. The top of the unit was mapped at a 3 m silty limestone interval which occurs approximately at the top of the cliffy dolomites.

The upper unit (€bpd) consists of alternating light and dark bands of very light-gray to dark-gray, fine- to coarse-grained, laminated to thick-bedded dolomite and silty dolomite.

Banded Mountain Member (€bbl-€bb5, 569 m). The Banded Mountain Member can be divided into 5 units (Fig. 4) which are present in the autochthon, Mormon thrust plate, parautochthonous slices beneath the Mormon Thrust, and above the Mormon Peak detachment. Nearly half of the outcrop of the central Mormon Mountains consists of rocks of the Banded Mountain (Plate I), and a detailed understanding of its internal stratigraphy is essential to understanding the structure of the range. In highly faulted sections above the Mormon thrust it was often difficult to distinguish the units, which usually were easily mapped in the structurally less complex autochthon. The Banded Mountain is 569 m thick in the East Mormon Mountains, and may thicken to as much as 750 m in the western Mormon Mountains, below the Mormon thrust. A complete section has not been measured above the Mormon thrust, but parts of some units can be identified and are similar in thickness to the autochthonous section. However, unit Cbb4, based on reconnaissance work south of the mapped area appears to be at least twice as thick in the upper plate of the Mormon Thrust than it is in the autochthon.

Unit €bb1 (116 m) consists of three parts, including, from bottom to top, the "silty unit" (9 m) discussed above, a black band (46 m), and a white band (61 m). The "silty unit" is an olive-gray, light brown-weathering wavy laminated silty dolomite, mottled with discontinuous lenses of dark brown-weathering silt and fine-grained quartz sand. The dark-gray band is composed of dark gray, medium-grained dolomite which contains minor amounts of silt and vug-filling, coarse-grained white dolomite. A 9 m-thick interval beginning about 13 m below the top of the black band contains a distinctive mottling of dark brown silt and fine sand. The white band is composed of light gray, very light gray-weathering, fine-to medium-grained, thinly laminated dolomite and silty dolomite. Silty portions of the white band weather light brown and very pale orange. The diagnostic features of this unit are the presence of the silty unit and dark brown siliceous mottled dolomite at the top of the black band.

Unit €bb2 (81 m) is divisible into a lower black band (approximately 70 m) and an upper sequence of interbedded white and black bands (11 m). The black band is a medium to dark gray, medium to light olive-gray-weathering, fine- to medium-grained, wavy-laminated to thin-bedded, blocky splitting, burrowed (forming characteristic Bonanza

King mottling, Fig. 5) dolomite. The upper 11 m of this unit contains light olive- to yellow-gray silty dolomites, and the uppermost 4 m is a conspicuous light olive-gray, very light-gray, pinkish and yellowish gray-weathering, thinly laminated dolomite. The lower part of unit Єbb2 tends to form steep slopes, whereas the upper 10 m tends to form more subdued, ledgy slopes.

Unit Єbb3 (177 m) is composed of a series of alternating white and black bands that are generally thinner (5-30 m) than those in units Єbb1 and Єbb2 (50-100 m). At the base of the unit, immediately above the conspicuous white band at the top of Єbb2, is a 3 m-thick bed of brownish gray, light gray-weathering dolomite which contains distinctive, curious "bathtub"-shaped silicified algal laminites (Fig 6). These occur sporadically throughout the unit, but are always present and most abundant in the basal portion of Єbb3. The remainder of the unit consists of light to dark gray, very fine- to coarse-grained, laminated to very thin-bedded, relatively silt-free dolomite which frequently displays characteristic Bonanza King burrow mottles (Fig. 5). The measured thickness of unit Єbb3 is 117 m and is probably a minimum for this unit, because it was cut by two faults where measured in the East Mormon Mountains. The uppermost part of Єbb3 is a 6 m-thick bed of silty dolomite.

Unit Єbb4 (143 m) is characterized by a higher percentage of siltstone and clastic carbonate than the other units. The base was mapped at a conspicuous, 8 m-thick dark gray-weathering, thin-bedded dolomite which occurs between two 6 m-thick, light gray, light brown-weathering, fine- to coarse-grained, laminated to thin bedded silty and sandy dolomites. These three units form a 20 m-thick "brown-black-brown" triad which is easily recognizable throughout the Mormon Mountains and East Mormon Mountains.

The lower silty dolomite of the triad corresponds to the upper silty dolomite in unit Ꞓbb3. The basal triad is succeeded by 44 m of light gray, light olive-grey-weathering, medium- to coarse-grained, thin-bedded, dolomitized calcarenite which is commonly characterized by a distinctive oolitic texture with 1-3 cm wide lenses of fine-grained dolomite and silty dolomite. Above this unit is a 10 m thick, slope-forming, pale yellow-brown, grayish orange-weathering, fine- to medium-grained, laminated to very thin-bedded dolomitic siltstone with tabular rip-up clasts. These rocks are overlain by 76 m of light to dark gray, fine- to coarse-grained, laminated to very thin-bedded, blocky to massive splitting dolomite that locally contains burrow mottling. In the western Mormon Mountains, these dolomites appear white from a distance, intercalated with two thin black bands, one about 35 m below the top of unit Ꞓbb4, the other about 50 m below the top. In the East Mormon Mountains, where secondary dolomitization has been more intense, the color banding is not distinctive. The lower black band is characterized by large-scale, cuspate, convex-upward wavy laminations with an amplitude of several meters and wavelength of about 10 m. Immediately below this horizon, in both the East Mormon and Mormon Mountains, is a 6 m-thick bed of light brownish gray, yellowish gray-weathering, wavy laminated silty dolomite.

Unit Ꞓbb5 (110 m) forms a prominent, cliffy black band in the western Mormon Mountains, and the contact between this unit and the underlying white band of uppermost Ꞓbb4 is very easily mapped (Fig. 4). However, as in unit Ꞓbb4, secondary dolomitization has obscured this color contrast in the East Mormon Mountains. The rather uniform lithology of this unit is a medium to brownish gray, light olive- to dark gray-weathering, fine- to coarse-grained, laminated to thin-bedded, burrow mottled, massive-splitting detrital dolomite. The upper 40 m of this unit contains columnar stromatolites, and, significantly, the first appearance of bedding-parallel chert lenses and stringers which are common in much of the overlying Paleozoic section.

The term "Peasley Limestone" was applied to unit Ꞓbpl by Wheeler (1943) and Olmore (1971). Wheeler did not discuss any units above the "Peasley", and Olmore informally designated everything above Ꞓbpl and below the Pogonip Group equivalent in the East Mormon Mountains as undifferentiated Cambrian dolomites. The Peasley Limestone is now defined as the basal member of the Highland Peak Formation (Merriam, 1964). The Highland Peak contains much more shale and limestone and is much thicker than coeval rocks in the Mormon Mountains. Bonanza King nomenclature has been widely applied to much more similar sections to the southwest in the Spring Mountains and Muddy Mountains (e.g. Burchfiel and others, 1974) and is thus applied here. The Bonanza King Formation forms the lowest part of Tschanz and Pampeyan's (1970) "undifferentiated Devonian to Cambrian limestones and dolomites".

Trilobite-brachiopod hash is common in clastic layers of the Bonanza King, and a few small intact gastropods were collected from the base of Ꞓbb5. The presence of Middle Cambrian fossils in the underlying Bright Angel Shale (Wheeler, 1943; this report) and early Late Cambrian fossils in the overlying Dunderburg Shale (according to regional studies) indicate a Middle Cambrian age for most of the Bonanza King Formation, and a Late Cambrian age for an unknown amount of its upper part. Material collected from the dolomitic siltstone in unit Cbb4 (81FP-160F; Fig. 3) did not contain any diagnostic fossils.

Nopah Formation (€n; 102 m)

The Nopah Formation was named by Hazzard (1937) for exposures in the Nopah Range of Inyo County, California. The Nopah Formation occurs in the autochthon, parautochthonous slices beneath the Mormon thrust and above the Mormon Peak detachment, but does not occur between the Mormon thrust and the Mormon Peak detachment within the mapped area. The Nopah Formation may be divided into two parts, a 12 m-thick lower siliceous unit, the Dunderburg Shale Member, and a 90 m-thick upper dolomite unit. The Dunderburg Shale, originally named by Hague (1883) from exposures in the Eureka District of east-central Nevada, is a remarkably widespread unit in the Cordilleran miogeocline. In the map area, it consists of a medium to light gray, slope-forming silty dolomite with grayish orange to pink angular rip-up clasts of silty dolomite, interbedded with greenish-gray medium-to coarse-grained glauconitic sandstones. The top 3 m consists of platy dolomitic siltstone and some shale.

The upper dolomite unit consists of light olive-gray, fine- to coarse-grained, wavy laminated to very thin-bedded dolomite mottled with yellowish brown, yellowish gray and grayish orange-weathering, fine-grained silty dolomite. At various intervals the dolomites contain bedding parallel white-weathering chert stringers. Solution collapse breccias (Fig. 7), which locally occur in every Paleozoic formation from the Nopah Formation to the Sultan Limestone, contain clasts up to several meters in diameter. The lower 53 m of the upper dolomite unit is generally cliffy and blocky splitting, while the upper 37 m has a well-bedded appearance and is slightly recessive.

No fossils were found in the Nopah Formation in the Mormon Mountains, but an Upper Cambrian age has been documented in many other areas by trilobite faunas in the Dunderburg Shale and lower Ordovician (Canadian) fossils the lower part of the overlying Pogonip Group (see below). Rocks mapped as Dunderburg Shale in this study may not be true Dunderburg, because age control in this part of the section is poor. An alternative possibility would be to consider the dolomitic siltstone 58 m above the base of Cbb4 to be Dunderburg. In this case, the overlying €bb4 and €bb5 would be correlative to the Nopah, and rocks now mapped as Nopah would belong to the Pogonip Group.

Pogonip Group (Op; 152 m)

The Pogonip Group, redefined by Nolan and others (1956) after originally being named by King (1878) from exposures in the Eureka District of east-central Nevada, forms a distinctive orangish brown-weathering recessive unit, above the more resistant, medium gray beds of the Nopah Formation. This major slope break and color change is prevalent throughout the southern Great Basin. In the map area, the Pogonip Group is exposed in the autochthon, parautochthonous slices beneath the Mormon thrust, and above the Mormon Peak detachment. The Pogonip may be divided into three units in the Mormon Mountains, a lower silty dolomite (87 m), a middle limestone and shale (18 m), and an upper oncolitic dolomite (47 m).

The lower unit consists of light olive to light-grey, pale orange, yellow- and pink-weathering, fine- to coarse-grained, laminated to thin-bedded dolomite with intervals containing white weathering nodular chert. This unit contains abundant intraformational solution collapse breccias (Fig. 7). A few thin beds of dolomitic siltstone, glauconitic sandstone, and bioclastic calcarenite are present.

The middle unit is composed of medium gray, fine- to coarse-grained, very thin wavy bedded, fossiliferous limestone interbedded with light olive-gray, laminated to very thin-bedded shaley mudstones.

The upper unit is composed of light brownish gray to light olive-gray, brownish, pinkish, and yellowish gray-weathering, medium-grained, laminated to thin-bedded, oncolitic (Girvanella) silty dolomite (Fig. 8).

The upper unit lithologically resembles part of the Aysees Member (Byers and others, 1961) of the Antelope Valley Limestone (Nolan and others, 1956) as defined by Guth (1979) in the Sheep Range (about 80 km southwest of the Mormon Mountains), because of its abundant Girvanella and a distinctive gastropod fauna dominated by Palliseria and Maclurites.

The middle and upper units are generally highly fossiliferous, containing abundant gastropods, brachiopods, and orthocone cephalopods. Faunas from the Pogonip in the East Mormon Mountains collected by Tchanz and Pampeyan (1970, p. 27) were identified by A.R. Palmer and E.L. Yochelson and were considered to be Lower Ordovician. H.G. Johnson (Olmore, 1971, p. 10) identified Sowerbyella, an index fossil to the Ordovician, from a section in the East Mormon Mountains (location not specified) from the Pogonip Conodonts collected for this study were identified by John Repetski. Sample 81FP-161F, collected from the base of the Pogonip, yielded 16 species of conodonts which clearly belong to Fauna D, which is Lower Ordovician (Canadian), and indicating these rocks are correlative with the Goodwin Limestone as defined by Nolan and others (1956). Sample 81FP-158F yielded 13 species belonging to the lower part of Fauna E of Late Canadian age, indicating time-equivalence to the Ninemile Formation of Nolan and others (1956). Sample 81FP-164F contained 18 species which are Late Fauna E in age, which corresponds to Latest Canadian time, indicating that these beds also correlate with the Ninemile

Formation, and indicate that the upper oncolitic, silty dolomite is definitely correlative with the Antelope Valley Limestone.

Eureka Quartzite (3-5 m)

Overlying the Pogonip Group (but included in map unit Ou because of its thinness) is a unit lithologically correlative with the Eureka Quartzite, originally named by Hague (1883) but later redefined by Kirk (1933). It consists of a light brownish gray, thin-bedded silty dolomite interbedded with light olive-gray and light brown, fine- to coarse-grained cross-laminated quartzite.

Conodants processed from Sample 81FP-162F collected from this unit were identified by Anita Harris, who found 15 very incomplete elements of *Aphelognathus* cf. *A. politus* (Hinde). She states (written commun., 1982):

"*Aphelognathus politus* together with about 3 other species is one of the chief components of the lower part of the Ely Springs Dolomite...[in southern Nevada] the basal part of the Ely Springs Dolomite contains sandy to sandstone intervals that represent reworked Eureka sands... [these sands] yielded biostratigraphically diagnostic conodonts of middle Late Ordovician age (Maysvillian)... Your sandy dolostone to dolarenite is undoubtedly part of this phase of the Ely Springs Dolomite."

Thus, a time-stratigraphic equivalent of the Eureka is not present in the Mormon Mountains, but rather a thin, much younger, reworked Eureka sand in the lowest beds of Ely Springs Dolomite.

Ely Springs Dolomite (Oe; approx. 108 m)

Overlying the orange-weathering Pogonip Group and the reworked Eureka Quartzite is a light gray unit which is correlative with the Ely Springs Dolomite, originally defined by Westgate and Knopf (1932). It occurs in the autochthon, in parautochthonous slices beneath the Mormon thrust, and above the Mormon Peak detachment, but not between the Mormon thrust and the Mormon Peak detachment. It is composed of medium gray, medium to light gray-weathering, aphanitic to medium-grained, laminated to thin-bedded dolomite with distinctive 1-5 mm diameter vugs filled with coarse-grained white crystalline dolomite. This unit commonly weathers to a texture reminiscent of a frothy basalt flow. The upper part of this unit is commonly mottled light and medium gray dolomite which can be mistaken for parts of the Bonanza King Formation. This unit frequently contains solution collapse breccias similar to the Pogonip Group, except the matrix is often pure quartz sand rather than silty dolomite.

The presence of a Maysvillian conodont fauna at the base of the Ely Springs and a major regional disconformity between Devonian and Ordovician rocks (e.g. Burchfiel and others, 1974; Longwell and Mound, 1967; see below) at the top of the formation indicate that it is entirely middle- to late-Upper Ordovician.

Sultan Limestone (Ds; 202 m)

The Sultan Limestone, named by Hewett (1931) for exposures in the Goodsprings District in Clark County (Fig. 1), occurs in the autochthon, parautochthonous slices beneath the Mormon thrust, and above the Mormon Peak detachment. All three members of the Sultan Limestone as defined by Hewett can be recognized in the Mormon Mountains: the lower Ironside Dolomite, the middle Valentine Limestone, and the upper Crystal Pass Limestone. The Ironside Dolomite was mapped as unit Dsi, and the Valentine and Crystal Pass Limestones were mapped together as unit Ds (Plate I).

Ironside Dolomite (Dsi; 54 m). The Ironside Dolomite is one of the most distinctive lithologic units of the southern shelf-platform transitional facies of the Cordilleran miogeocline. In the Mormon Mountains, it consists of two units. The lower one, which forms a slope above the more resistant beds of the Ely Springs Dolomite, consists of grayish orange-weathering, fine-grained, laminated to thin-bedded silty dolomite interbedded with light olive to pinkish gray, medium- to coarse-grained dolarenite with a quartz sandstone bed at the base. The upper one is a medium gray to olive-black, olive-gray to light olive-gray-weathering, fine- to coarse-grained, laminated to thin-bedded dolomite with characteristic bedding parallel rows of 1-2 cm diameter vugs filled with white coarsely crystalline dolomite. It is locally fossiliferous, containing brown silicified stromatoporoids, oncolites, and *Amphipora*. Stromatoporoids are not as abundant in the Ironside Dolomite in the Mormon Mountains as they are to the south in the Spring Mountains.

Valentine Limestone (Ds; 79 m). The Valentine Limestone may be divided into two units, a lower dolomite and an upper limestone with their contact about 30 m above the base, although at some localities the Valentine is entirely dolomite and this subdivision cannot be made. The lower unit is a medium to light gray, light olive-gray to light gray, fine- to coarse-grained dolomite with interbeds of quartz sandy dolomite and large bodies of dark brown silicified quartz sand and algal structures. The upper unit consists of light olive-gray, sometimes pinkish gray-weathering, aphanitic to very fine-grained, very thin-bedded, massive-splitting, stylolitic quartz-sandy limestone with some remains of silicified brachiopods, gastropods, and corals. Both the upper and lower units contain conglomerates with clasts up to boulder size set in a matrix of quartz sand and limestone, perhaps formed by solution collapse.

Crystal Pass Limestone (Ds; 69 m). The distinctive Crystal Pass Limestone is a brownish-gray, banded light and medium gray-weathering, aphanitic to very fine-grained, laminated to very thin-bedded styloitic limestone. A distinctive 5 m-thick marker horizon 30 m below the top of the Crystal Pass Limestone is a very pale orange, pinkish gray to brown-("desert varnish") weathering, cross-laminated calcareous quartz sandstone that forms a prominent light-colored ledge. Another interval, 8 meters thick beginning 11 meters below the top of the Crystal Pass Limestone, consists of varicolored, silty to clayey, thin-bedded limestone that forms a prominent bench. Immediately below this interval is a horizon containing very large (up to 5 cm) silicified brachiopods. One of the forms in this interval was identified as Cyrtospirifer (B.C. Burchfiel, pers. commun., 1979), an index to the Late Devonian. Locally, the Crystal Pass Limestone exhibits a conglomeratic texture similar to that of the Valentine Limestone. In its lower part, scattered fine-to coarse-grained, rounded quartz sand grains are common.

The age of the Sultan Limestone is Middle (?) Devonian at the base and Latest Devonian at the top. In addition to Cyrotospirifer, sample 81FP-165F, collected from the same horizon yielded Late Devonian (Late Famennian) conodonts (Fig. 3) according to C.A. Sandberg. Conodonts at the base of the lower unit of the Ironside in the southern part of the East Mormon Mountains (sample 81FP-157F; sample 81FP-163F, collected from the same horizon in the Mormon Mountains, was barren; Fig. 3) were identified by John Repetski, and included Pandorinellina cf. P. insita (Stauffer), which ranges from latest Middle Devonian to Late Devonian. To the south in the Spring Mountains, Harrington (1982) assigns the Ironside Dolomite to the Late Devonian. Finer resolution of the age of the basal Ironside in the Mormon Mountains must await further sampling of this interval.

Sample 81FP-158F, also taken from the southern part of the East Mormon Mountains (Fig. 3c) and stratigraphically only a few meters below sample 81FP-157F, contained Late Early Ordovician conodonts (Late Canadian) as discussed above. Thus, between this locality and the north part of the East Mormon Mountains, the Ely Springs Dolomite, reworked Eureka Quartzite and Antelope Valley Limestone equivalent are all eliminated beneath the sub-Middle (?) Devonian unconformity. A tectonic boundary with perhaps several kilometers of right-lateral offset ("Moapa Peak Shear Zone" of Olmore, 1971) exists between the two areas, but the rapid disappearance of over 150 m of section at the disconformity may be largely a stratigraphic effect.

The Sultan Limestone is equivalent to the Muddy Peak Limestone in the Muddy Mountains (Longwell and others, 1965).

Monte Cristo Limestone (Mm; approx. 279 m)

The Monte Cristo Limestone was named by Hewett (1931) from exposures in the southern Spring Mountains. All five members defined by Hewett were recognized in the Mormon Mountains by Langenheim (1956), who raised it to Group status in the Arrow Canyon Range (Langenheim and others, 1962). The Monte Cristo Limestone is present in the autochthon, in parautochthonous slices beneath the Mormon thrust, and above the Mormon Peak detachment.

Dawn Limestone (Mmd; 78 m). The cliff-forming Dawn Limestone may be divided into two units of roughly equal but variable thickness, a lower limestone and an upper cherty limestone. The lower limestone unit consists of alternating brownish and medium light-gray, olive and light olive-gray-weathering, very fine- to very-coarse grained, thin- to medium-bedded bioclastic to fossil fragmented limestone containing numerous solitary and colonial corals, gastropods and brachiopods. The upper unit is lithologically similar but contains 10 to 20 percent dark-gray, light brown-weathering, lensy silicified limestone. Both units have been altered by secondary dolomitization and weather medium- to dark-gray. The Dawn Limestone is commonly cross-laminated.

Anchor Limestone (Mma; 35 m). The slope forming, lithologically distinctive Anchor Limestone gradationally overlies the Dawn Limestone and is composed of medium to dark gray-weathering, fine-grained, laminated to very thin-bedded limestone with roughly half of the formation composed of very thin beds of silicified limestone and chert (Fig. 9).

Bullion Limestone (Mmb; approx. 110 m). Overlying the recessive Anchor Limestone is the Bullion Limestone characteristically forming spectacular cliffs. It is a brownish gray, light olive to medium gray-weathering, very fine- to coarse-grained, thin- to thick-bedded, locally bioclastic crinoidal limestone with several horizons in which nodular chert and silicified limestone make up 10-20% of the rock.

Arrowhead Limestone (3 m). The thin Arrowhead Limestone lies between the Bullion and Yellowpine Limestones, and was not mapped separately because of its thinness. It is a brownish black, yellowish to light olive-gray-weathering, aphanitic to very fine-grained, very thin-bedded limestone with pinkish orange weathering clayey-silty wavy laminae which form partings between limestone beds. Although thin, the Arrowhead Limestone usually forms a conspicuous recessive seam between the massive cliffs formed by the Bullion and Yellowpine Limestones.

Yellowpine Limestone (Mmy; approx. 53 m). The Yellowpine Limestone is essentially identical to the Bullion Limestone, and mapping the two as separate members depends largely on being able to recognize the Arrowhead Limestone. Usually, however, the Yellowpine Limestone contains very large solitary corals (up to 10 cm long) which are not typically present in the Bullion Limestone.

