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STRUCTURE OF THE NORTHERN CEDAR MOUNTAINS, WEST-CENTRAL NEVADA: A STUDY UTILIZING BALANCED CROSS-SECTIONS AND SURFACE DATA

by

LAUREN SHELLEY BROWN

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

MASTER OF ARTS

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STRUCTURE OF THE NORTHERN CEDAR MOUNTAINS,
WEST-CENTRAL NEVADA:
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ABSTRACT:

Thrust imbricates of the Late Triassic Luning and Jura-Cretaceous Dunlap Formations, constituting the upper plate of the regionally extensive Luning thrust, locally exhibit younger-over-older relations in the Northern Cedar Mountains. Regional relations, supported by data from the Cedar Mountains, indicate a predominantly southerly transport direction for the northerly dipping Luning thrust, which is traced through the study area. Two fold sets are observed in the Cedar Mountains, the first striking NE, the second NW. Unlike their similarly-striking counterparts elsewhere in the region, the Cedar Mountain folds and associated thrusts verge north, suggesting northerly tectonic transport. Balanced cross-sections based on both regional and detailed mapping data suggest that these structures are the product of backthrusting which occurred as a late-stage response to southerly transport of imbricated sheets of the Luning allochthon over a north-dipping decollement ramp. Local younger-over-older relations result from backthrusting of these thrust imbricates.

Mid-Cretaceous intrusions postdate thrusting and predate an episode of southerly-directed low-angle detachment faulting. The low-angle
Detachments unroof a mid-Cretaceous plutonic complex and the lower parts of the Luning allochthon.
ACKNOWLEDGEMENTS

I would like to acknowledge Dr. John S. Oldow's numerous and important contributions to this study. I would also like to thank Dr. Norman J. Silberling for his biostratigraphical work; my mother, sisters and brothers for their love, patience and support; and Maryann O'Brien, to whom in many ways I owe the completion of this work.
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INTRODUCTION

The Northern Cedar Mountains in west-central Nevada expose pre-Tertiary rocks associated with the Luning-Fencemaker fold and thrust belt (Oldow, 1984) (Figure 1). In this region Mesozoic carbonate and clastic shallow marine rocks of the Triassic Luning Formation and continental rocks of the Jurassic-Cretaceous (?) Dunlap Formation experienced late Mesozoic deformation which produced polyphase folds associated with predominantly south-directed thrusting. Late (?) Mesozoic quartz monzonite plutons intrude structures and are often truncated by NNW-trending Basin and Range faults and Tertiary volcanics.

This study was undertaken as part of an ongoing reevaluation of the pre-Tertiary structure and stratigraphy of the rocks constituting the Luning-Fencemaker fold and thrust belt. Backthrusts and younger-over-older fault relationships recognized in the Northern Cedar Mountains posed difficulties in modelling and kinematics of deformation in the area. The application of balanced cross-section techniques to surface data, taken in conjunction with regional stratigraphic relationships, produced a unique family of solutions. The results of the study have implications for such questions as the kinematics of backthrusting at a ramp, the production of younger-over-older relationships by thrust faulting, and post-compressional tectonics at the edge of thrust belts. They demonstrate the power of balanced cross-section techniques when used in conjunction with detailed surface
FIGURE 1. Generalized map of part of the western United States illustrating major tectonic features. Shaded areas are: northeastern Oregon (NO); the Klamath Mountains (KM); the Sierra Nevada (SN); and the Mesozoic marine province of the northwestern Great Basin (NGB). The hachured line represents the approximate western limit of sialic crust deduced from the initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ contour. Bold lines are known or inferred faults: the Pine Nut fault is an inferred strike-slip system formed in the Jura-Cretaceous; the Luning-Fencemaker Belt and Sevier Belt are thrust fault systems (teeth on upper plate); GT is the Golconda thrust; and RMT is the Roberts Mountain thrust (teeth on upper plate). (Adapted from Oldow, 1984a, p. 247)
mapping and argue that many structural mapping problems should be subjected to this type of analysis even when the criteria strictly justifying balanced cross-sections are not satisfied.

**Surface maps and conventional cross-sections: strengths and limitations**

Surface mapping provides virtually the only information on Mesozoic deformation in the Great Basin, as Cenozoic Basin and Range structures dominate data produced by geophysical explorations of the subsurface. The results of the mapping done during this study point out both the strengths and weaknesses of surface maps and conventional cross-sections as indicators of subsurface structure.