The Monte Cristo Limestone is equivalent to the Redwall Limestone of the Grand Canyon region and the Rogers Spring Limestone in the Muddy Mountains (Langenheim, 1963).

It should be noted that the thicknesses of Mississipppian rocks measured for this report are in disagreement with those measured by Langenheim (1963), who reported for the members, from bottom to top 15 m, 43 m, 116 m, 5 m, and 94 m (total 273 m) compared to 78 m, 35 m, 110 m, 3 m and 53 m (total 279 m) for this report. In that exactly the same section was measured here and by Langenheim, it is clear to me that Langenheim's Dawn-Anchor thickness is simply too thin by a factor of two and his Yellowpine thickness is too thick by a factor of two.

The Monte Cristo Limestone is primarily Early Mississippian in age, possibly extending into early Late Mississippian time (Longwell and others, 1965).

<u>Bird Spring Formation</u> (MPb; 200 m+).

A few tens of meters of the lowermost Bird Spring Formation are exposed in the autochthon below the Mormon thrust and above the Mormon Peak detachment. It is light to dark gray, light to light olive-gray weathering, aphanitic to coarse-grained cherty limestone with interbeds of quartz sandstone. It forms a characteristic "staircase" outcrop pattern. The basal beds of the Bird Spring Formation are probably latest Mississippian in the Mormon Mountains, based on studies in the Meadow Valley Mountains (Longwell and others, 1965). The top of the Bird Spring Formation is clearly Permian in age (Longwell and others, 1965), but the youngest beds within the map area are probably no younger than

Pennsylvanian. The base of the Bird Spring Formation probably rests disconformably on early Upper Mississippian beds of the Yellowpine Limestone, but no physical evidence for such an unconformity was observed.

Tertiary-Quaternary Landslide and Alluvial Deposits (TQls; Qa)

Overlying strata of all ages with marked angular unconformity are Tertiary (?) - Quaternary landslide and alluvial deposits. The landslides are extremely poorly sorted, angular, well-cemented breccias which are present in two localities: one on the west side of the map area and one (or perhaps several) just west of the "corral" in the South Fork of Toquop Wash (Fig. 2; Plate I). Clast composition of both these landslide masses reflect bedrock composition upslope from them. Their lower contacts are characteristically highly irregular. No effort was made to distinguish various different types of alluvial cover other than the landslide debris.

STRUCTURE

Four major structural levels produced by large-scale thin-skinned faulting of the miogeoclinal package are divided by, from bottom to top, faults beneath parautochthonous slices below the Mormon thrust, the Mormon thrust, and the Mormon Peak detachment. These will be referred to here, from bottom to top, as the autochthon, parautochthon, Mormon thrust plate and the Mormon Peak allochthon (Fig. 10). The parautochthon and Mormon thrust plate were emplaced over the autochthon by east-directed thrusting during the Mesozoic Sevier orogeny (Armstrong, 1968). In Miocene time, both the autochthon and thrust stack were dismembered by a complex system of west-dipping, low-angle normal faults, the largest of which is the Mormon Peak detachment. The normal faulting caused varying amounts of eastward tilting of rock units in the map area, such that in general structurally lower parts of the Sevier orogen (autochthon and Precambrian basement) lie to the west and structurally higher parts (parautochthon and Mormon plate) lie to the east.

Mormon Thrust System

The Mormon thrust system, exposed in the eastern half of the mapped area (Plate I, Fig. 3) is composed of the Mormon thrust plate, which is exclusively made up of the Banded Mountain Member of the Bonanza King Formation, and structurally lower parautochthonous slices which contain every formation from unit 4 of the Banded Mountain to the Yellowpine Limestone. The geometry of detachment in the Mormon thrust system indicates that it was formed by east-directed, ramp-décollement style thrusting. The thrust faults are generally either bedding-parallel or cut up-section to the east (Plates I and II). Figure 11 depicts the geometry of the thrust system by showing, in map view, the lithologies immediately below ("footwall map") the sole of the thrust system. This map indicates that the base of the thrust system overrides Mississipiian rocks in western areas of exposure, and then ramps up across Pennsylvanian rocks to the east. Although bedding parallel detachment in the Mississippian over at least 5 kilometers parallel to transport direction (shown later to be northeast) is clear, it is not possible to determine whether a relatively steep ramp (>30°) occurs in the Bird Spring Formation in the eastern portion of the map. However, previous mapping (Olmore, 1971, Tschanz and Pampeyan, 1970) and reconnaissance mapping outside the map area indicates that the Mormon thrust system is constrained to override Mesozoic strata in the Tule Spring Hills and East Mormon Mountains not more than 10 km to the east, and is observed to cut upsection at an angle of about 30° through Permian rocks exposed in a window in the northeastern Mormon Mountains.

It is curious that detachment in the footwall over much of the Mormon Mountains occurs in the Bullion and Yellowpine Limestones, both massive, relatively coarse grained, poorly bedded units, rather than in the well-bedded, finer-grained limestones immediately above and below (Plate III). Apparently the décollement chose to localize in the least mechanically anisotropic unit in that part of the section.

The hanging wall of the Mormon thrust is similarly enigmatic. It detached everywhere in strong dolomites of the Banded Mountain Member of the Bonanza King Formation, largely in units 2 and 3 to the west and unit 4 to the east, when only a few hundred meters below the thrust, the Bright Angel Shale apparently remained intact. A similar geometry of décollement thrusting exists to the south in the Spring Mountains (Burchfiel and others, 1981; Willemin and others, 1981), where the major thrusts are detached just a few hundred meters above the Bright Angel Shale in strong Bonanza King dolomites. This observation in the Mormon Mountains combined with Bohannon's (1981) mapping to the south in the Muddy Mountains defines a 200 km-long segment of the Cordilleran foreland fold and thrust belt in which the major frontal décollement ignored strength anisotropies in the sedimentary wedge and broke the "hard way" (Fig. 12).

The parautochthonous slices beneath the Mormon thrust dip northeastward and are complexly folded by northwest-trending, northeast vergent overturned folds (Plate I, Plate II, section F-F', Fig. 13). Because bedding in the parautochthonous slices generally intersects their bounding faults at a high angle, their source must have been a ramp between the Cambrian and Mississippian décollement portions of the thrust system.

The fact that the parautochthonous slices dip northeast strongly suggests that the ramp from which they were stripped had a northwest strike. The strike of the rotated ramp slices should mimic the strike of the ramp from which they were derived. Also, only a northeast transport direction of the slices would produce the northeast-overturned folds (Fig. 14). Thus, a major phase of northeasterly transport, estimated to be N60°E±30° (Fig. 14) is indicated for the Mormon thrust system. In addition, Figure 11 suggests that the ramp east of the Mississippian décollement portion of the thrust system also strikes northwest because the line of pinchout of autochthonous Bird Spring Formation strikes northwest. Ramps need not always strike perpendicular to transport direction, but the consistency of this observation with the northeast vergent folds in the parautochthon is supportive of northeast transport. The northeast transport direction of the Mormon thrust system suggested here does not take into account any possible rotations of the entire system about a vertical axis after its emplacement.

No systematic pattern of folding was observed in the Mormon thrust plate. Both southeast- and northeast-overturned folds are present, along with both northeast- and northwest-striking, open to tight upright folds, with no preferred axial orientation.

Folds in the autochthon were observed at one locality (Fig. 11, location B). Here, recumbent folds directly beneath the Mormon thrust are overturned to the southeast (Plate I). The vergence of these folds suggests that the Mormon thrust system may have had a period (or periods) of southeast transport. A thrust fault between two parautochthonous slices (location A, Fig. 11) has been folded and overturned to the north. This indicates that folding of slices, and therefore at least part of the movement on the Mormon thrust, followed imbrication of these two parautochthonous slices.

The location of the ramp between the Cambrian and Mississippian décollement portions of the Mormon thrust is unknown. The source terrane (original underpinnings) of the Mormon thrust plate must lie west of westernmost autochthonous outcrops of its oldest units, which are Banded Mountain dolomites (Fig. 11), but the upper segment of the ramp may have overlapped these exposures. The easternmost possible position of the top of the ramp corresponds to the western limit of exposures of autochthonous Monte Cristo limestone, which essentially corresponds to the westernmost outcrops of the Mormon thrust itself (Fig. 11). It seems likely that the ramp lay over the western part of the mapped area, because older-over-younger imbrications within the Bonanza King Formation there are common, particularly in the area west of Davies Spring (Plate I). These slices could be interpreted as parautochthonous slices torn away from the ramp. However, these imbrications are intimately associated with younger-over-older, west-directed normal faults beneath the Mormon Peak detachment. Thus, the older-over-younger relationships may simply be a consequence of juxtaposing pieces of higher thrust allochthons onto the autochthon by normal faulting and be unrelated to the original position of the Cambrian-Mississippian ramp.

An important aspect of the thrusted terrane is that it provides a fairly accurate means by which to measure the amount of tilting due to the Miocene extensional event. Cross-sections and reflection seismic profiles of thrust belts which have not been disrupted by later events typically show that the autochthonous sedimentary package and bedding-parallel portions of the basal thrust fault are very gently inclined (0-5°) away from the foreland (e.g., Bally and others, 1966; Cook and others, 1979). Thus, although no post-Sevier pre-extensional sedimentary deposits are present in the mapped area, elements of the thrust belt itself serve as an approximate guide to pre-extension paleohorizontal.

No data within the mapped area yield information as to the precise timing of thrust faulting. Regional constraints, discussed in Wernicke (in prep.), suggest a mid-Jurassic to Late Cretaceous age for compressional deformation. The total amount of shortening represented by the Mormon thrust system is at least 30 km, based on reconnaissance outside the map area, also discussed in Wernicke (in prep.).

Eastern Imbricate Normal Fault Belt

A complex group of west-dipping low-angle normal faults and associated high angle faults that disrupt the thrust terrane is present in the eastern part of the map area. The total amount of extension accommodated by these faults increases from only a few hundred meters in the north to perhaps as much as 4 km in the south.

This is shown clearly by the Horse Spring fault (Plate I, Plate II, Figs. 10 and 15) which has approximately 3 km of offset near Whitmore Mine (Plate II, D-D'), and completely dies out only 6 km to the north. Along section A-A', the Horse Spring fault may be represented by a series of normal faults which cut autochthonous Devonian and Mississippian rocks and basal slivers of the Mormon thrust system, but cumulative extension along these faults is only about 200 m.

A conspicuous kinematic problem of the Horse Spring fault is the difference in tilt between the hanging wall and footwall in its central area of exposure, as seen in section D-D'. As noted in Wernicke and Burchfiel (1982), planar normal faults cannot cause differential tilt between hanging wall and footwall unless the blocks deform internally in some way. Concave upward listric faults, because they are curved, force differential rotation and tilt the hanging wall more steeply than the footwall. Since footwall strata are inclined 20-30° more steeply than hanging wall strata, and no evidence was found for significant internal deformation of the fault blocks, neither the planar nor concave upward listric models are adequate.

Palinspastic restoration of section D-D' elucidates the basic problem (Fig. 16): the fault cuts through the footwall section much more quickly than the hanging wall section, which requires that the two were not initially in direct contact. It seems that a wedge of material must have once lain between the two. The reconstruction in Fig. 16 was done so as to minimize the volume of this wedge by restoring the hanging wall to its easternmost possible position. The simplest kinematic scheme to account for this missing wedge is to have it squeezed out to the west from between the hanging wall and footwall, similar to propelling a watermelon seed out from between the thumb and index finger. The geometry of the faults bounding the wedge derived from the reconstruction indicates that the upper one was an east-directed thrust fault emplacing older rocks over younger. The lower, more steeply dipping fault, in conjunction with the Horse Spring fault, forms a downward steepening listric normal fault zone, whch serves to accomodate differential tilt between the hanging wall and footwall. From this, it could be generalized that relatively undeformed regions are separated from imbricate normal fault systems by concave upward listric fault zones if the blocks dip toward the boundary (Wernicke and Burchfiel, 1982) and by concave downward listric faults if the blocks dip away from the boundary, as appears to be the case here.

Drag folding directly beneath the Horse Spring fault is well developed where it crosses the Bright Angle Shale-Bonanza King Formation contact, along a segment between Horse Spring and Hackberry Spring. Here the footwall rocks are folded over until locally they dip to the west. Numerous bedding plane faults within the limestone unit of the Papoose Lake Member are apparently folded over, and slices of basal Papoose Lake are discordantly faulted across the Bright Angel Shale (sections F-F' and B-B').

The Horse Spring fault dips west to northwest north of Wiregrass Spring, but abruptly changes to a southwest dip of about 30 degrees in its southern exposures. In its southeasternmost exposures, the fault becomes difficult to define, and is represented on Plate I as a vastly oversimplified zone of faults in the vicinity of hill 4322. Chaos (in the sense of Noble, 1941) is spectacularly developed along southeastern segments of the fault, particularly in a canyon 0.5 km due north of the petroglyphs, where several hundred meters of section are omitted across a complex fault zone (Fig. 17; see Wernicke and Burchfiel, 1982, for an explanation of the genesis of chaos).

In general, the southern portion of the eastern imbricate belt is much more complex than the relatively simple tilted-block geometry of the northern portions. Near bedding-parallel faults are far more abundant, and brecciation is so intense that in most instances it took considerable searching to find an outcrop suitable for taking an attitude. Despite the busy appearance of Plate I in this area, it barely begins to document the actual degree deformation.

Faults structurally beneath the Horse Spring fault generally do not display differential tilt between hanging wall and footwall, and are thus believed to be of planar rotational type as described in Wernicke and Burchfiel (1982). If the Paleozoic rocks in these fault blocks were roughly horizontal prior to extension, then the extensional strain in the eastern imbricate belt is approximately 75 percent increase over original width (Fig. 3, Wernicke and Burchfiel, 1982), assuming an average dip of bedding of 40° and an average fault dip of 35°. Combined with the offset on the Horse Spring fault of no more than about 3.0 km, the total extension along section D-D' of the eastern imbricate belt is approximately 4 km. The total extension along F-F' is a similar amount, since a reconstruction of the section should not allow overlap of the Mormon thrust. It is striking that such a small amount of extension can cause the severe deformation observed in the southern part of the area.

The direction of extension in the eastern imbricate belt cannot be determined with precision, but two lines of evidence suggest that it is east-northeast. The first is the sense of rotation of autochthonous

Paleozoic rocks. In areas where slickenside striae are abundant within imbricate normal fault terranes, their trend consistently parallels the dip direction of rotated strata (e.g., Anderson, 1971; Davis and others, 1980; Davis and Hardy, 1981). Thus, assuming that the autochthonous Paleozoic section was roughly horizontal or gently inclined to the west prior to imbricate normal faulting, the extension direction may be inferred from their sense of tilt. Shown in Figure 18 is a histogram of tilt direction of autochthonous sedimentary rocks from which a maximum elongation direction of N80°W±20° can be estimated. The second, in accord with the first result, is derived from two intersecting fault planes which do not offset each other, located 1.7 km ENE of peak 7083 (just north of section D-D'). Regardless of the chronology of movement on these two faults, this geometry is only possible if the slip-line of at least one of the faults is parallel to their line of intersection. The orientations of both fault planes are well constrained by measurement in the field (as shown on Plate I), and their line of intersection trends S82°W and plunges 25° (Fig. 19), in surprisingly good agreement with the extension direction based on the average dip direction of rotated strata.

As discussed in Wernicke (in prep.), a large, east-directed subhorizontal normal fault exposed in the Tule Spring Hills and East Mormon Mountains must underlie the Mormon Mountains at a depth of no more than a few kilometers, and thus probably functions as a basal detachment for the eastern imbricate belt. By inspection of Plate I, the eastern imbricate belt and inferred basal detachment unmistakeably involve autochthonous crystalline basement and thus their formation was not influenced by the Mormon thrust, despite their close structural proximity. Figure 20 shows one fault of the eastern imbricate belt which cuts the Mormon thrust and rotates it and the underlying autochthonous Mississippian rocks to the east. These facts are contrary to the popular notion that low-angle normal faults developed within older compressional thrust belts are reactivated thrust faults because the thrusts represent planes of weakness. The eastern imbricate belt and its basal detachment developed discordantly across the older structural grain, suggesting that (1) pre-existing "planes of weakness" are not required for the initiation of low-angle normal fault systems and (2) pre-existing structural features such as thrust faults do not always influence the localization of low-angle normal fault systems. The relationships observed here suggest that the fracture of crystalline basement rocks had a lower yield stress than "backsliding" on the Mormon thrust.

Central High-Angle Fault System

In the central, structurally simplest part of the map area, a system of high-angle faults are present in the autochthon structurally above the Horse Spring fault but below the Mormon Peak detachment (Fig. 21). The faults are, with a few exceptions of small normal displacement (10-100 m), and dip between 50° and 70°. Some of the faults change orientation along strike. Collectively, they account for very little extensional strain, but their spatial organization seems to be consistent with the approximately S80°W extension direction inferred from the eastern imbricate normal fault system, as discussed below.

The faults may be divided into two sets: a north-northeast striking set and a northwest striking set (Fig. 22). The apparent sense of throw on the north-northeast striking set is consistently down to the west-northwest, and on the northwest striking set, generally down to the southwest. This pattern is best developed away from the Horse Spring fault. As one moves to the northeast toward the fault, the orientation and sense of throw of the faults become more irregular, and they tend to become more closely spaced.

Assuming average strikes of N49°W and N23°E (Fig. 22) and a dip of 60° for the two sets in the more organized, western part of the system (most of the scatter, particularly in the northwest striking set, are from faults to the north adjacent to the Horse Spring fault), they plot on a stereogram as a near perfect conjugate shear system (Fig. 23), with σ_3 oriented at N76°E35, trending nearly parallel to the principal elongation direction inferred from the eastern imbricate normal fault system (Figs. 18 and 19). Unfortunately, actual slip direction on these faults is unknown due to lack of offset linear features and slickenside striae, so it is not possible to rigorously support this interpretation. Another, albeit more complex interpretation, would be to consider slip on the faults as purely down-dip and to call attention to the fact that the two sets occur largely (though problematically for this interpretation, not completely) in separate spatial domains: a northwest-striking set to the south and the north-northeast-striking set to the north. In this case, two separate principal elongation (and presumably, least principal stress) directions could be inferred, and perhaps a third corresponding to the domain of north-northwest striking faults in the northeast part of Figure 21.

The fact that one of the faults cuts the Mormon Peak detachment at one end and is truncated by it at the other strongly suggests synchronous movement on the high-angle faults and at least the final stages of movement on the Mormon Peak detachment (Plate I; fault A, Fig. 21). None of the high-angle faults can be shown to postdate final movement on the Horse Spring fault.

Mormon Peak Detachment

The structurally highest tectonic unit in the map area is the Mormon Peak allochthon, marked at its base by a conspicuous topographic ledge formed by the Mormon Peak detachment (Fig. 24). The detachment surface forms a smoothly contoured dome (Fig. 25) which mantles the topography of the range. Shown on Figure 25 is a set of topographic contours which serve to constrain the projection of the detachment east of peak 6365 (it must project above ridges consisting of lower plate strata). The amplitude of the dome is about 1 km over the width of the map area.

Beneath the detachment, a similar broad domal structure is expressed by the orientation of autochthonous Paleozoic strata. Fig. 26a shows strike and dip symbols from Plate I in this domal structure, without the central high-angle fault system and formational contacts. Figure 26b shows a structure contour map on the base of the Tapeats Sandstone in the same area, drawn without attempt to extrapolate faults of the central high-angle fault system to depth. Figure 26b should not be taken as a literal interpretation of the position of the Tapeats, because in the southeastern part of the diagram it is almost certainly cut out on the Horse Spring Fault. Rather, it is intended to depict the present form of the domal structure, and the base of the Tapeats was chosen as a convenient datum. Although Figures 26a and 26b have approximately the same form, they differ in that the contour on the Tapeats has an eastward elongate domal crest while the structural form indicated from attitudes shows a localized crest. The localized crest in Figure 26a is coincident with the western part of the elongate crest in the Tapeats, and the eastern part of the elongate crest is the locus of maximum relative uplift of northeast- and northwest-striking faults of the central high-angle fault system that are upthrown to the east. The difference in form of the two diagrams may thus be attributed to the high-angle faulting. Since the locus of uplift of the central high-angle fault system does not coincide with the center of the well-defined dome in bedding attitudes, the two structures probably did not form synchronously.

An important problem is the relationship between the domal structure expressed in the autochthonous Paleozoic section and the domal form of the detachment surface. Inspection of Figures 25 and 26 suggests that the two structural forms are certainly not perfectly coincident, the detachment appearing to be much smoother and having less total relief. Figure 27 shows two cross-sections comparing the form of the lower plate strata and the detachment, and it is clear, particularly in cross-section A-A', that if the curvature of the detachment were removed, a pronounced warping of the autochthonous Paleozoic would remain. Although there is no *a priori* reason to assume that the detachment was initially planar (whereas Paleozoic bedding obviously was), the crude coincidence (on the scale of the range) of the lower plate doming with the detachment fault dome suggests that the two may be an expression of the same process. It is clear that the warping of the Paleozoic section predated a substantial amount of movement on the detachment. If the folding of the lower plate accompanied the early history of movement on the detachment there must have been enough movement thereafter to "smooth out" the form of the detachment by truncation and removal of the early-domed surface. Alternatively, the warping of lower plate strata may be related to Mesozoic deformational events, in which case its spatial association with the domed detachment is merely coincidental. I favor the former hypothesis, on the basis of the occurrence of a number of geometrically similar Tertiary detachment terranes in the southern Basin and Range province, that have domes in lower plate mylonitic foliation that are remarkably coincident with domes in the detachments, but the lower plate structures are in detail clearly discordant to the faults (e.g., Davis and others, 1980; Rehrig and Reynolds, 1980).

Since it has been shown that (1) the central high-angle fault system was active during the final stage of emplacement of the Mormon Peak allochthon, (2) the doming of lower plate strata significantly predates these stages, and (3) the doming of lower plate strata and the development of the central high-angle fault system are probably diachronous, it follows that doming largely preceded development of the central high-angle fault system. Both doming and faulting are here interpreted to be synchronous with movement on the Mormon Peak detachment. I further speculate that if lower plate warping, which must represent true folding, accompanied early movement on the detachment, the present domed configuration of the fault is an expression of later folding, rather than an initially curviplanar surface of breakage.

The detachment cuts downsection toward the west in the footwall. The easternmost klippe in the map area rests upon the Mormon thrust plate only a few hundred meters above the base of the Mormon thrust system, and the next klippe to the west upon autochthonous Devonian and Mississippian rocks. The remainder of the klippen to the west cut gradually downward through the autochthon, until the westernmost exposed klippen rest upon Precambrian basement (Plates I and II). As discussed above, the décollement-style thrust system provides a reasonably good pre-extension horizontal. Assuming a regional dip of 3° for the Tapeats Sandstone,

Figure 28 shows the approximate position of the Mormon Peak detachment relative to the Tapeats. The detachment cuts downward about 4 km over a horizontal distance of 18 km, indicating the initial dip of the detachment averaged across the map area was about 10°.

In common with the eastern imbricate normal fault system, intense brecciation occurs adjacent to the detachment, and in both the upper and lower plates persists for several tens of meters away from it. The breccias are characterized by considerable clast rotation and highly variable grain size and sorting (Fig. 29). In contrast, the thrust faults of the Mormon thrust system generally do not have clast-rotated breccias asociated with them. Where present they are volumetrically minor, and usually occur within a few meters (rather than tens of meters) of the faults. These observations are consistent with those of a number of workers in the Basin and Range province who have noted the same difference in degree of fracturing between compressional and extensional faults (e.g., Guth, 1981; Seager, 1970). In cases where it is ambiguous whether a low-angle fault is a thrust or a low-angle normal fault, the presence of large amounts of clast-rotated breccia favors the latter.

In addition to pervasive brecciation, the Mormon Peak allochthon is highly internally deformed, and might best be characterized as a tectonic megabreccia. Consistently, bedding parallel faults omit stratigraphic section, high-angle faults are normal, and both either tangentially merge with or are truncated by the detachment. The allochthon as a whole, however, is essentially composed of an intact Paleozoic section ranging from mid-Banded Mountain Member dolomites at the base up to the

Mississippian and Pennsylvanian Bird Spring Formation at the top, and has been broken into a series of mildly tilted normal fault blocks (Fig. 30). Although the Paleozoic sections within the blocks were probably somewhat deformed during the Sevier orogeny, no evidence was found within any of the blocks to suggest major pre-block faulting deformation in the map area, although mapping by Tschanz and Pampeyan (1970) in the Meadow Valley Mountains to the west suggests that substantial Mesozoic compressional deformation is present within the Mormon Peak allochthon there (Fig. 31). In the Mormon Mountains, the normal faults separating the tilted blocks generally dip opposite the tilt direction, as would be expected from simple imbricate normal faulting of a near-horizontal package of rock. In addition, at several localities in the north part of the Mormon Mountains about 5-10 km north of the northern boundary of the map area, tilted Tertiary ignimbrites rest disconformably or with very slight angular unconformity on the Bird Spring Formation in the Mormon Peak allochthon. For these reasons, it appears that the magnitude and direction of tilting of Paleozoic rocks in the Mormon Peak allochthon in the map area is largely the result of Tertiary imbricate normal faulting.

The direction of tilt of the blocks, shown in Figures 32 and 33, is separable into east and west tilted domains. The west tilted domains show a preferential tilt toward the southwest, while east tilted domains show no preference between southeast and northeast (most of the southeast tilts occur in blocks near peak 6365; Fig. 30).

The boundaries of the tilt domains in the western part of the map area where they can be seen within the allochthon trend roughly east-northeast, parallel to the tilt direction of the blocks (Figures 32 and 33). The change in polarity of tilting across the boundaries does not take place along a discrete fault zone, but seems to be accommodated along a diffuse zones (Plate I). Two domain boundaries in the central and eastern part of the map area trend perpendicular, or at least at a high-angle, to tilt direction, and are of "antiformal" type, since the blocks dip away from the boundary (Stewart, 1980). The basic pattern of tilt-parallel and tilt-perpendicular boundaries observed in the Mormon Peak allochthon is highly analogous, and perhaps homologous, to regional tilt patterns observed in the Basin and Range province (Rehrig and others, 1980; Stewart, 1980).

The east-northeast extension direction inferred from the tilt direction of upper plate rocks is consistent with that inferred from the eastern imbricate belt and central high-angle fault system. The slightly scattered pattern of tilts observed in Fig. 32 is what might be expected from an imbricate normal fault terrane superimposed on a package of randomly (but not steeply) tilted strata.

As in the footwall, the Mormon Peak detachment cuts downsection to the west in the Mormon Peak allochthon. Klippen in the western part of the map area contain sections with up to several hundred meters of Banded Mountain at their base. Further east, in the klippe which includes Mormon Peak, the section is only as old as the Pogonip Group with the exception of a small sliver of Banded Mountain. In the easternmost klippe, the oldest strata are the upper part of the Ely Springs Dolomite. An important exception to this general trend, as well as the apparent lack of compressional structures within the Mormon Peak allochthon in the map area, occurs in the klippe which includes peak 7365 (Plate II, cross-section A-A'), where two small slivers are present, one consisting of Banded Mountain and the other of Nopah. These thin slivers, along with a conformable (but tectonically abbreviated) overlying section including rocks as young as Mississippian, rest upon a slice of the Sultan Formation, which in turn lies directly on the Mormon Peak detachment.

The above relationships suggest that the minimum offset on the Mormon Peak detachment is about 15 km (Fig. 34). Since the footwall is detached in mid-Banded Mountain strata of the Mormon thrust plate in the easternmost exposures of the detachment, the nearest possible matching rocks in the hanging wall lie west of the line of pinchout of suitably thick Banded Mountain sections. Although thin scraps of Banded Mountain lie east of line a-a' on Figure 34, they are not thick enough to provide a suitable replacement for the large amount of strata missing in the footwall. These slices were torn away from the base of the allochthon and left in an intermediate position between the more complete hanging wall sections to the west and their original underpinnings to the east, in a manner analogous to the formation of chaos structure.

Matching Banded Mountain strata in the Mormon Peak allochthon with Banded Mountain in the Mormon thrust plate requires that the Paleozoic section in the allochthon be part of the Mormon thrust plate. However, three independent lines of evidence suggest that these rocks were derived from a higher tectonic element of the Sevier orogen, the Glendale-Muddy Mountain thrust plate.

The first line of evidence is based on regional considerations of the position of the allochthon. Reconnaissance mapping by me and by Tschanz and Pampeyan (1970) outside the area of Plate I indicates that the Mormon Peak allochthon is structurally continuous with folded late Paleozoic and early Mesozoic strata exposed in the Meadow Valley Mountains (Fig. 31). These folded rocks are part of a regionally continuous synclinorium of folds and small thrust faults that occurs in the upper plate of the Keystone-Muddy Mountain-Glendale thrust plate and involves Paleozoic rocks of miogeoclinal affinity (Fig. 12). In common with the Muddy Mountains and Spring Mountains (Bohannon, 1981; Burchfiel and others, 1974; Axen, 1980), the southern Mormon Mountains contain two major thrust faults developed in Paleozoic strata of more cratonal character than those in the folded belt (Armstrong, 1968). The Mormon thrust plate dips gently southward to the south of the mapped area. About 10 km south of the southernmost klippe of the Mormon Peak allochthon, Banded Mountain dolomites of the Glendale thrust plate are thrust over the Jurassic Aztec Sandstone and the Cretaceous (?) Overton fanglomerate (Wernicke, in prep.; Bohannon, in press). I believe (based on reconnaissance work) that these Mesozoic rocks form a contiguous stratigraphic sequence with Cambrian strata to the north in Mormon thrust plate. Therefore, the Mormon thrust is not the highest thrust fault in the frontal part of Sevier orogen at the latitude of the Mormon Mountains; yet rocks in the Mormon Peak allochthon are, suggesting that the Mormon Peak allochthon could not be derived from the Mormon thrust plate, but instead must be part of the Glendale-Muddy Mountain plate.

The second line of evidence is the juxtaposition of a Cambrian through Mississippian section above a slice of Devonian limestone in the klippe of the Mormon Peak allochthon containing Peak 7365, discussed above. In view of the consistent deformational style of stratigraphic omission and juxtaposition of higher structural levels on lower that accompanies the Mormon Peak detachment, the anomalous Devonian slice may be a fragment derived from the Mormon thrust plate, indicating that overlying Cambrian thrugh Mississippian rocks are part of a higher thrust plate than the Mormon plate. Alternatively, however, it is possible that the Devonian slice is a little-transported fragment from the autochthon. Yet another possibility is that the slice is part of the Glendale plate that was faulted beneath the Cambrian dolomites by an anomalous structurally-lower-over-higher imbrication during transport. Thus, the Devonian slice is suggestive, but not compelling, evidence in favor of the present hypothesis.

A third piece of evidence is the occurrence of a small slice of well-indurated carbonate-pebble conglomerate beneath the Mormon Peak allochthon along the detachment at the northern point of the klippe containing peak 6365. This conglomerate does not resemble any known Paleozoic rocks in the region, nor does it obviously correlate with any Early Triassic through Early Jurassic deposits in the area. It does, however, appear similar to thin, sub-thrust, syn-orogenic conglomerates of Jurassic and Cretaceous age present to the south in the Spring Mountains and Muddy Mountains in places where the thrusts have overridden the land surface (Axen, 1980; Brock and Engelder, 1977; Burchfiel and others, 1981). On the tentative basis that these are sub-thrust conglomerates, their present tectonic position is most easily explained if they were derived from beneath an overland portion of the structurally higher Glendale thrust, because the Mormon Peak detachment intersects the Mormon thrust only where it is subterranean.

In view of the circumstantial nature of these three lines of evidence, and their dependence on reconnaissance mapping, the assignment of Paleozoic rocks in the Mormon Peak allochthon to the Glendale thrust plate must be regarded as tentative until further detailed mapping of the region is completed. If true, then the amount of structural section omitted across the detachment in its western areas of exposure could approach 10 km, a thickness two and a half times that of the Cambrian through Cretaceous section. It also implies that the minimum offset on the Mormon Peak detachment is much greater than the 15 km required simply to match Cambrian sections of the Mormon Peak allochthon with those in the Mormon thrust plate.

The age of extensional faulting in the map area is mid- to late-Miocene. Reconnaissance mapping by me and by Ekren and others (1977) in the northern Mormon Mountains indicates that imbricately normal-faulted mid- to late-Miocene volcanic rocks are present in the Mormon Peak allochthon (Fig. 31). The highest unit in the volcanic section, resting conformably on mid-Miocene ash-flow tuffs, is a basaltic andesite that yielded a whole-rock K-Ar age of 8.5±0.3 Ma (Armstrong, 1970, sample 203I). Assuming this to be its true age, it follows that the Mormon Peak detachment must have still been moving at this time, since upper plate normal faults either merge with or are truncated by the detachment. Although the volcanic sections are conformable, lack of angular unconformities does not necessarily indicate that extensional tectonism had not begun prior to eruption of the basaltic andesite. In fact, the Miocene section in the Meadow Valley Mountains is still flat lying (Fig. 31), even though it has been transported at least 15 km to the west-southwest on the Mormon Peak detachment. Thus, no data in the immediate area directly bear on the precise time of onset of extensional tectonism. It is probable that extension in the Mormon Mountains was partly coeval with west-southwest directed extension in the Lake Mead area, over 60 km to the south of the mapped area, described by Anderson (1971; 1973), Anderson and others (1972), and Bohannon (1979; in press). They have shown on the basis of structural and sedimentologic analysis of synorogenic clastic and volcanic deposits, that the most intense west-southwest directed extension occurred between about 14 and 10 Ma.

Unconformably overlying the extended terrane are flat-lying clastic rocks of the Muddy Creek Formation, which fills the modern Basin and Range valleys. The maximum age of these deposits in the vicinity of the Mormon Mountains is unknown, and it is therefore possible that low-angle normal faulting in the Mormon Mountains continued well past 8.5 Ma, perhaps into the Pliocene, but this is not likely in view of Anderson's and Bohannon's work.

Similar to the eastern imbricate fault system and its inferred basal detachment, the Mormon Peak detachment does not appear to be spatially coincident with older Mesozoic thrust faults. With respect to the footwall, the detachment cuts obliquely through the Mormon thrust plate, autochthon, and Precambrian basement, an extremely unlikely trajectory for a former thrust fault. As noted above, with respect to the hanging wall, the detachment appears to cut upsection to the east rather than follow a single stratigraphic horizon. In view of the remarkably consistent décollement within Cambrian dolomites in the hanging walls of frontal Sevier thrusts to the south in the Muddy Mountains and Spring Mountains (Longwell and others, 1965; Burchfiel and others, 1974, Fig. 12), it seems unlikely that the base of the Mormon Peak allochthon is a reactivated thrust.

DISCUSSION

The Mormon Mountains have many of the characteristics of a widespread group of terranes which have been termed "metamorphic core complexes" (Davis and Coney, 1979), including a domal range form, a domal, structurally-higher-over-lower, low-angle dislocation surface veneering the topography of the range, and imbricate normal faulting both above and below (e.g., Shackelford, 1980) the dislocation surface. Unlike the metamorphic core complexes, rocks beneath the dislocation surface show no indication of ductile deformation or mylonitization. As hypothesized by Davis and Coney (1979) and Rehrig and Reynolds (1980), the deformation in the lower plates of the metamorphic core complexes, characteristically consisting of a shallowly-dipping mylonitic foliation containing a mineral lineation which trends remarkably parallel to the upper plate extension direction, serves as an accommodating mechanism for upper plate extension. However, in some of the core complexes, the mylonitic lineations were shown to predate upper plate imbricate normal faulting by more than 40 Ma (Davis and others, 1980). In others, the tectonites formed only 5-10 Ma prior to imbricate normal faulting (Reynolds and Rehrig, 1980; Keith and others, 1980). In either case, the data seem to be incompatible with models which view lower plate ductile strain simply as a synchronous, *in situ* accommodation of upper plate imbricate normal faulting, a mechanism of crustal extension suggested by many workers prior to the recognition of the metamorphic core complex terranes (e.g., Wright and Troxel, 1973; Proffett, 1977). There has been considerable debate among workers in the lineated terranes as to whether or not the tectonites are genetically related to upper plate extensional deformation and the basal detachments themselves. Geologists who regard the two events as separated by several tens of million years do not consider the two to be genetically related, (e.g., Davis and others, 1980), while those who interpret them as being separated by only a few million years favor models suggesting they are part of the same deformational system (e.g., Coney, 1980). The data presented here indicate that all of the attributes of Mormon Mountains geology in common with metamorphic core complexes described above need not be genetically related to the formation of a mylonite tectonite, and provides further documentation for the conclusion of Davis and others (1980) that doming, imbricate normal faults, and large, low-angle, structurally-higher-on-lower dislocation surfaces need not be associated with the formation of mylonite tectonite. The occurrence of an unmetamorphosed Paleozoic section beneath the Mormon Peak detachment conclusively shows that the lower plate was extended only a few hundred meters (in the north) increasing to no more than 4 km (in the south) during emplacement of the Mormon Peak allochthon. Thus, lower plate extension cannot be invoked as a driving mechanism for upper plate translation and extension, since the Mormon Peak allochthon has been translated more than 15 km relative to the lower plate. A mega-landslide or gravity-slide mechanism for the emplacement of the Mormon Peak allochthon is equally untenable, because there clearly are no "toe" relationships indicative of shortening in ranges to the west. In the Meadow Valley Mountains, there seems to be no indication of any complex mid- to late-Miocene deformation. As mentioned above, throughout the Meadow Valley Mountains and in Meadow Valley Wash, mid-Miocene ash-flow tuffs as old as 12.5±0.3 Ma (K-Ar, sanidine, sample 203F, Armstrong, 1970) and an underlying undated lacustrine limestone lie in near horizontal depositional contact upon folded and east-vergent thrust faulted Paleozoic and Mesozoic strata (Tschanz and Pampeyan, 1970; Ekren and others, 1977), and thus could not have been significantly deformed as a result of movement on the Mormon Peak detachment (Fig. 31). It is thus geometrically required that the detachment continue into the subsurface to the west beneath the Meadow Valley Mountains.

The 10° initial dip of the detachment deduced from footwall geometry and the observation that the detachment cuts stratigraphically upward to the east in the hanging wall indicates that the Mormon Peak allochthon was wedge-shaped. The apparent westward structural simplification of the allochthon is therefore probably due to its increased ability (because of its increased thickness) to resist internal fragmentation resulting from frictional drag on the detachment. A similar westward structural simplification exists in the lower plate, with the eastern imbricate belt giving way westward to the structurally simple central high-angle fault terrane. Perhaps this also is a product of a west-thickening, wedge-shaped allochthon above the detachment which probably exists at depth beneath the eastern imbricate belt. The overall picture of extensional strain is thus envisioned as large-scale, east-over-west simple shear of two imbricate, wedge-shaped slabs which are penetratively extended by a variety of extensional fault types (Wernicke and Burchfiel, 1982) concentrated in the thin, eastern portion of the slabs (Fig. 35). This is precisely the mechanism of large-scale strain accommodation usually observed in areas which have experienced large-scale compression (e.g., Milnes and Pfeiffner, 1980), except that the slabs are bounded by normal faults instead of thrust faults (Wernicke, 1981).

An observation with important implications for current models of Basin and Range extensional structure is the pattern of imbricate normal faulting in the Mormon Peak allochthon. Tilt direction of imbricate normal faults has commonly been used as the sole criterion to determine the polarity of the slip vector of basal detachments (e.g., Davis and Hardy, 1981; Dokka and Glazner, 1982), and some workers have used tilt geometry to infer complex patterns of ductile stretching and laminar flow within the middle and lower crust (e.g., Rehrig and others, 1980; Hamilton, 1981). The map data summarized in Figure 32 suggest that (1) tilt polarity may not be a reliable indicator of transport polarity, and (2) complex tilt patterns, similar to those observed throughout the Basin and Range province, may occur within a single extensional allochthon and be unrelated to any structures present in the footwall. The most reliable indicator of the polarity of the slip vector on detachments is the direction of stratigraphic and/or structural descent of the detachment, as is clearly discernable for the Mormon Peak detachment.

SUMMARY

Following conformable sedimentation during Early Cambrian through Early Jurassic time in the Mormon Mountains area, large-scale, east-directed, décollement-style thrust faulting disrupted the miogeoclinal wedge during the Late Jurassic (?) through Late Cretaceous Sevier orogeny (Armstrong, 1968). The Mormon thrust was active during this time, and moved to the northeast, dislocating slabs of Cambrian through Mississippian rocks from a ramp, folding them about northwest-trending axes, and accreting them to a autochthonous Mississippian décollement surface. In accord with regional relationships (Burchfiel and Davis, 1971; Carr, 1980; Axen, 1980; Bohannon, in press), the Glendale plate was subsequently emplaced on top of the Mormon plate.

Following a period of presumed tectonic quiescence from Paleocene through early Miocene (?) time, large-scale, thin-skinned extensional tectonism began to dismember the Sevier orogenic belt in a west-southwest direction, as inferred from several independent kinematic arguments. The major exposed extensional structure, the Mormon Peak detachment, developed obliquely across the major thrust plates of the Sevier orogen rather than reactivating the thrust faults. It has a minimum displacement of 15 km, and juxtaposes the highest structural levels of the Sevier orogenic belt with Precambrian crystalline basement of the autochthon. An inferred, structurally lower detachment of perhaps equal importance formed beneath the eastern imbricate belt within Precambrian crystalline basement, and thus also did not reactivate a pre-existing thrust. The Mormon Peak detachment formed with a dip of approximately 10 degrees to the west, and the initial stages of movement on the fault are interpreted to have been accompanied by warping of the entire system. Late in its history of movement, as broad warping continued, the central high-angle fault system became active, further deforming the system, and was locally active following final emplacement of the Mormon Peak allochthon. Based on cross-cutting relationships, the eastern imbricate belt was active after final emplacement of the easternmost klippe of the Mormon Peak allochthon, and the last movements on the easternmost faults of the central high-angle fault system. These relationships do not preclude partially synchronous movement on the Mormon Peak detachment and faults in the eastern imbricate belt, but do suggest that the emplacement of the Mormon Peak allochthon largely predated the development of the eastern imbricate belt. A complex pattern of tilt domains within the Mormon Peak allochthon apparently developed independently of lower plate structures. The current eastward dip of about 20° of the detachment in the easternmost klippe (Plate I) is probably the result of rotational imbricate normal faulting in the eastern imbricate belt.

After thin-skinned extensional tectonism, which may have persisted past 8.5 Ma, or perhaps in its waning stages, the modern topography developed, apparently as a result of both broad warping (e.g., the west side of the map area, Meadow Valley Wash and the Meadow Valley Mountains, Fig. 31) and high-angle "classic" Basin and Range faults (e.g., east side of the East Mormon Mountains, Fig. 1). The sequence of thin-skinned normal faulting followed by the formation of widely-spaced (10-30 km) basins and ranges has been observed throughout the Basin and Range by many workers (e.g., Zoback and others, 1981). The total amount of crustal extension accommodated within the mapped area is a minimum of 15-20 km, and perhaps a great deal more.

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Figure Captions

Figure 1. Index map showing access roads and location of study area with respect to major tectonic features and mountain ranges mentioned in text. E = Eureka District; H = Halfpint Range; G = Goodsprings district; N = Nopah Range; P = Providence Mountains; 1 = Rox NE quadrangle; 2 = Moapa Peak NW quadrangle; 3 = Davidson Peak quadrangle.

Figure 2. Localities in Plate I referred to in text, superimposed on contacts from Figure 10. Mormon Peak allochthon is shaded.

Figure 3. Locations of measured sections and fossil localities. Symbols at extremes of sections correspond to oldest and youngest map units (from Plate I) measured; symbols in parentheses indicate map unit from which fossils were collected. Topography and roads from U.S. Geological Survey 7.5-minute topographic maps.

Figure 4. Typical light-and-dark banding of the Bonanza King Formation, showing subdivisions of the Banded Mountain Member. Looking west from near peak 6365.

Figure 5. Characteristic pattern of burrowing found in dark colored dolomites of the Bonanza King Formation. Lens cap is about 5 cm in diameter.

Figure 6. Distinctive bathtub-shaped silicified algal laminites which define the base of unit Cbb3.

Figure 7. Breccias in the Pogonip Group, apparently formed by the collapse of solution cavities. These breccias are common in every unit from the Nopah Formation to the Sultan Limestone, but are particularly common in Ordovician rocks. Lens cap is approximately 5 cm in diameter.

Figure 8. Girvanella in Middle Ordovician (Antelope Valley Limestone-equivalent) part of the Pogonip Group. Lens cap is about 5.0 cm in diameter.

Figure 9. Intercalated chert and limestone of the Anchor Limestone. Lens cap is about 5.0 cm in diameter.

Figure 10. Major structural units in the central Mormon Mountains.

Figure 11. Map showing footwall geology of the Mormon thrust system. Barbed line shows map trace of the base of the thrust system from Plate I.

Figure 12. Major thrust plates of the Sevier orogenic belt in southern Nevada. The two lowest thrusts consistently detach in Bonanza King dolomites. Between the two highest thrusts is a broad, regional synclinorium consisting of miogeoclinal rocks which have been folded and faulted on small thrusts.