The most obvious map-scale structures in the study area are east-west trending fault-bound belts of the Triassic Luning Formation and Jurassic-Cretaceous (?) Dunlap Formation (Muller and Ferguson, 1939; Ferguson and Muller, 1949). The east-west trending belts were previously mapped (Muller and Ferguson, 1939; Ferguson and Muller, 1949; Mottern, 1962; Wetterauer, 1977) as parts of an east-west trending syncline/anticline pair, with the Jurassic-Cretaceous(?) Dunlap Formation forming the core of the syncline. Careful remapping (Plate 1) demonstrates that the Dunlap Formation is not the hinge of a syncline, but rather represents a single stratigraphic stack, fault-bound on both its northern and southern contacts with the Triassic Luning Formation. The northern fault, which has a very shallow southerly dip, places the Dunlap above the Luning in a younger-over-older relationship. The southern fault, with a steeper south dip, juxtaposes the Luning over the Dunlap in an older-over-younger relationship.

The regionally extensive Luning thrust, which forms the southern
boundary for upper-plate rocks of the fold and thrust belt, is exposed in the field area. Here the Luning thrust juxtaposes the highly deformed carbonates of the Luning Formation in the upper plate over the relatively undeformed clastics of the Dunlap Formation.

All of these Mesozoic structures are cut by a NNW-trending Cenozoic high-angle fault (Cz1, Plate 1). Motion on the Cenozoic fault, constrained by surface relationships, allows reconstruction of the subsurface morphology of the Mesozoic faults north and south of the exposed belt of the Dunlap Formation.

Analysis of mesoscopic fabric data indicates two phases of structures, D1 and D2. As will be developed later, D1 and D2 were formed during thrusting and have NNE and WWN trending axial planes respectively. Unlike their regionally developed counterparts, which dominantly are southerly vergent structures, the folds in the Northern Cedar Mountains predominantly exhibit a northerly vergence. The northerly-vergent structures are kinematically associated with south-dipping Mesozoic faults. Those faults which juxtapose rocks in an older-over-younger relationship may readily be interpreted as backthrusts, but the fault producing the younger-over-older geometry mentioned above (hereafter called the Dunlap fault) must be considered with more caution, as this relationship is most commonly produced by normal faulting.

Surface maps and conventional cross-sections thus provide a number of constraints and indications of possible scenarios for the structural history of the study area. They do not, however, distinguish
sufficiently between those apparent solutions which are kinematically possible and those which are not.

**Use of balanced cross-section techniques with surface data**

Balanced cross-section techniques (Dahlstrom, 1969) have been used successfully to analyze and predict subsurface structures in thrust belts (Bally et al., 1966). They provide a relatively simple framework for analysis of the kinematics of deformation by accounting for volumes of rock involved in deformation. This is accomplished by assuming that deformation is accomplished only by flexural slip folding and by faulting, with an implied stable volume of rock, which allows conservation of unit thicknesses and bed length in the cross-section for both deformed and undeformed states.

Among the benefits of using this approach is that it produces minimum shortening estimates, an important consideration since actual shortening is in general undeterminable.

Two factors may seem to mitigate against the general use of balanced cross-sections. It will be shown, however, that neither factor need pose an insuperable problem in the use of balanced cross-section techniques.

The first problem is that the technique is not usable without knowledge of the depth to and configuration of the basal detachment. In the past, the technique has been used principally with seismic data which constrains the basal decoupling surface. However, in areas near the leading edge of fold and thrust belts, the subsurface position of the basal detachment surface may sometimes be deduced within
reasonable limits from surface data. It will be shown that this is possible in the study area.

The second factor involves the assumptions allowing conservation of unit thickness and bed length. Field observations in the study area of such structures as cleavage, indicating volume loss, and mesoscopic folds, which shorten bed length in the deformed cross-section, indicate that these assumptions are not valid on a mesoscopic scale. Both indicate amounts of shortening ignored in construction of the balanced cross-section. They may nevertheless allow flexural-slip as an approximation of the mechanism of megascopic shortening. Solutions reached using flexural-slip deformation may be tested to determine the degree to which the boundary conditions must be relaxed in order to substantially change the model. Field observations are then used to determine whether the rocks were deformed in such a way as to cause such changes. It will be shown that the bulk strain of Mesozoic rocks in the Northern Cedar Mountains is not sufficiently great to warrant significant relaxation of the concentric deformation condition assumed during construction of the balanced cross-section.
REGIONAL SETTING

The study area encompasses part of the upper-lower plate boundary of the Luning-Fencemaker fold and thrust system in north-central Nevada, in the northwest Great Basin (Figure 1.) Recent papers by Oldow (1984a) and Oldow and others (1984) synthesize stratigraphic and structural relations in the region. The thrust system involves rocks of the Mesozoic marine province, complexly folded and thrust to the east and south over Paleozoic and Mesozoic rocks of continental and shallow marine affinities. The Luning-Fencemaker thrust belt is truncated on the west against the Mesozoic Sierran arc along the Pine Nut system, which is interpreted to be a Juro-Cretaceous transpressional fault active during thrusting to the east (Oldow, 1983 and 1984a; Oldow et al., 1984).