Figure 13. Northeast-vergent, recumbent isoclinal fold in Devonian rocks within the parautochthon, about 1 km north of the corral.

Figure 14. Axes of macroscopic, flexural-slip folds in parautochthonous slices suggesting northeastward transport of the thrust system.

Figure 15. View looking north of Horse Spring fault, placing Ordovician to Mississippian rocks over Cambrian rocks and Precambrian basement.

Figure 16. Reconstruction showing space problem that exists between hanging wall and footwall of the Horse Spring fault. Stratigraphic labels from Plate I.

Figure 17. Chaos structure along Horse Spring Fault near the petroglyphs, where two faults cut out all of the undifferentiated Ordovician and Valentine and Crystal Pass members of the Sultan Formation.

Figure 18. Tilt direction histogram of eastern imbricate belt suggesting

Figure 28. The descent of the Mormon Peak detacment from just above the Mormon thrust in eastern exposures to the base of the Tapeats Sandstone 18 km to the west suggests an initial dip of about 10° for the detachment.

Figure 29. Typical clast-rotated breccias associated with extensional faulting in the Mormon Mountains. Large boulder in upper right corner is about 0.5 in diameter.

Figure 30. View south toward peak 6365 (high peak on left) showing mildly tilted imbricate normal fault blocks in Mormon Peak allochthon. Note the topographic ledge formed by the detachment, and the structural simplicity of Banded Mountain strata in the footwall.

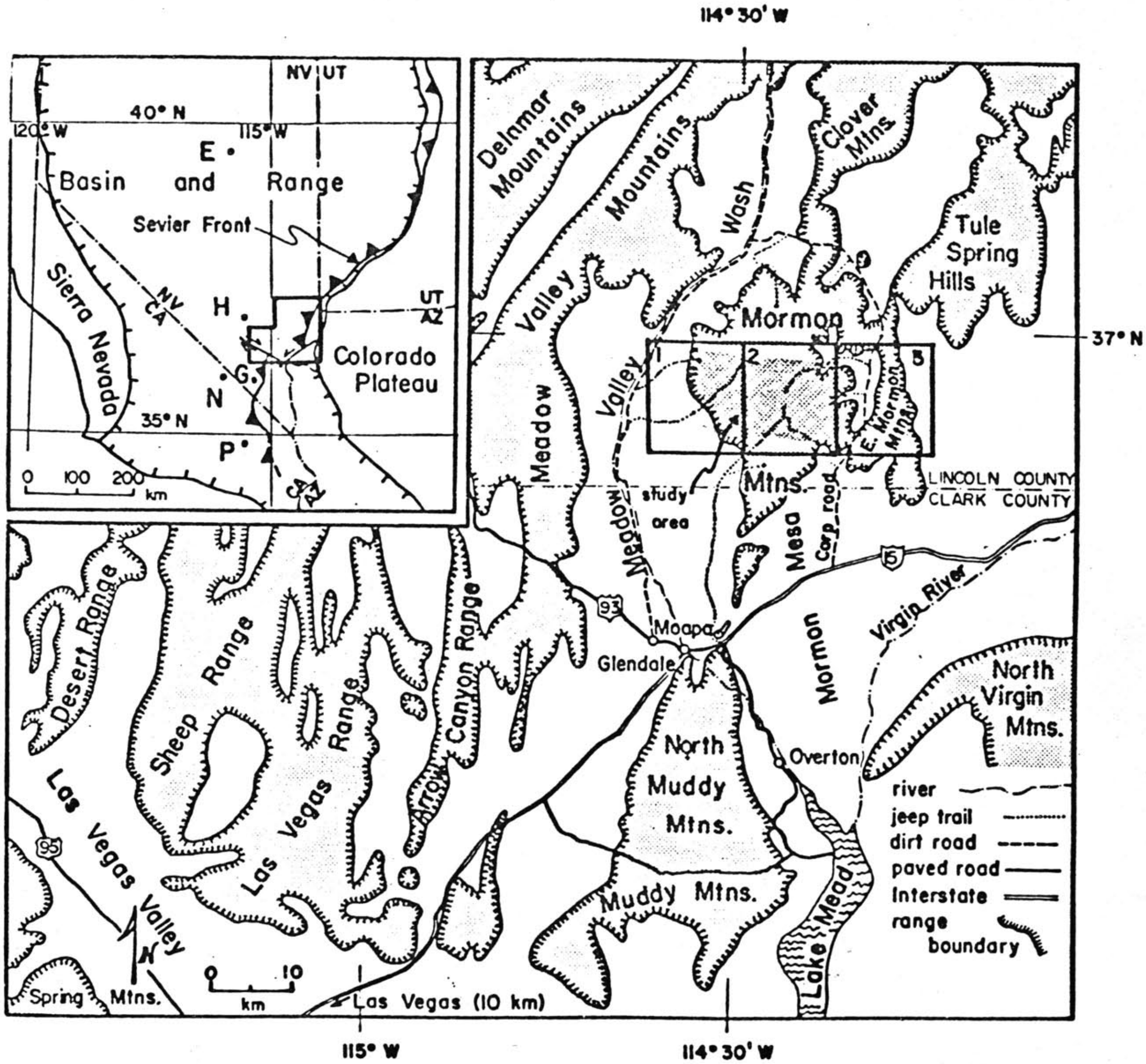
Figure 31. Tectonic map showing the structural relationship between the Mormon Mountains and Meadow Valley Mountains. Geology from Wernicke (this report), Wernicke (in prep.) Tschanz and Pampeyan (1970), and Ekren and others (1977). All mapping except from this report is reconnaissance.

Figure 32. Equal-area plot showing preferred west-southwest tilt direction of west tilted rocks and due east tilt direction for east tilted rocks. Most of the tilts are less than 30°, with over 90% less than 45°.

Figure 33. Tilt domains in the Mormon Peak allochthon, separated by tilt direction-parallel and inferred tilt direction-perpendicular boundaries. Note that these boundaries are apparently unrelated to any lower plate structures.

Figure 34. East pinchout of thick Cambrian sections in the Mormon Peak allochthon west of Mormon Peak requires at least 15 km displacement on the Mormon Peak detachment, if east-northeast transport is assumed.

Figure 35. First-order scheme for the accommodation of extensional strain in the Mormon Mountains area. Large, imbricate slabs are sheared past one another on low-angle normal faults while their shallow ends experience relatively penetrative extensional strain by a variety of normal fault types. In this diagram, not all of the displacement between deep, structurally simple portions of the slabs are accommodated by extension at the shallow ends.



114° 30' W
NV UT
40° N
120° W
115° W
E
Basin and Range
Sevier Front
Sierra Nevada
NV CA
H
UT AZ
Colorado Plateau
G
N
35° N
P
0 100 200 km
Delamar Mountains
Meadow Valley Mountains
Wash
Clover Mtns.
Tule Spring Hills
Mormon
37° N
1
2
3
Meadow Valley
E. Mormon Mtn.
Mtns.
study area
Carp road
LINCOLN COUNTY
CLARK COUNTY
Mesa
15
Virgin River
Desert Range
Sheep Range
Las Vegas Range
Arrow Canyon Range
93
Moapa
Glendale
Mormon
North Virgin Mtns.
North Muddy Mtns.
Overton
Las Vegas Valley
95
Muddy Mtns.
Lake Mead
river
jeep trail
dirt road
paved road
Interstate
range boundary
0 10 km
Spring Mtns.
Las Vegas (10 km)
115° W
114° 30' W

FIG. 1

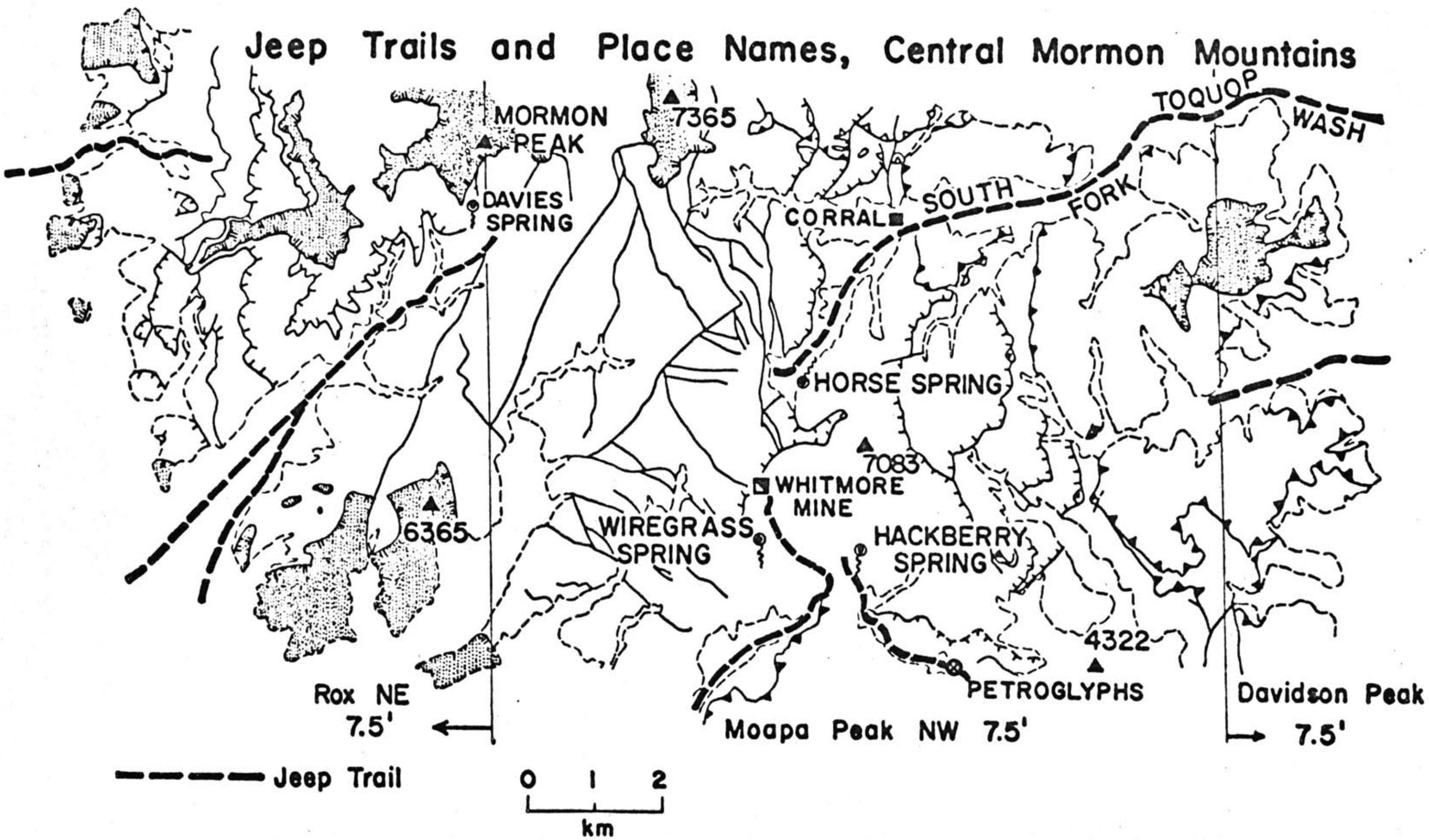

Jeep Trails and Place Names, Central Mormon Mountains
MORMON PEAK
7365
DAVIES SPRING
CORRAL
SOUTH
FORK
TOQUOP
WASH
HORSE SPRING
7083
WHITMORE MINE
6365
WIREGRASS SPRING
HACKBERRY SPRING
4322
PETROGLYPHS
Rox NE 7.5'
Moapa Peak NW 7.5'
Davidson Peak 7.5'
Jeep Trail
0 1 2
km

FIG. 2

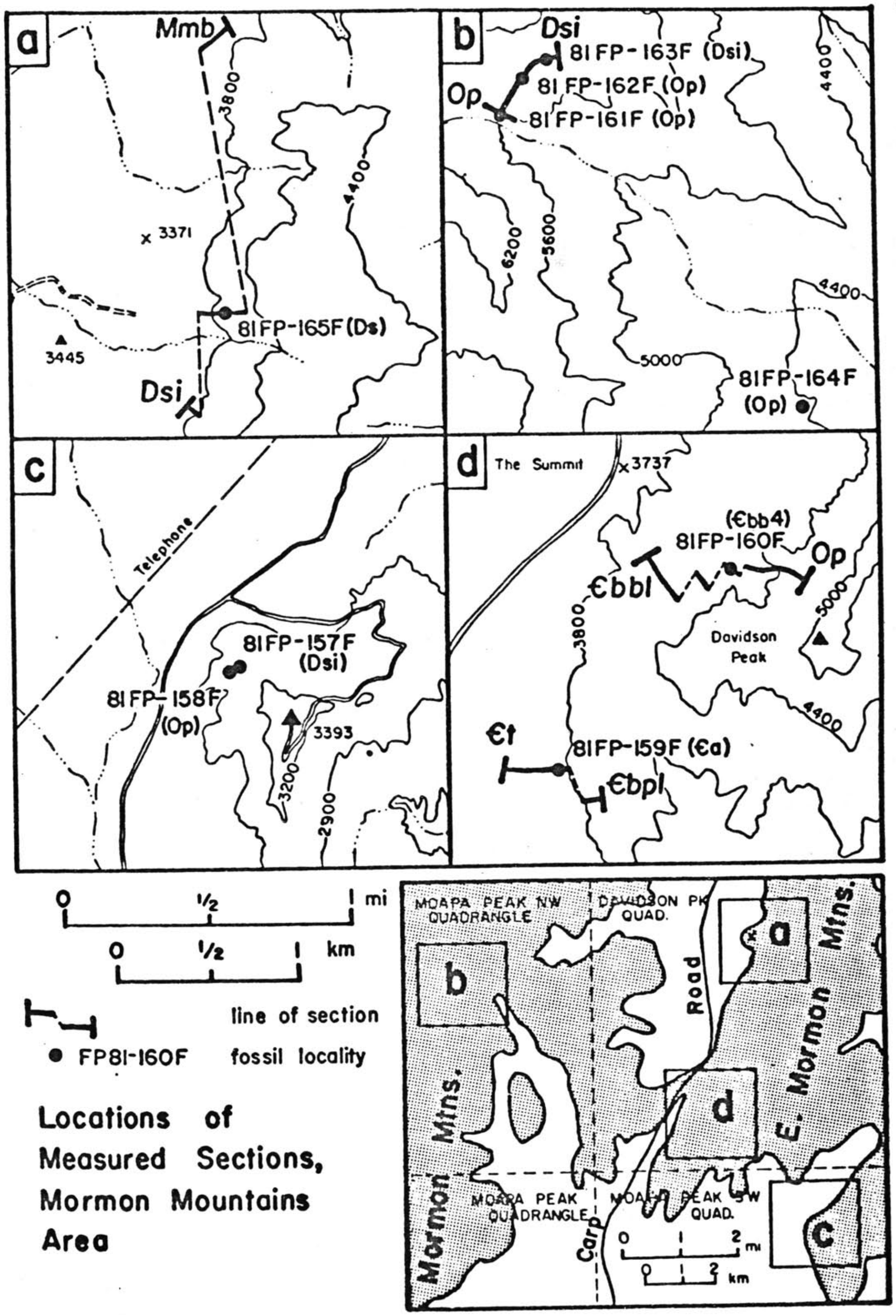

a
Mmb
3800
4400
x 3371
3445
81FP-165F(Ds)
Dsi
b
Dsi
81 FP-163F (Dsi)
81 FP-162F (Op)
Op
81 FP-161F (Op)
4400
6200
5600
4400
5000
81FP-164F
(Op)
c
Telephone
81FP-157F
(Dsi)
81 FP-158F
(Op)
3393
3200
2900
d
The Summit
x 3737
(€bb4)
81FP-160F
Op
€bbl
3800
5000
Davidson
Peak
4400
€t
81FP-159F (€a)
€bpl
0
1/2
1 mi
0
1/2
1 km
line of section
FP81-160F
fossil locality
Locations of
Measured Sections,
Mormon Mountains
Area
MOAPA PEAK NW
QUADRANGLE
DAVIDSON PK
QUAD.
Road
E. Mormon Mtns.
Mormon Mtns.
MOAPA PEAK
QUADRANGLE
MOAPA PEAK SW
QUAD.
Carp
0
2 mi
0
2 km

FIG. 3

Fig. 5

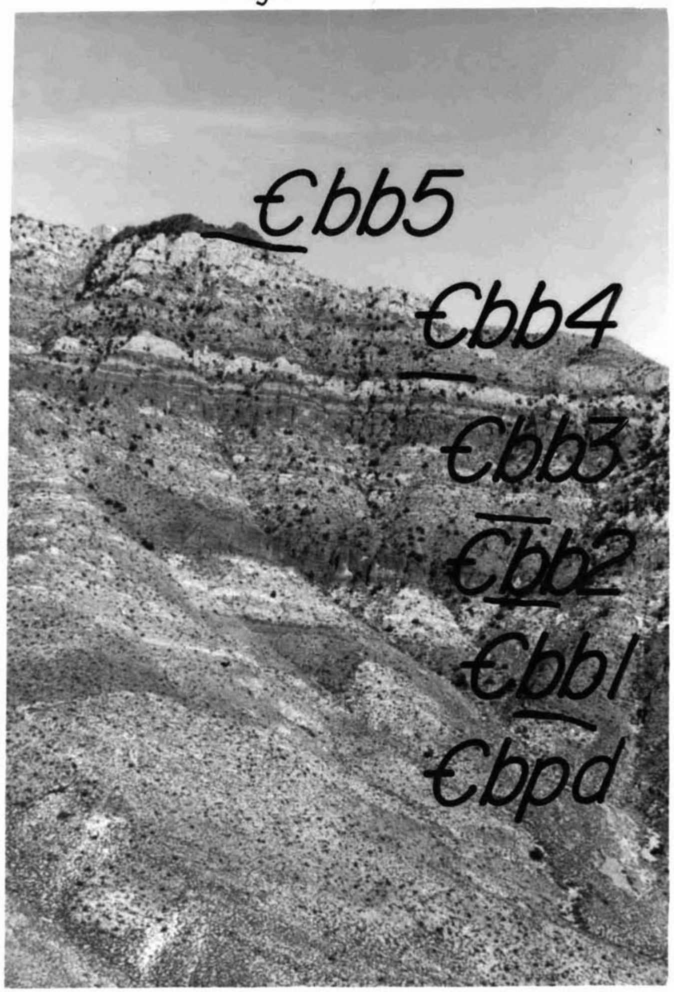


Fig. 4

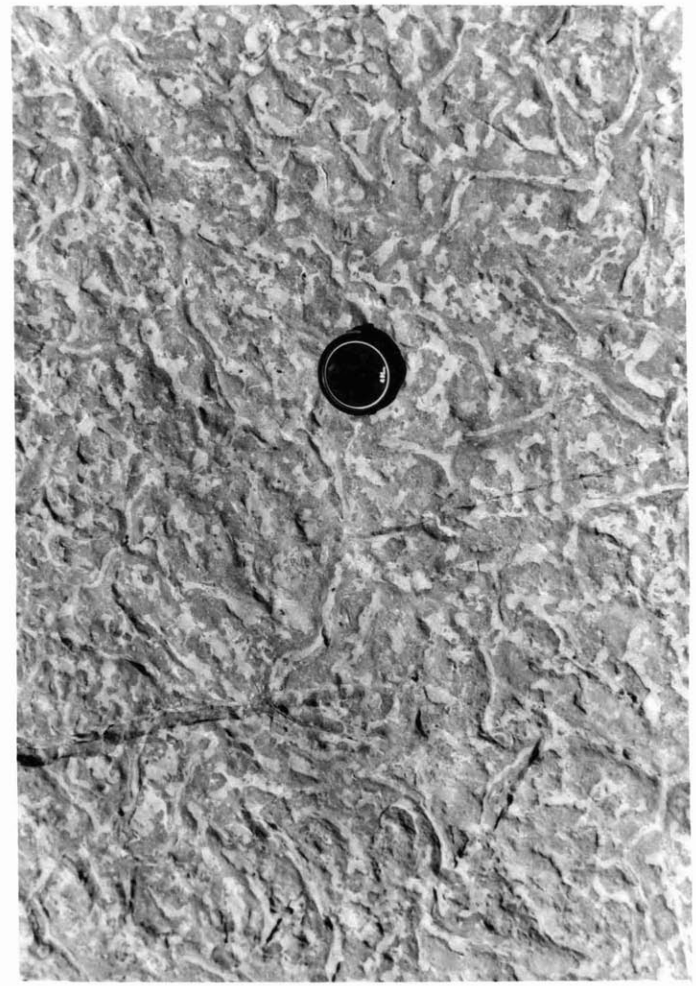
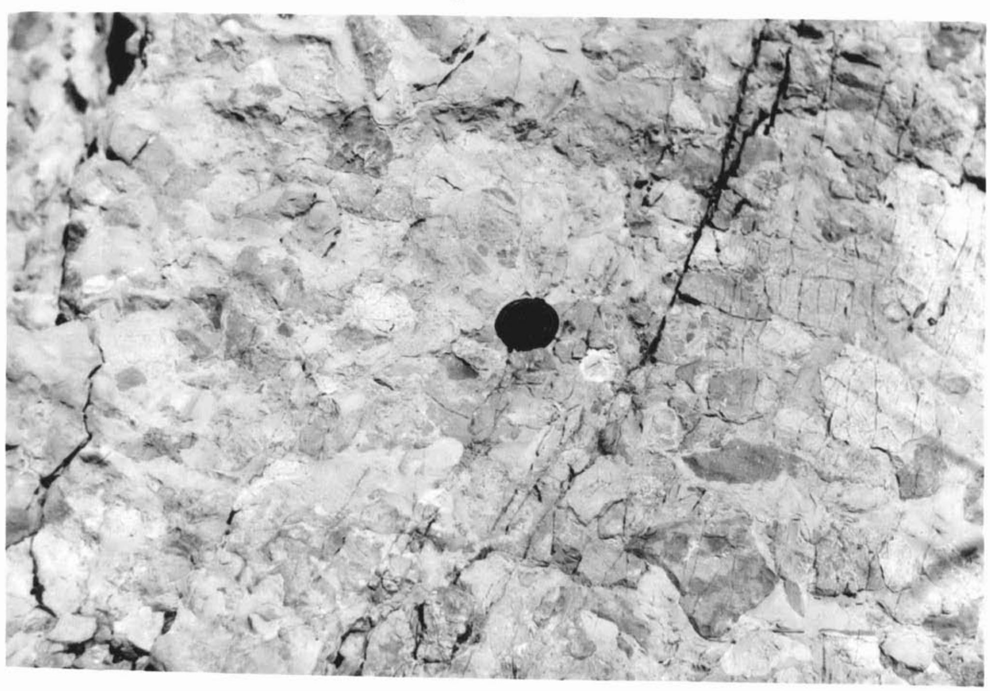


Fig. 6

Fig. 7



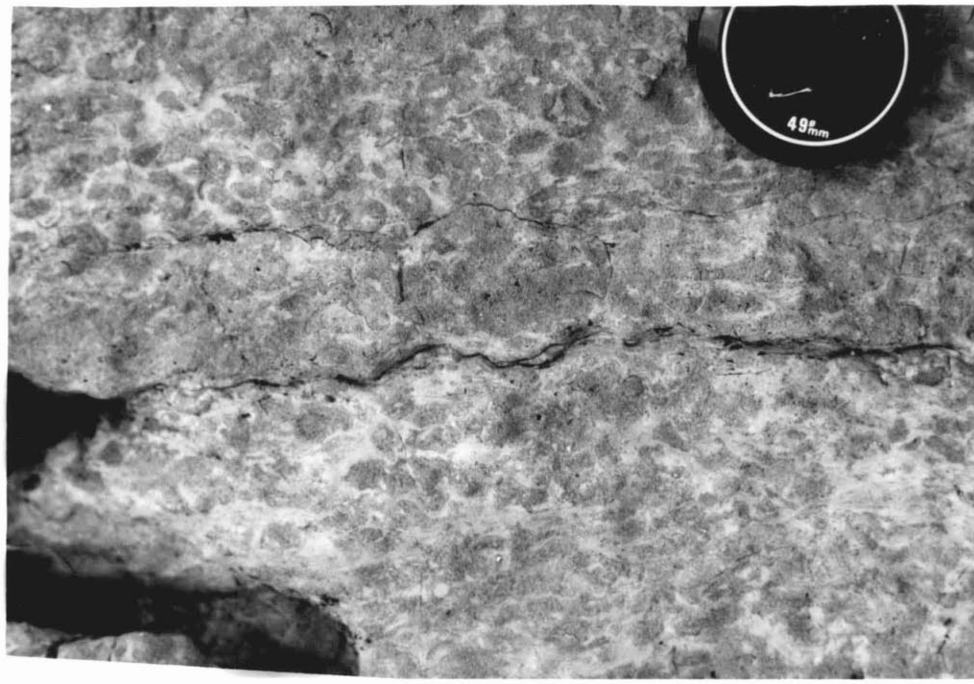


Fig. 8

Fig. 9

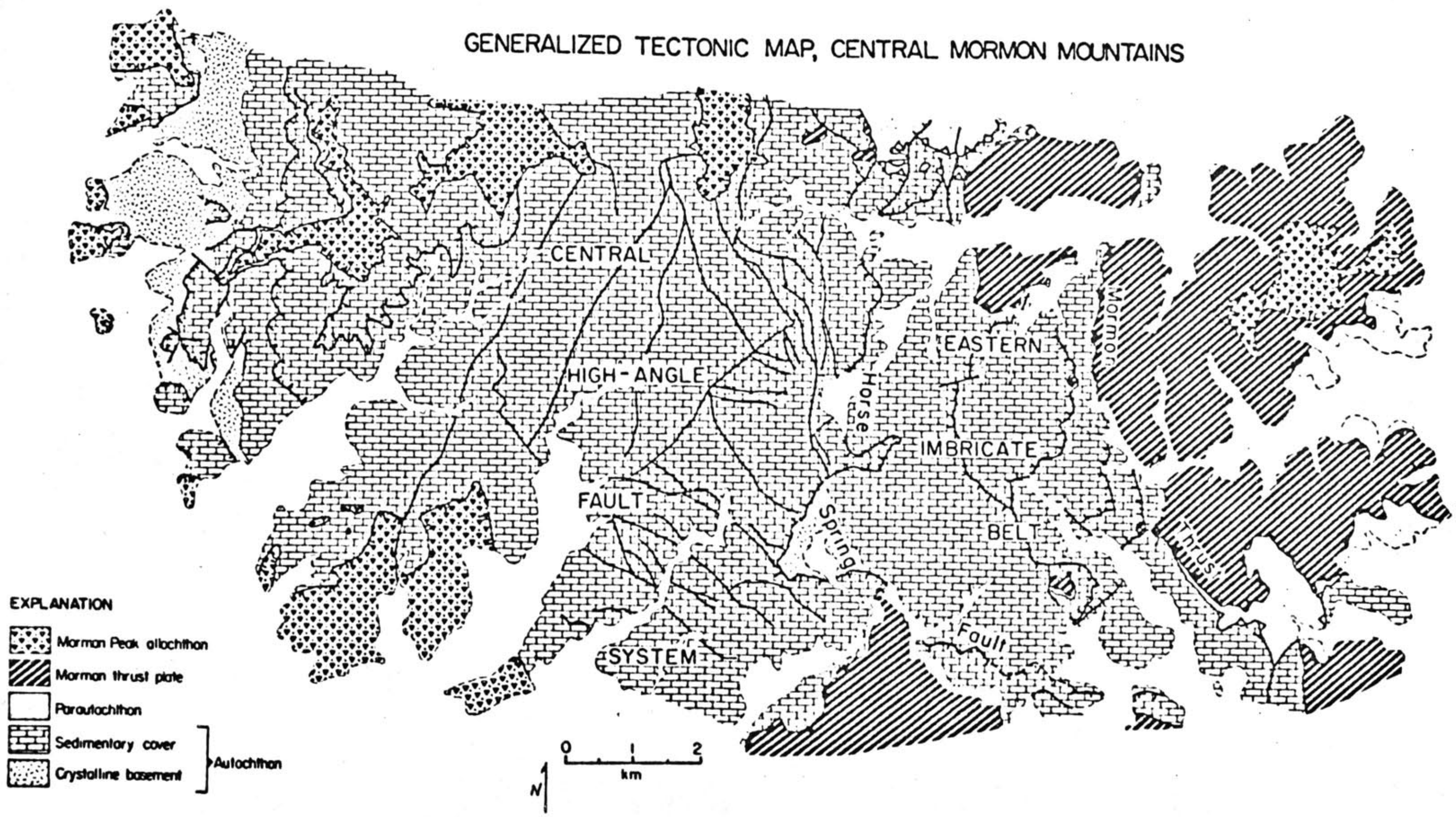

GENERALIZED TECTONIC MAP, CENTRAL MORMON MOUNTAINS
CENTRAL
HIGH-ANGLE
FAULT
SYSTEM
EASTERN
IMBRICATE
BELT
Horse
Spring
Fault
Mormon
Thrust
EXPLANATION
Mormon Peak allochthon
Mormon thrust plate
Parautochthon
Sedimentary cover
Crystalline basement
Autochthon
0
1
2
km
N

FIG. 10

FOOTWALL GEOLOGY, MORMON THRUST SYSTEM

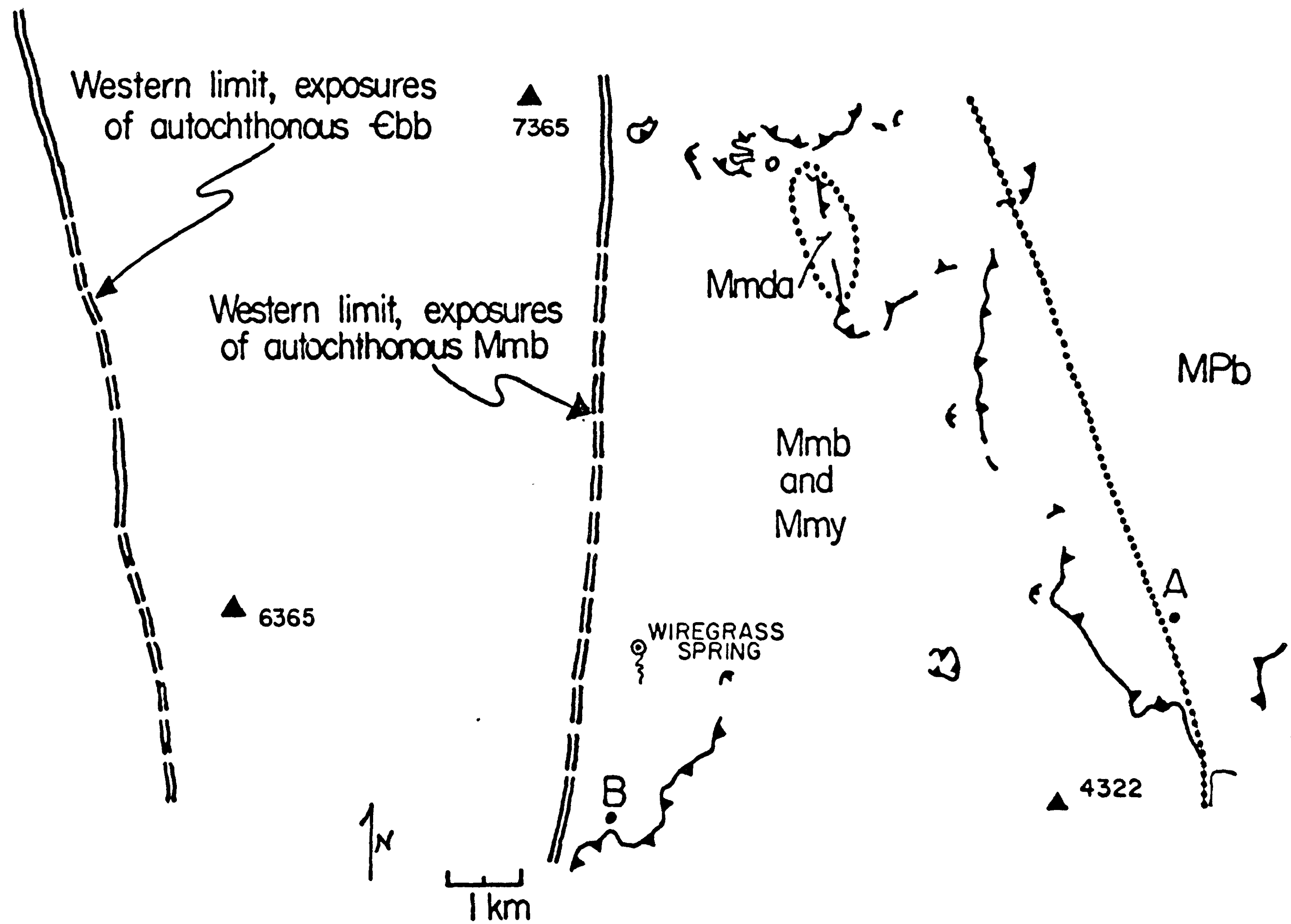


FIG. 11

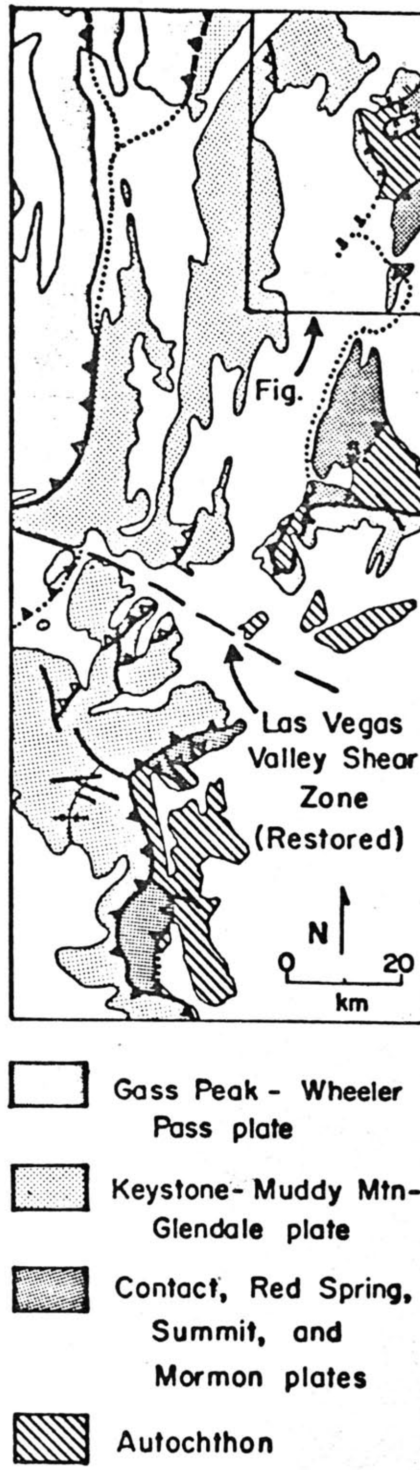
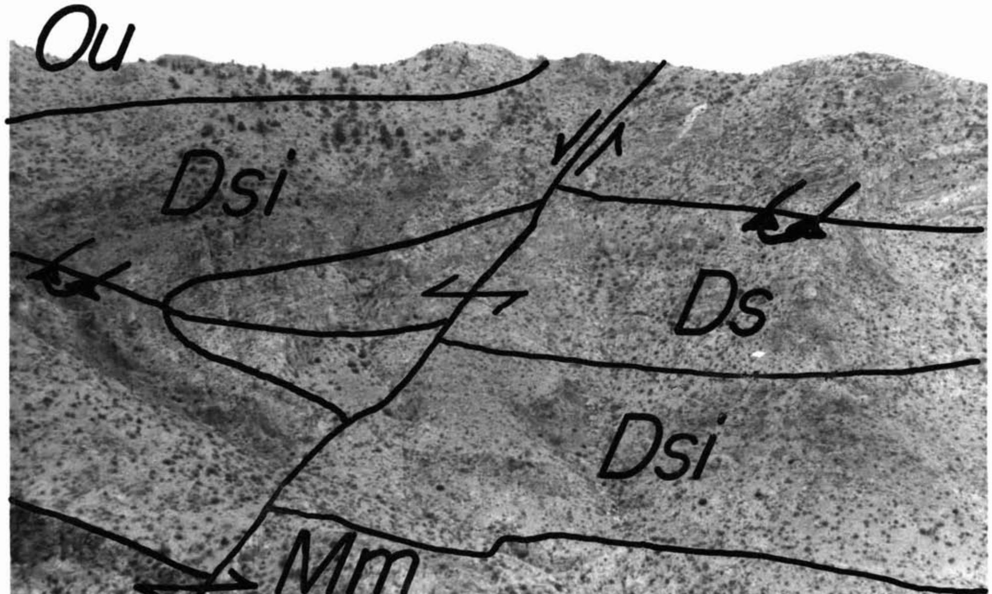
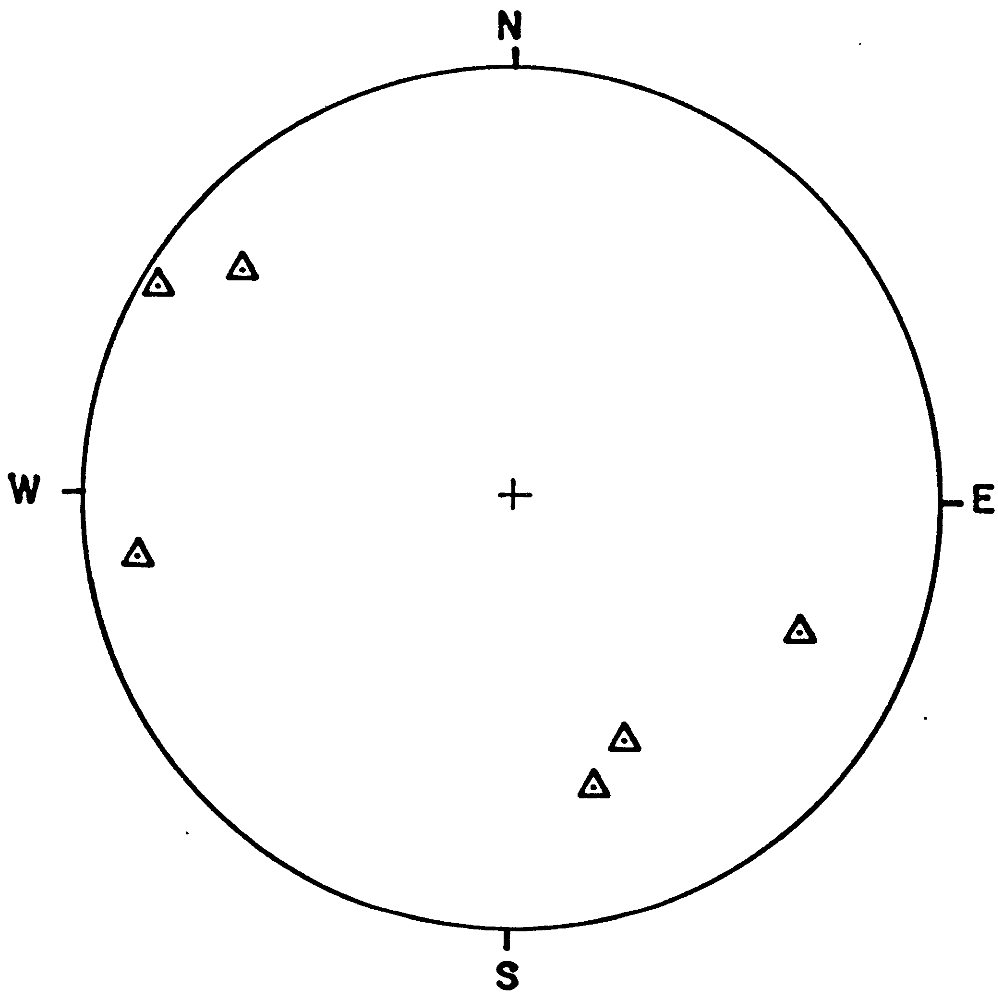


Fig.
Las Vegas
Valley Shear
Zone
(Restored)
N
0
20
km
Gass Peak - Wheeler Pass plate
Keystone- Muddy Mtn- Glendale plate
Contact, Red Spring, Summit, and Mormon plates
Autochthon

FIG. 12

Fig. 13





Fold Axes in Parautochthonous Slices

FIG. 14

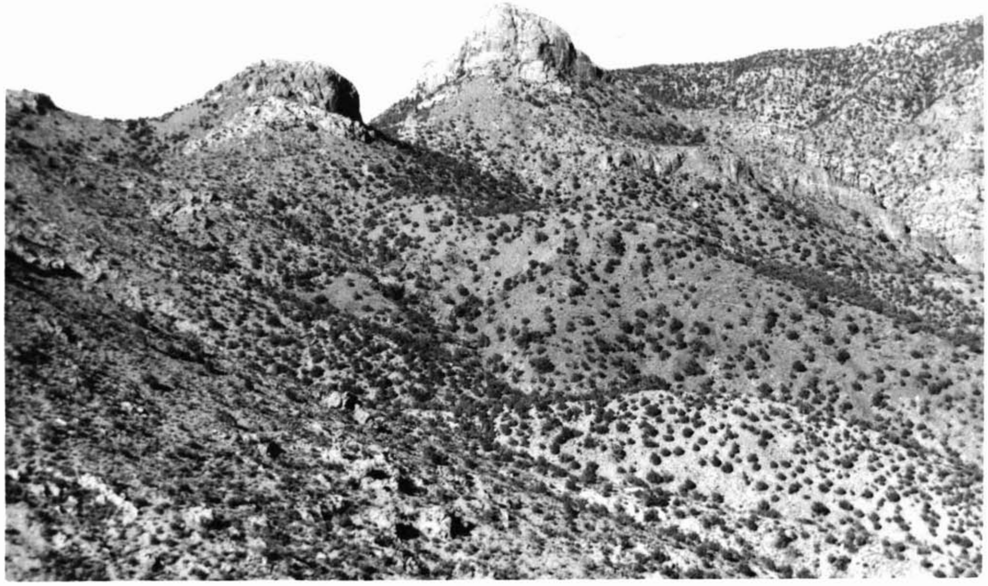


Fig. 15

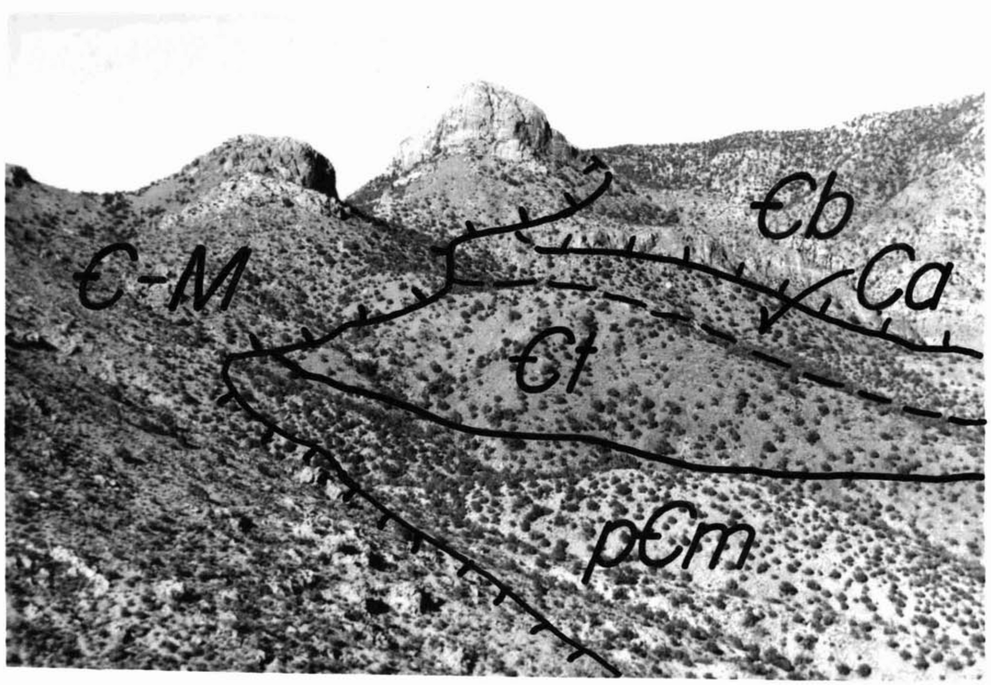
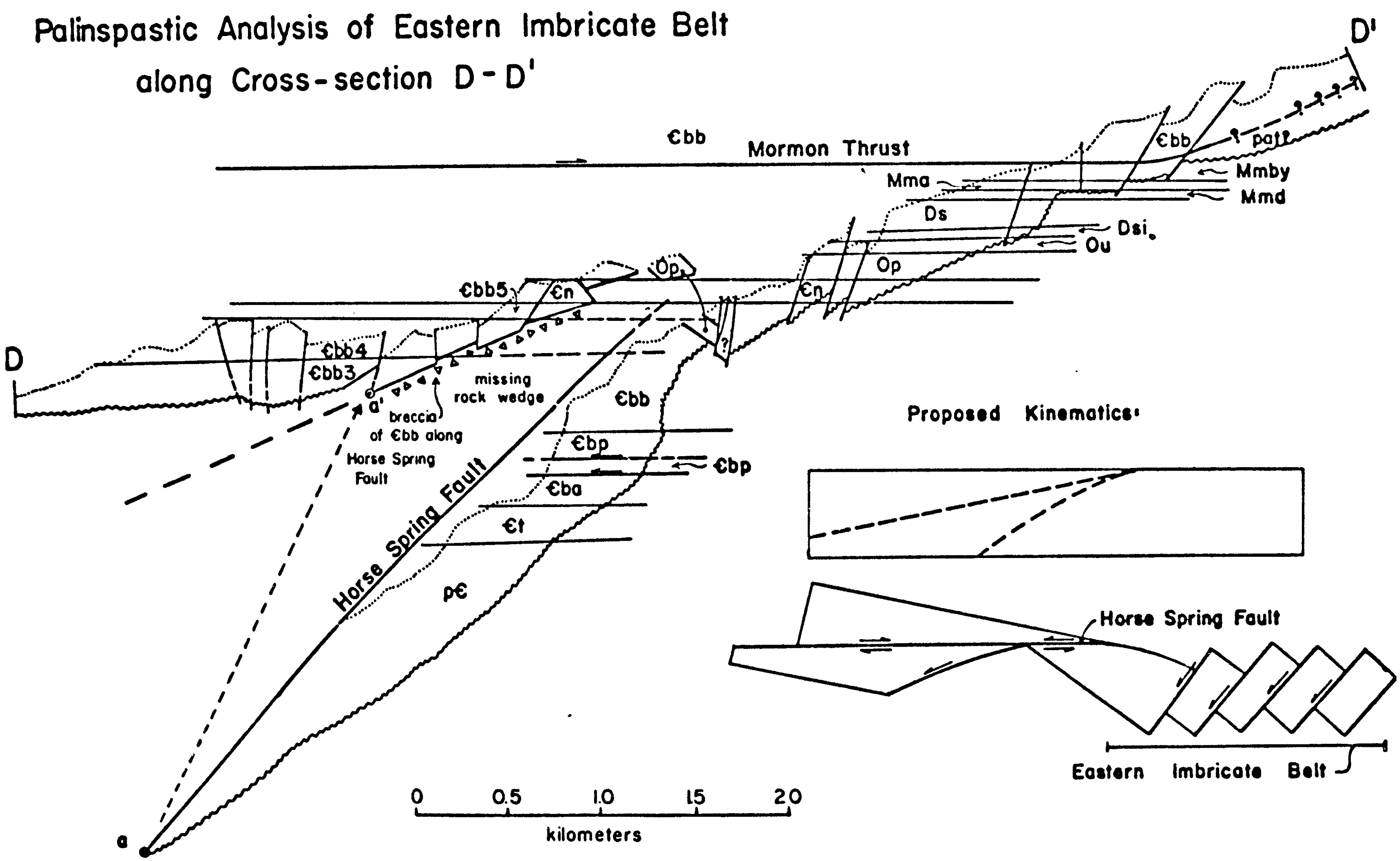


FIG. 16



Palinspastic Analysis of Eastern Imbricate Belt
along Cross-section D - D'
D
D'
Єbb
Mormon Thrust
Єbb
pal?
Mmby
Mmd
Mma
Ds
Dsi
Ou
Op
Єn
Op
Єn
Єbb5
Єbb4
Єbb3
missing
rock wedge
a'
breccia
of Єbb along
Horse Spring
Fault
Horse Spring Fault
Єbb
Єbp
Єbp
Єba
Єt
pЄ
a
0
0.5
1.0
1.5
2.0
kilometers
Proposed Kinematics:
Horse Spring Fault
Eastern Imbricate Belt

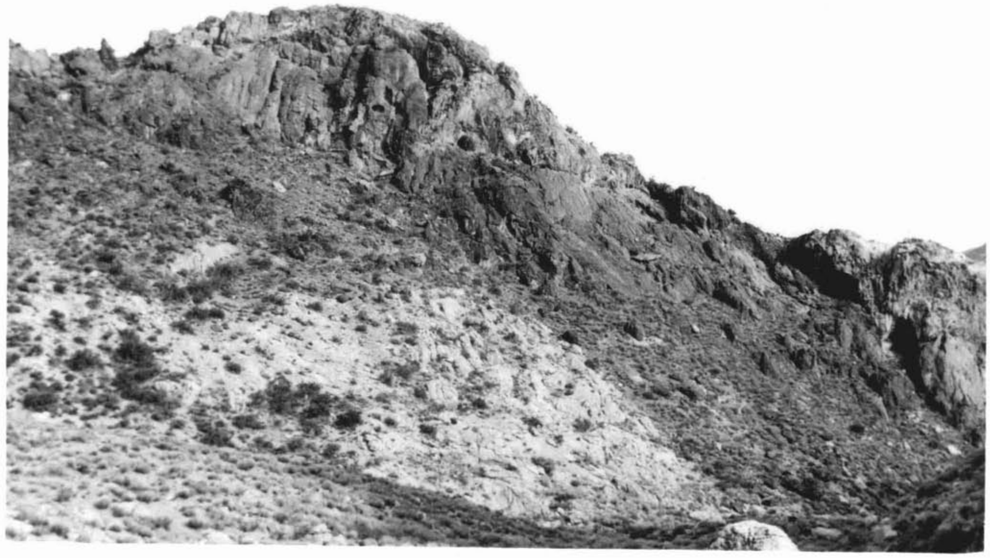
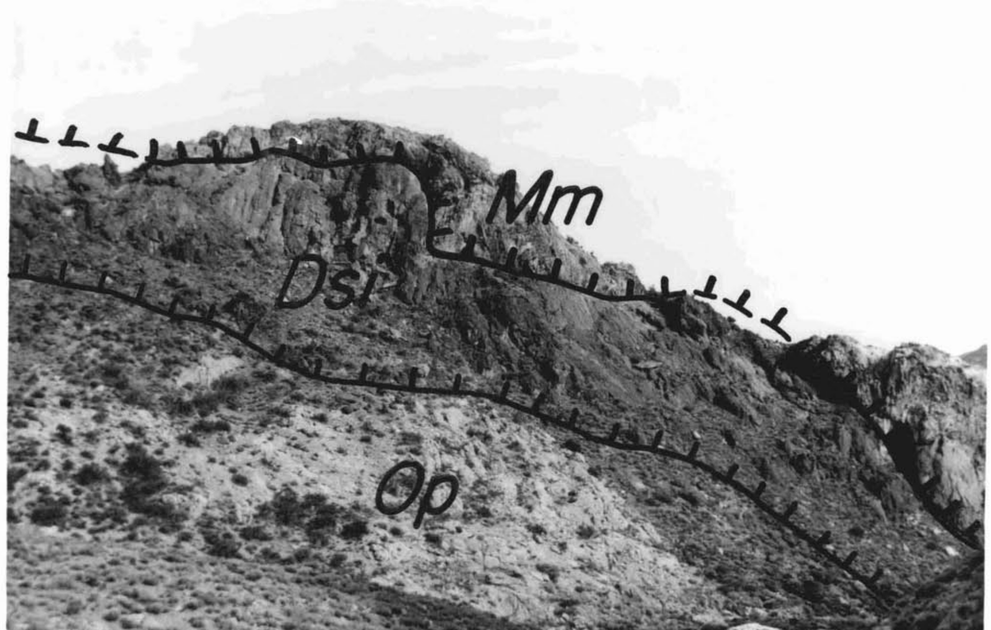


Fig. 17



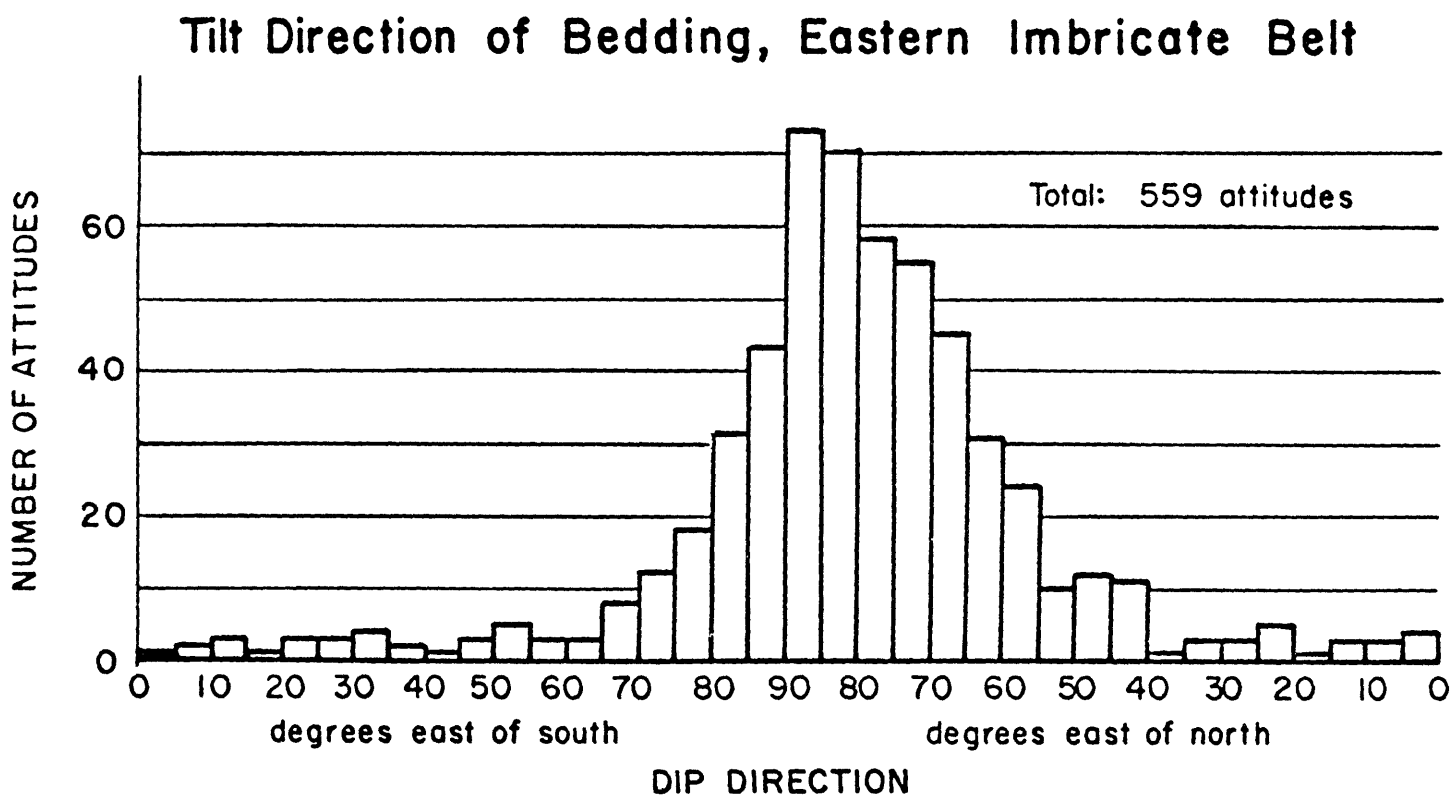
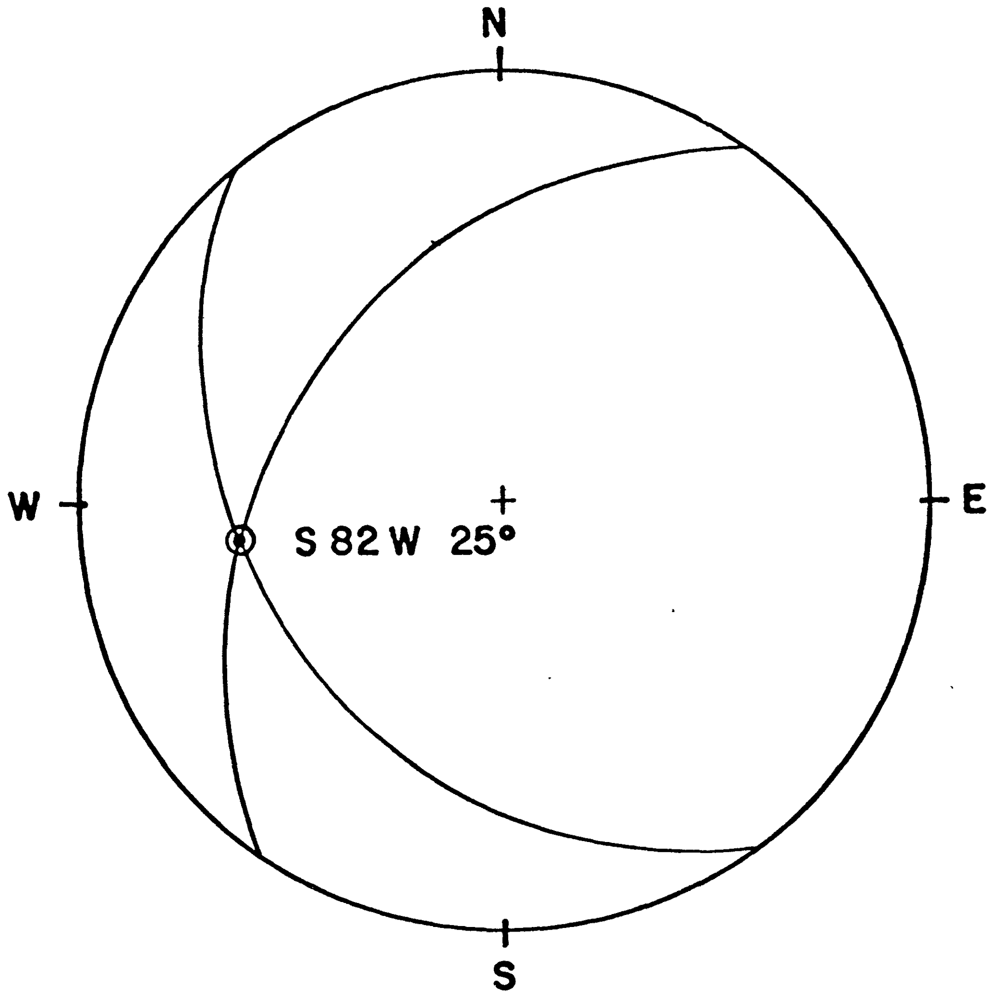

Tilt Direction of Bedding, Eastern Imbricate Belt
Total: 559 attitudes
NUMBER OF ATTITUDES
0
20
40
60
0 10 20 30 40 50 60 70 80 90 80 70 60 50 40 30 20 10 0
degrees east of south
degrees east of north
DIP DIRECTION

FIG. 18



Slip-line of Intersecting Fault Planes

FIG. 19

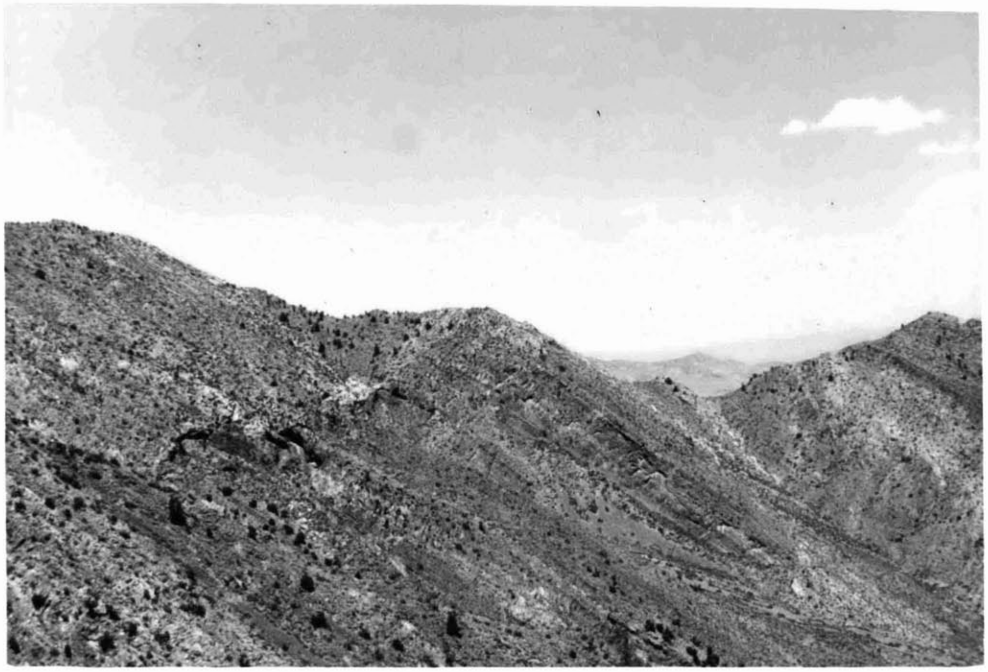
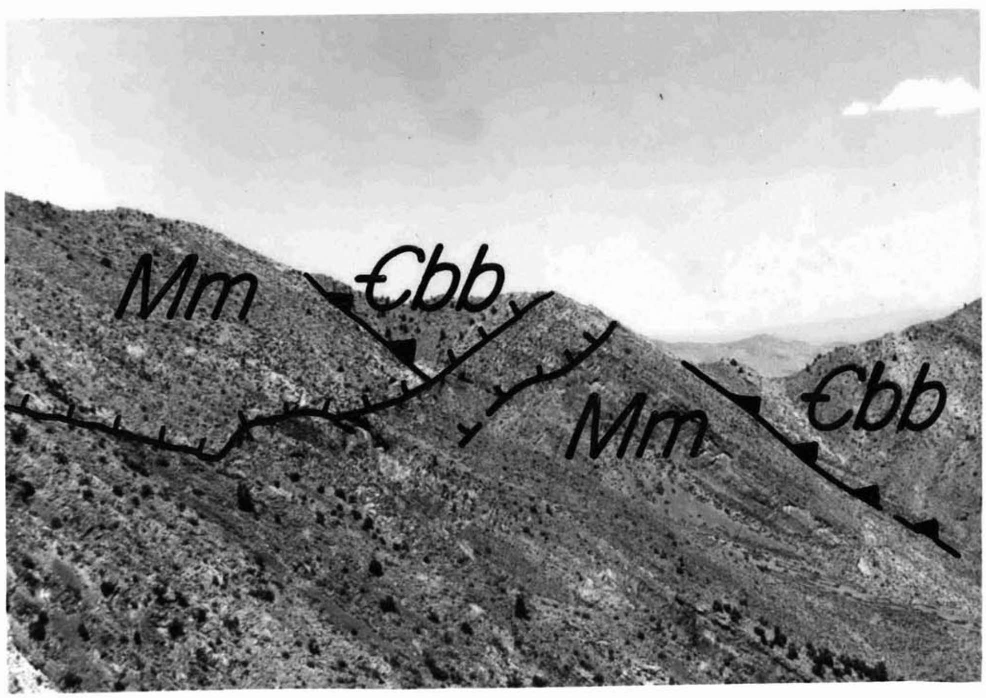


Fig. 20



Orientation and Sense of Throw, Central High-Angle Fault Terrane

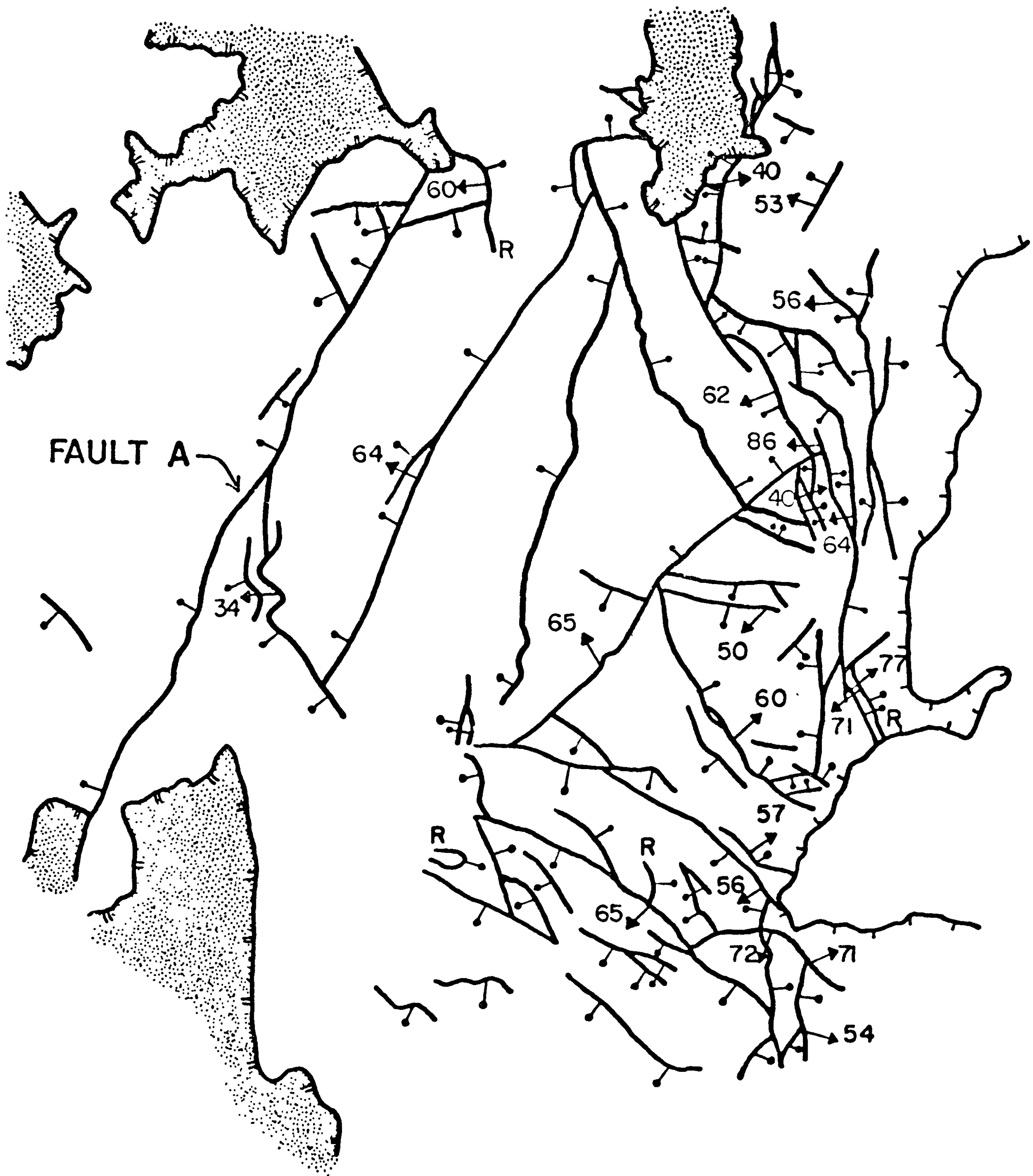


FIG. 21

Strike of Faults

Central High-Angle Fault System

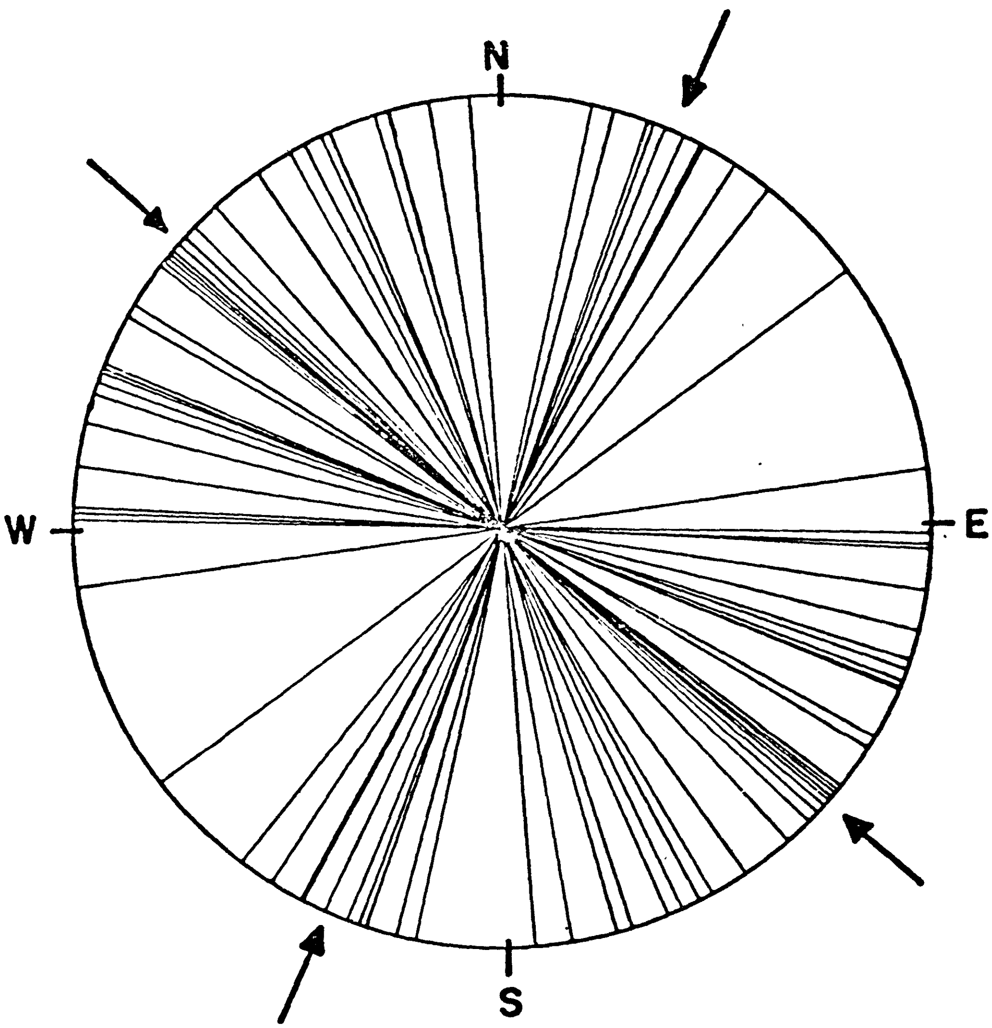


FIG. 22

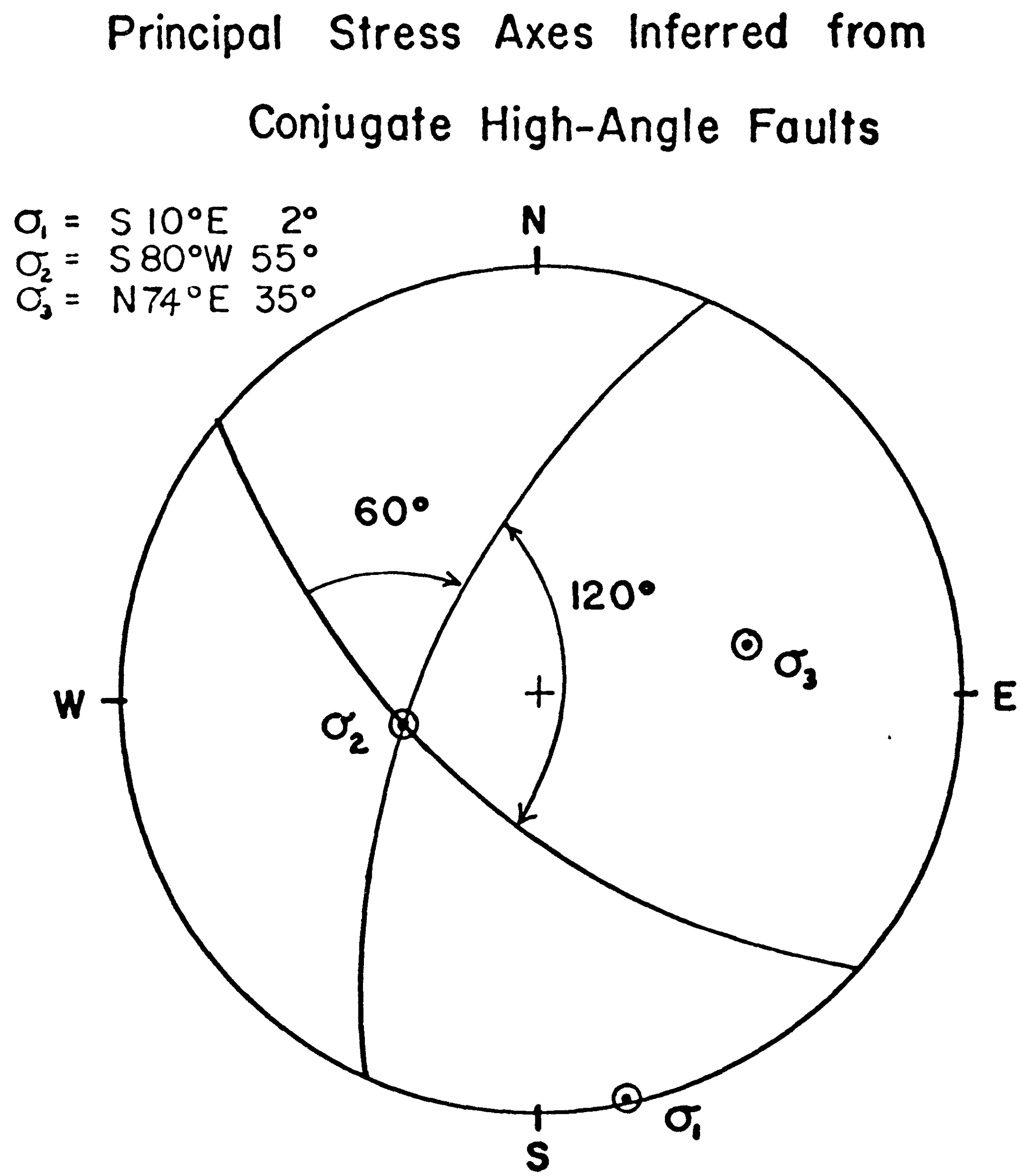

Principal Stress Axes Inferred from
Conjugate High-Angle Faults
σ1 = S 10°E 2°
σ2 = S 80°W 55°
σ3 = N74°E 35°
N
W
E
S
60°
120°
σ2
σ3
σ1

FIG. 23

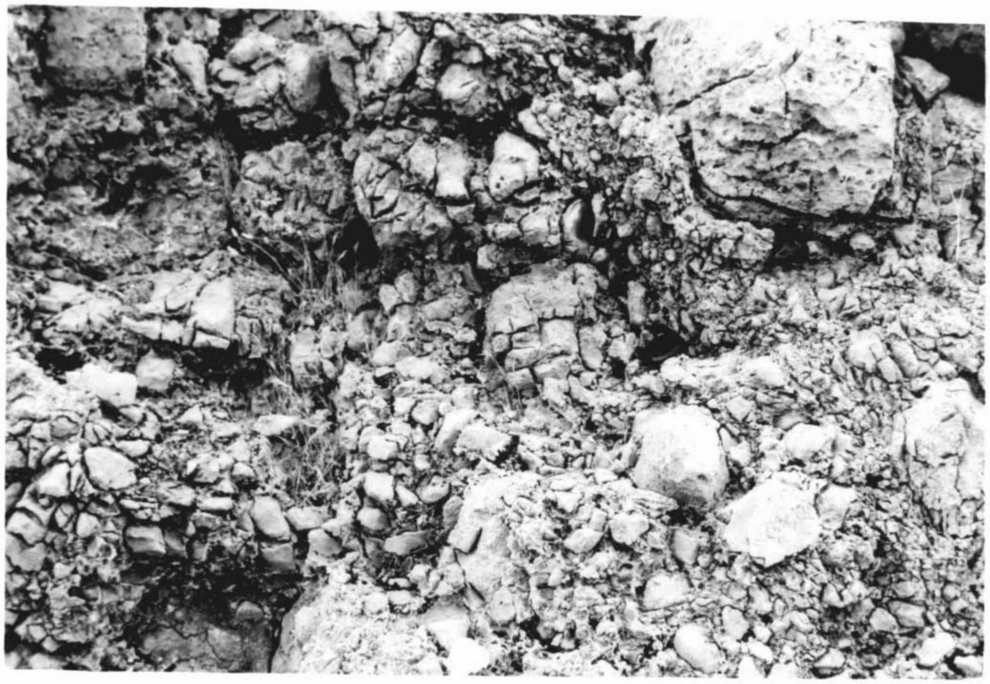
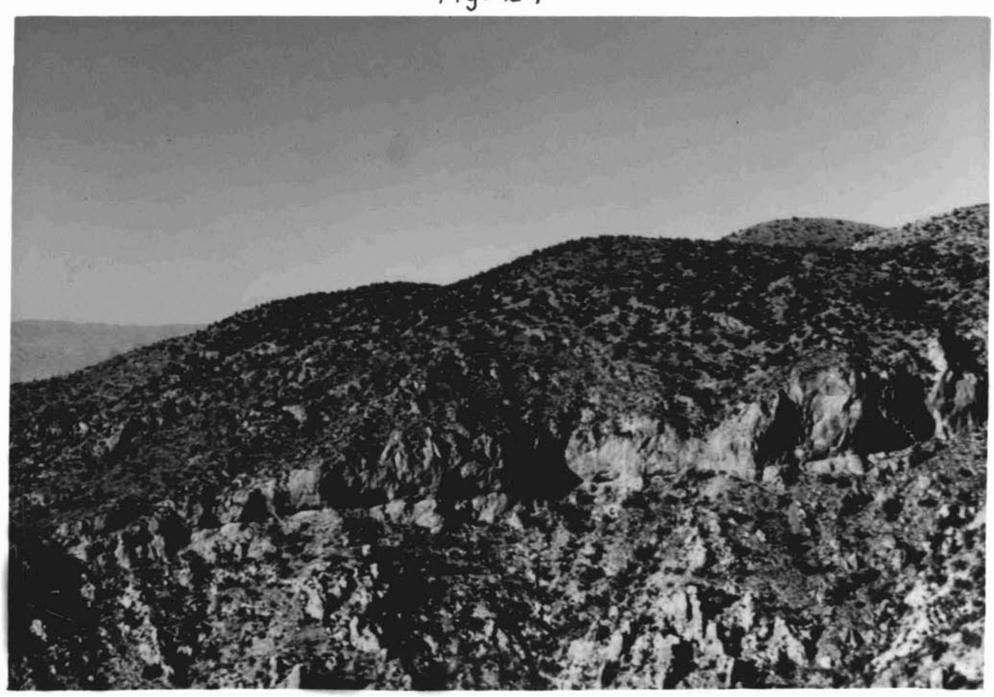
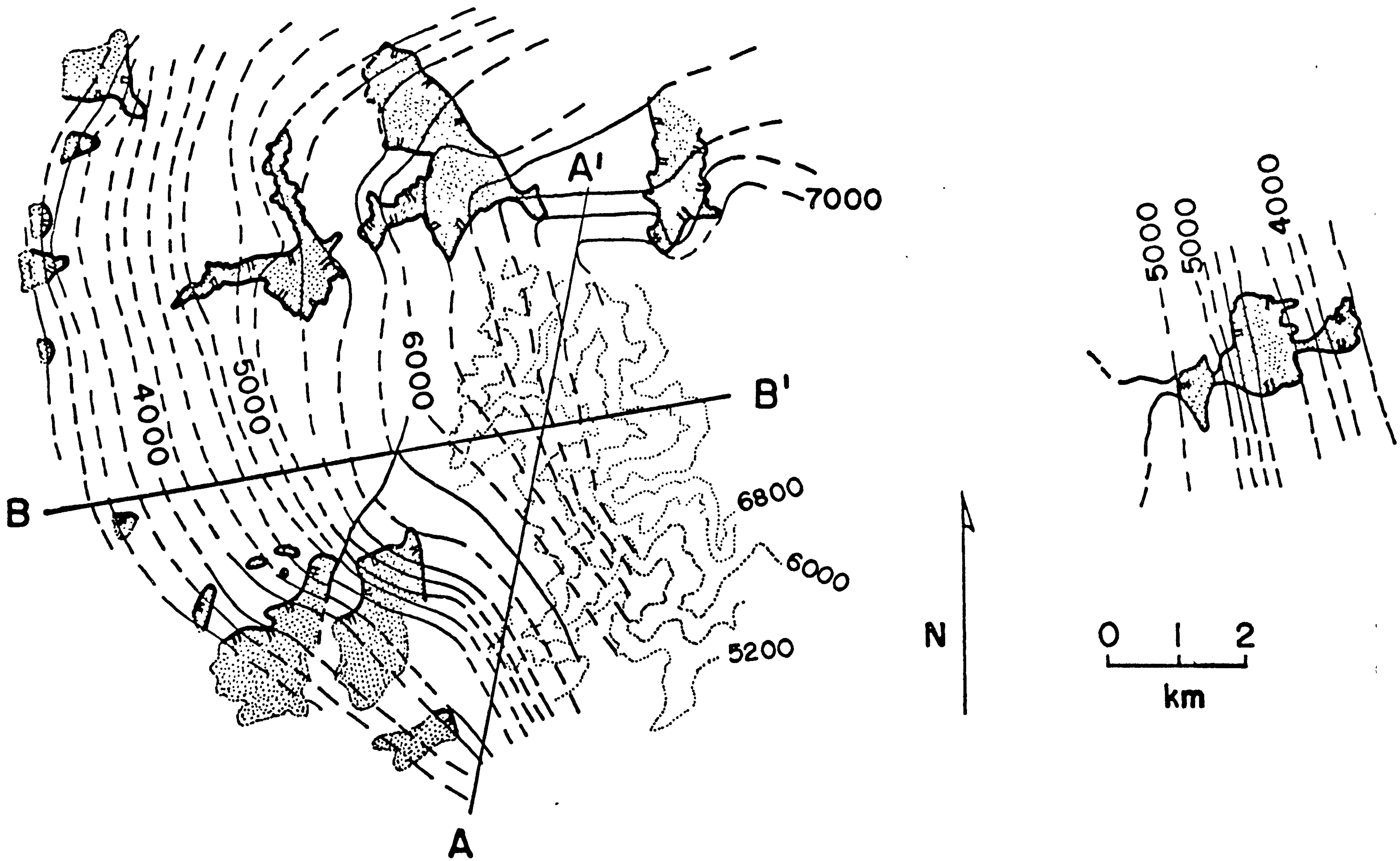


Fig. 29

Fig. 24

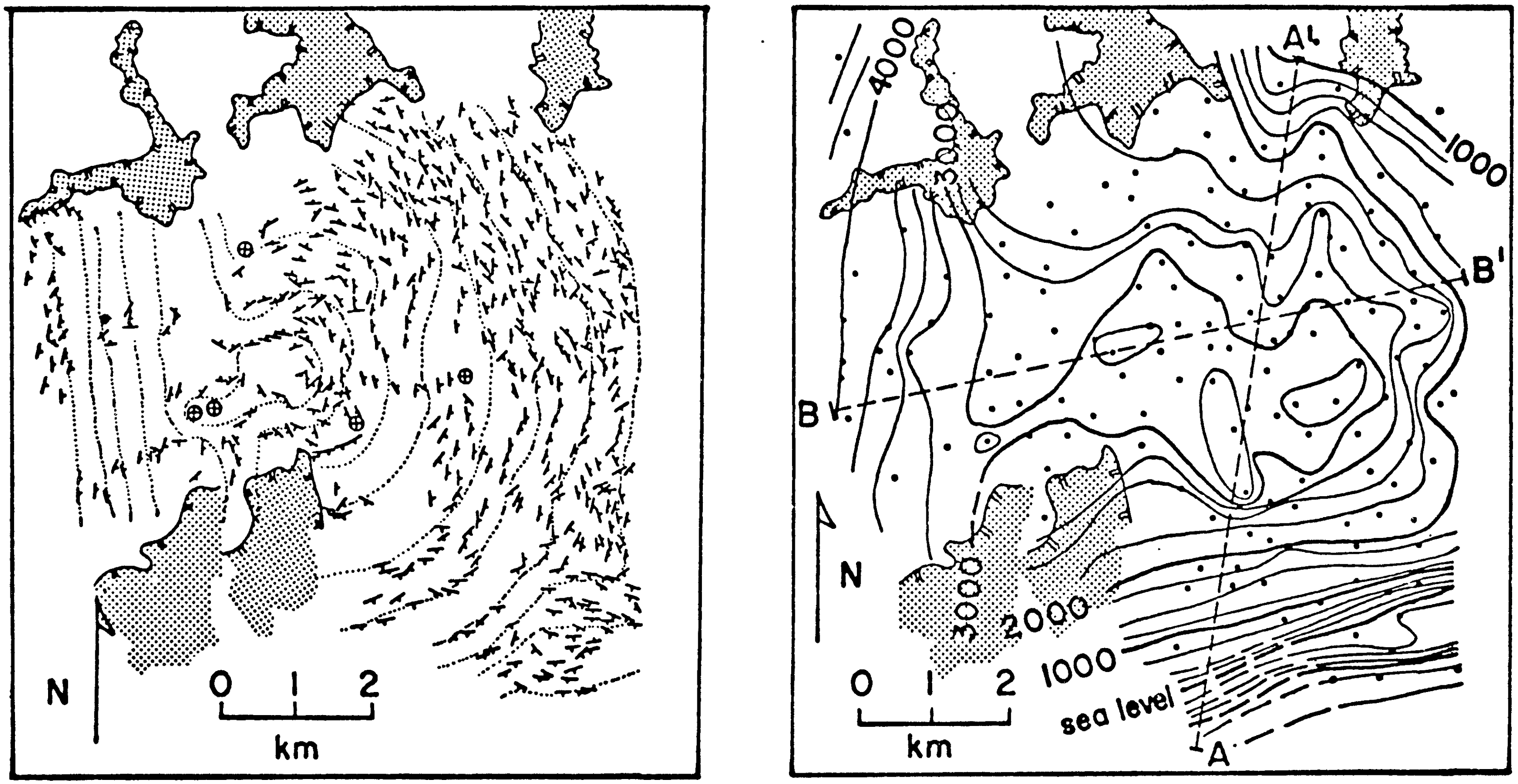




Elevation of Mormon Peak Detachment (in feet)

FIG. 25

FIG. 26



a) Form of Domal Structure in Autochthon

b) Elevation of Base of Tapeats Sandstone (feet)

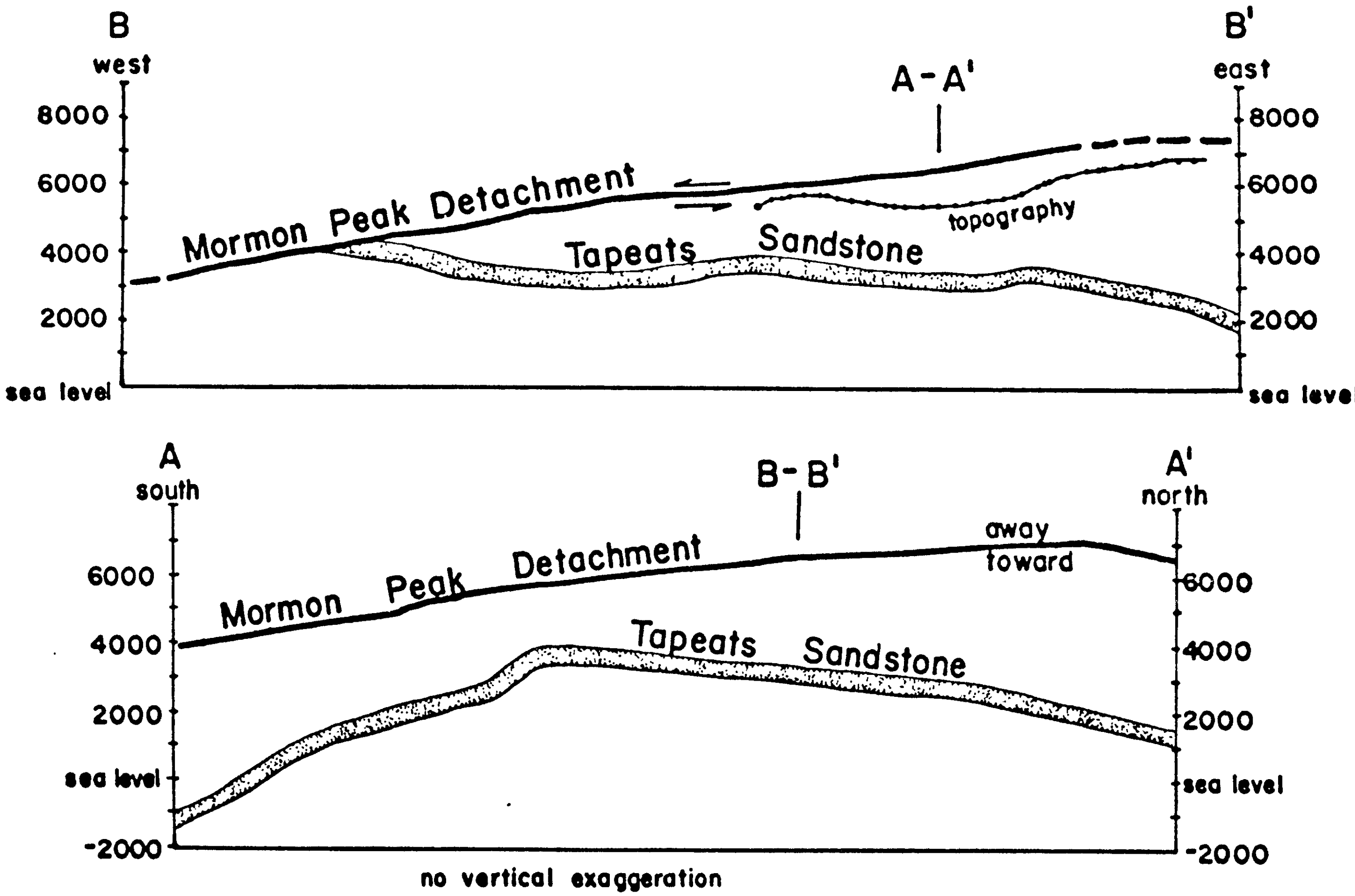
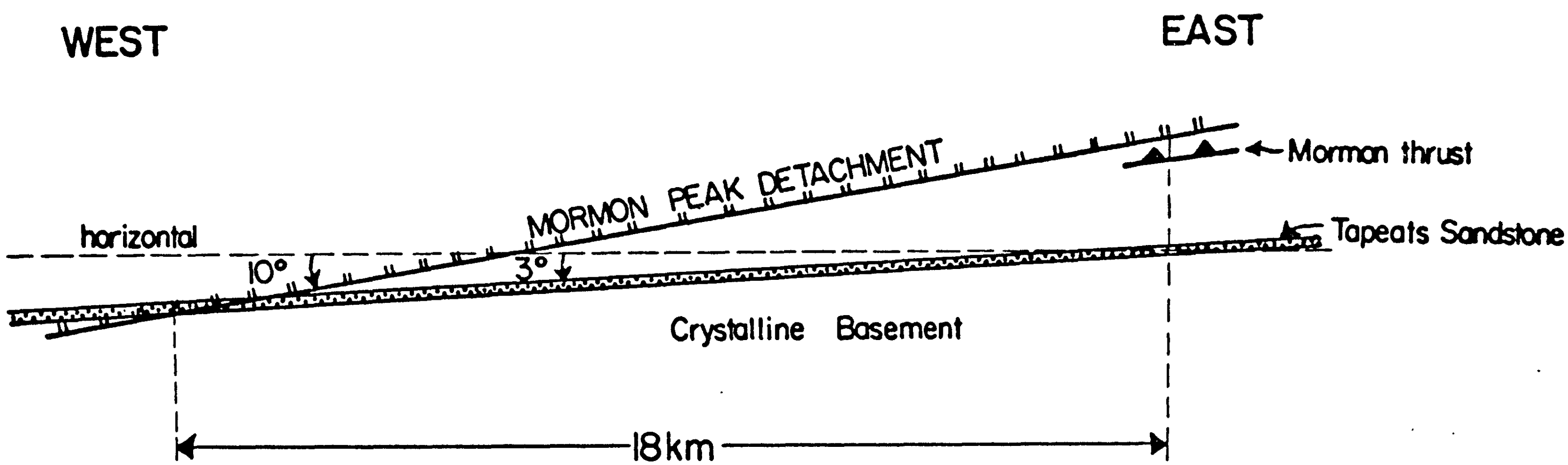

B
west
B'
east
A - A'
8000
6000
4000
2000
sea level
Mormon Peak Detachment
topography
Tapeats Sandstone
A
south
B - B'
A'
north
away
toward
6000
4000
2000
sea level
-2000
Mormon Peak Detachment
Tapeats Sandstone
no vertical exaggeration

FIG. 27

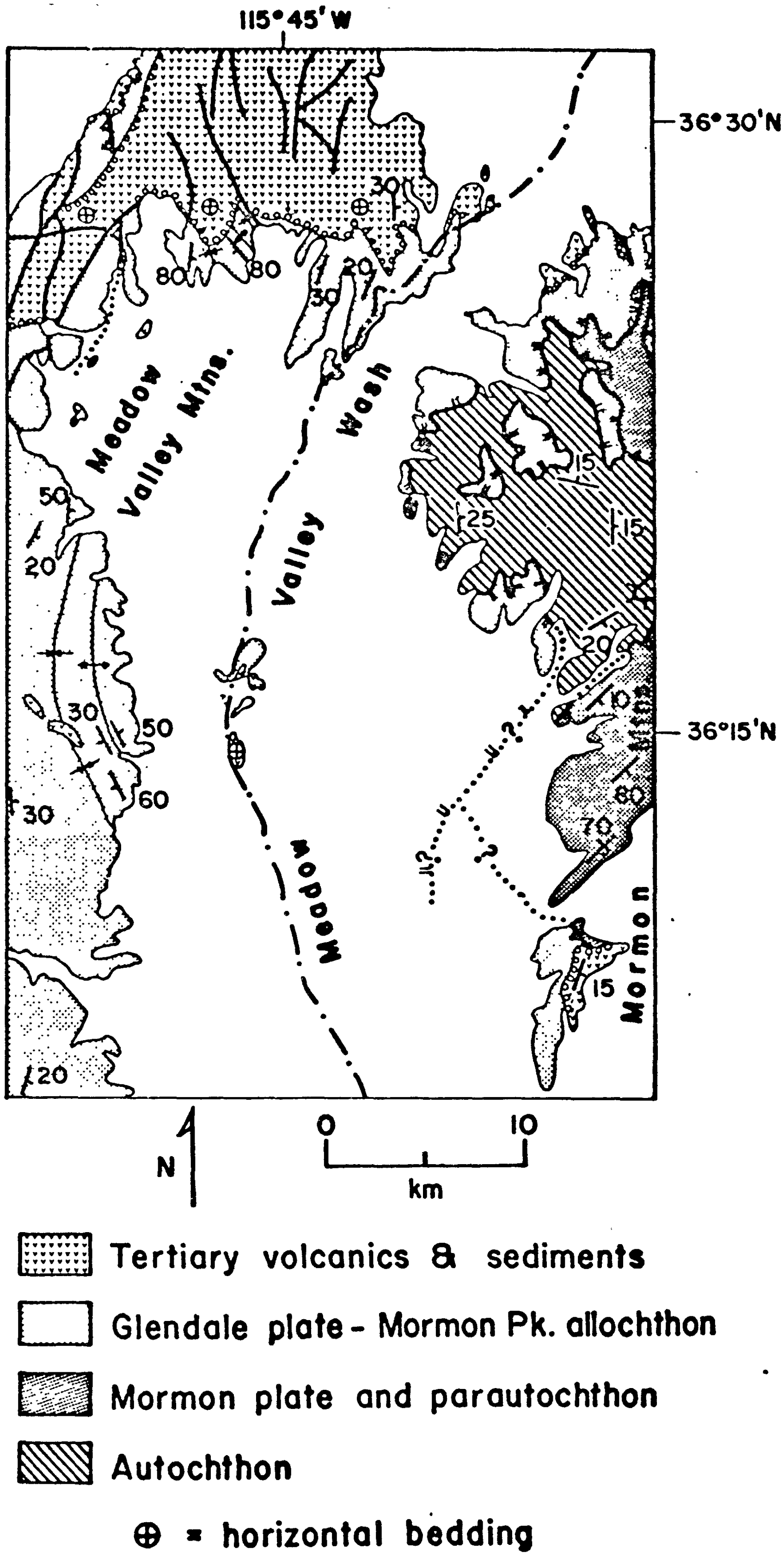
FIG. 28

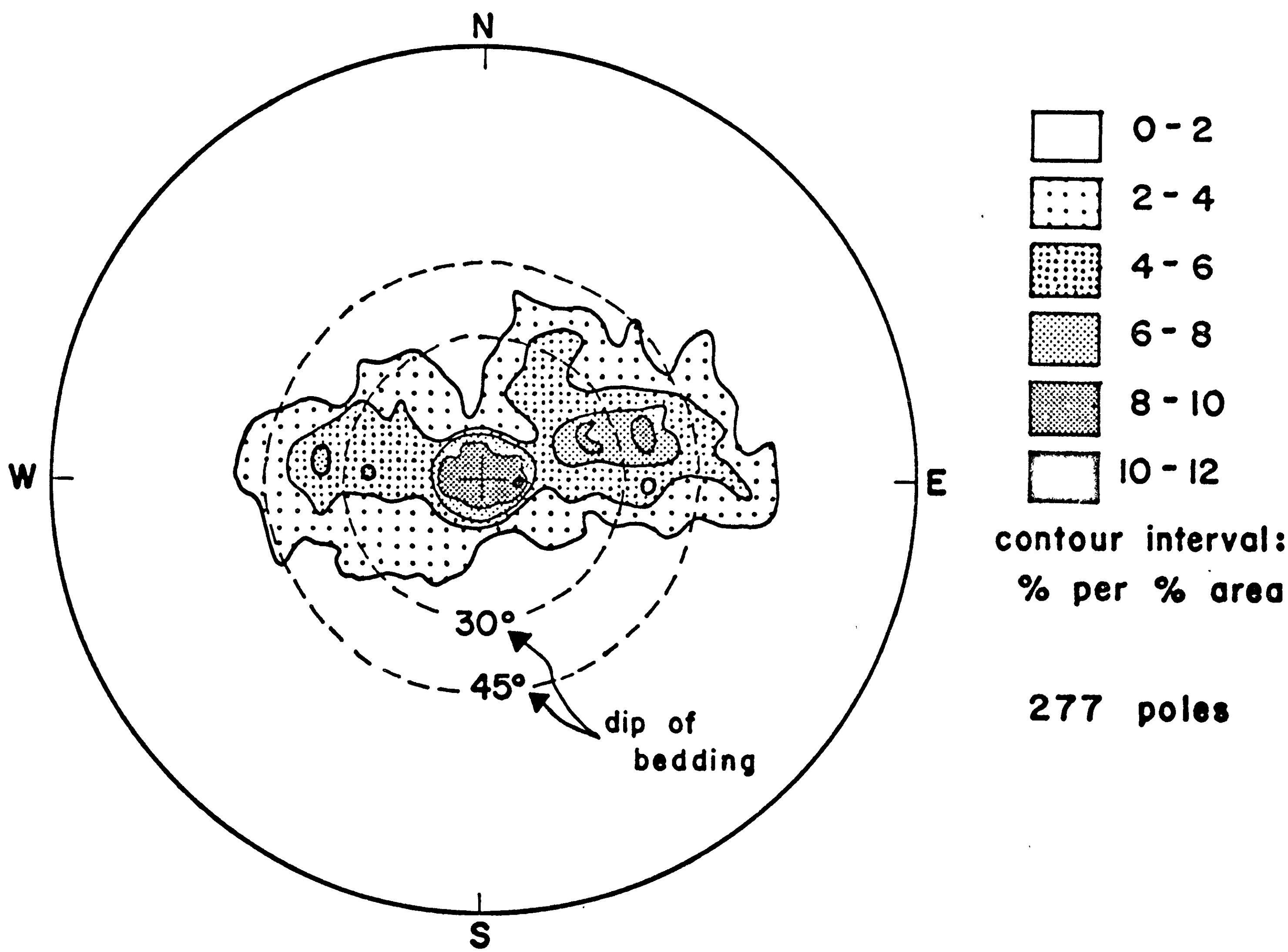


WEST
EAST
MORMON PEAK DETACHMENT
Mormon thrust
horizontal
Tapeats Sandstone
10°
3°
Crystalline Basement
18km

Fig. 30

FIG.31


115°45'W
36°30'N
36°15'N
Meadow Valley Mtns.
Valley Wash
Meadow
Mormon
Mtns.
0
10
km
N
Tertiary volcanics & sediments
Glendale plate - Mormon Pk. allochthon
Mormon plate and parautochthon
Autochthon
⊕ = horizontal bedding



Poles to Bedding in Mormon Peak Allochthon

FIG. 32

Direction of Tilting in Mormon Peak Allochthon

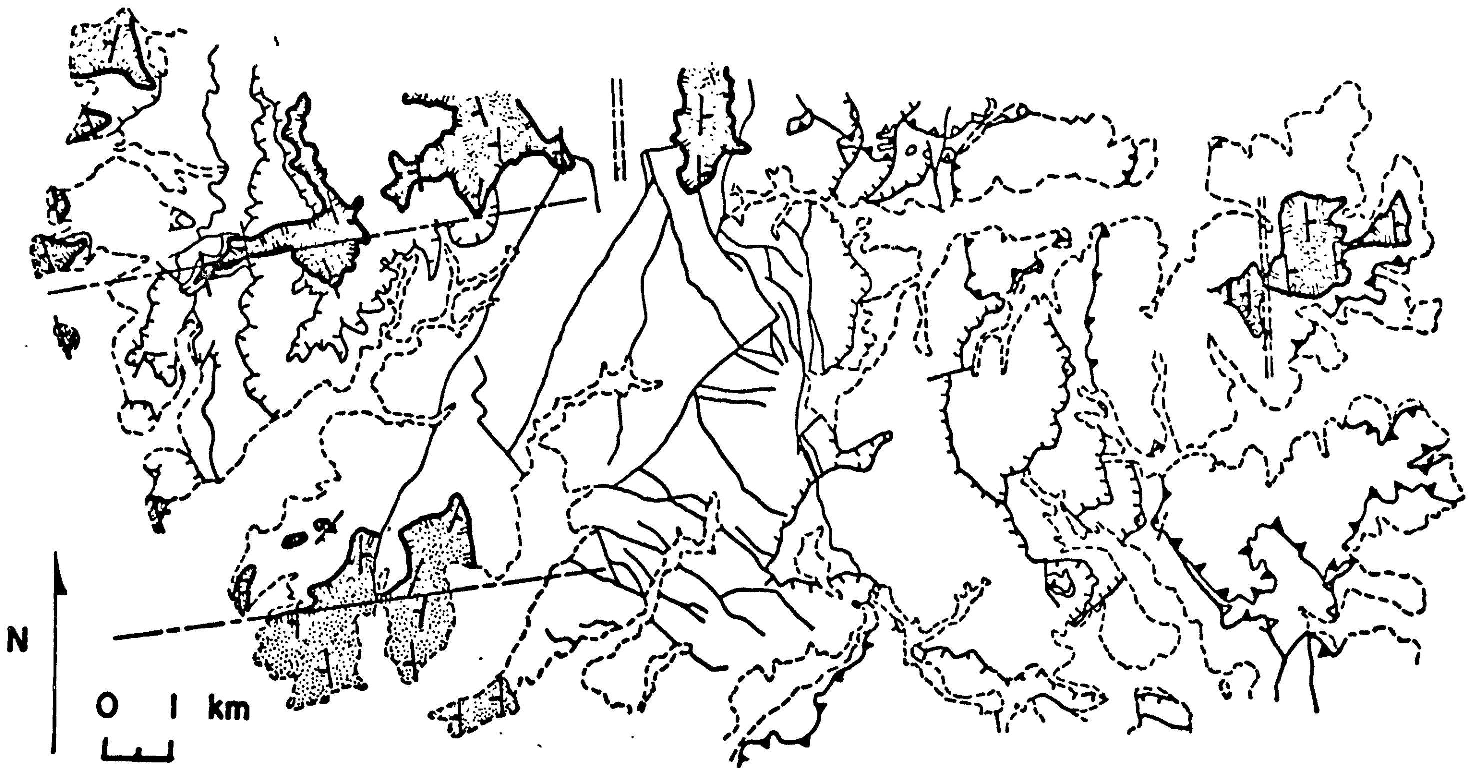


FIG. 33

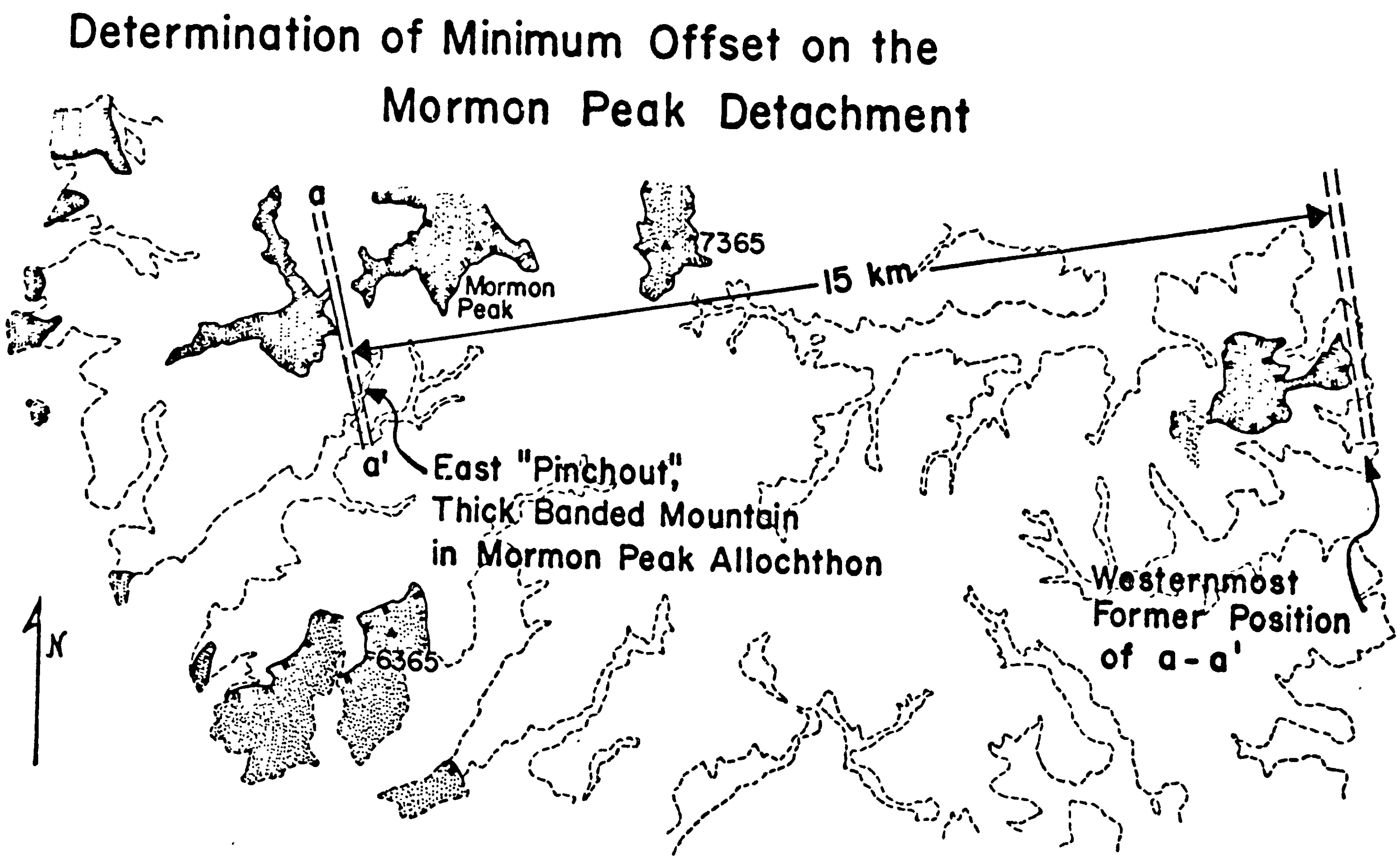

Determination of Minimum Offset on the
Mormon Peak Detachment
a
a'
Mormon Peak
7365
15 km
6365
N
East "Pinchout",
Thick Banded Mountain
in Mormon Peak Allochthon
Westernmost
Former Position
of a-a'

FIG. 34

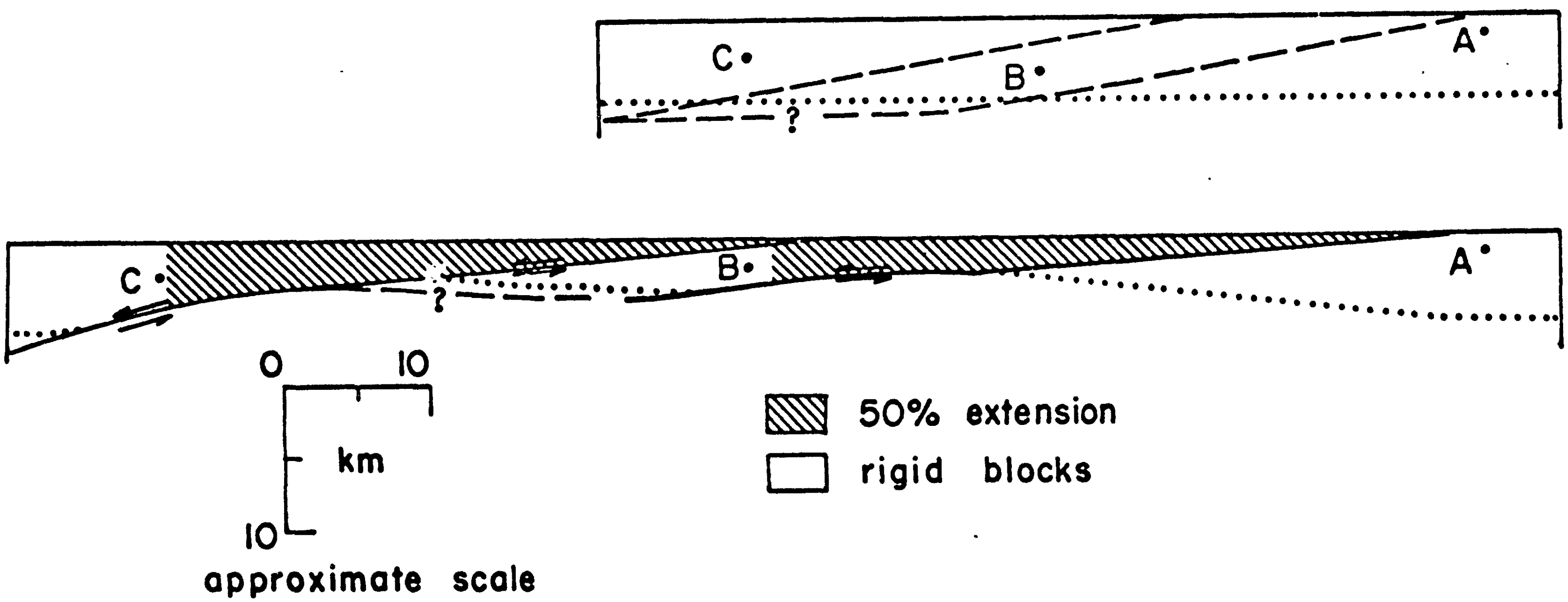

C
B
A
?
C
B
A
?
0
10
km
10
approximate scale
50% extension
rigid blocks

FIG. 35

PLATE 1

GEOLOGIC MAP OF THE CENTRAL MORMON MOUNTAINS, LINCOLN COUNTY, NEVADA

Plate II. GEOLOGIC SECTIONS, CENTRAL MORMON MOUNTAINS

NO VERTICAL EXAGGERATION

Plate III. STRATIGRAPHIC COLUMN, CENTRAL MORMON MOUNTAINS