Mesozoic marine province rocks consist of volcanics, terrigenous clastics, and carbonates deposited in and around a marginal backarc basin east of the Mesozoic Sierran arc (Muller and Ferguson, 1939; Silberling and Roberts, 1962; Speed, 1978a; Oldow, 1983 and 1984a). The basin was open to the north and elsewhere bounded by shallow-marine shelf and subaerial environments (Oldow, 1984a). Eastern and southern basin margins roughly coincide with an inferred crustal boundary between the western edge of sialic North America and oceanic-affinity crust that underlies the Luning-Fencemaker allochthon. Deposition apparently was controlled by subsidence of noncontinental crust (Speed, 1987a). Location of the crustal boundary is estimated by a number of workers (Roberts et al., 1958; Eaton, 1963; Hill and Pakiser, 1966;
Doe, 1973 and 1978; Armstrong et al., 1977; Kistler, 1978; Cogbill, 1979; Prodahl, 1979; Kistler et al., 1981, Leeman, 1982; Farmer and Depaulo, 1983) using stratigraphic relationships, geophysical data including regional gravity and seismic refraction studies, and strontium and other isotopic systematics. Strontium isotopic studies of Mesozoic plutons, considered autochthonous with respect to their host rocks, have provided the most abundant data, and the initial strontium contour, $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ is used to delineate the boundary (Figure 1).

The origin of noncontinental crust underlying the basin is controversial. It is proposed to have originated as a backarc basin during the Paleozoic (Burchfiel and Davis, 1972 and 1975) or alternatively by tectonic accretion of an oceanic microplate during the Permo-Triassic (Speed, 1979). The curvilinear trend from southerly to westerly of the inferred crustal boundary, the Luning-Fencemaker thrust, Golconda thrust, Roberts Mountain thrust, and Mesozoic and Paleozoic facies patterns, have been cited as tectonic in origin (Albers, 1967; Wetterauer, 1977). However, recent structural (Oldow, 1984b) and paleomagnetic studies (Oldow and Geissman, 1982; Geissman, et al., 1984) support an earlier interpretation of Ferguson and Muller (1949) that the shape of the boundary is a preserved feature formed during Precambrian (Stewart, 1972) or possibly later rifting (Kistler, 1978), which has profoundly affected subsequent deposition and deformation, including that which formed the Luning-Fencemaker thrust system.
Regional development of structures

Three phases of contraction structures are regionally-developed in the rocks of the Mesozoic marine province. These phases are sequenced by superimposition of structures and are here termed D1, D2, and D3 in chronological order of occurrence. The upper age limit for D1 and D2 is constrained throughout the region as pre-Late Cretaceous (100-90 my) by crosscutting plutons (Speed and Kistler, 1980; Oldow, 1984a). Timing of inception of D1 deformation appears to have differed in different parts of the region. A reported syntectonic pluton dated by K/Ar as between 165 and 145 my established that deformation began at least locally during the Middle or Late Jurassic (Speed, 1974; Willden and Speed, 1975). In other locations, rocks affected by D1 deformation are dated faunally as Early Cretaceous (Willden, 1958 and 1964) and by Rb/Sr as 103.5 my (Speed and Kistler, 1980). Deformation was probably time-transgressive, occurring in a migrating belt or belts (Oldow, 1984a). D2 deformation is constrained as post-103.5 my and pre-90 my by the youngest rocks deformed and by crosscutting plutons (Oldow, 1984a). For reasons discussed below, the timing of D3 deformation cannot be constrained except as post-D2.

D1 structures formed in response to great NW-SE shortening during emplacement of the Luning-Fencemaker allochthon. Deformation is expressed in NE-trending tight to isoclinal folds with axial plane cleavage. D1 structures are well developed in the allochthon, but are found only locally in the lower plate, in a belt subjacent to the overlying thrusts (Oldow, 1981a and 1981b).
D2 deformation is expressed in NW-trending close to tight folds with locally developed axial plane cleavage. It locally formed minor thrusts, generally SW-directed. This phase is much better developed in the lower plate than D1, and is uniform in orientation in lower-plate exposures. In the upper plate, D2 exhibits great primary variability both in orientation and intensity (Oldow, 1981a and 1981b).

D3 fold sets have both N-S and E-W trending axial planes. These sets are found in the same mountain ranges but are never superimposed and may have developed contemporaneously. Because of their slight to moderate limb oppression and local development, their relation to Mesozoic plutons crosscutting D1 and D2 is unclear. Age limits for this deformation are therefore undetermined, except that it postdates D1 and D2 and predates Tertiary volcanism.

Lithotectonic assemblages

Lithologic and structural differences in time-correlative packages of pre-Tertiary rocks within the province have prompted its division into lithotectonic assemblages by Oldow (1984a) (Figure 2). Supporting data for the following discussion is synthesized in that paper. Assemblage boundaries in several locations are known thrusts; at other locations faults are inferred. Five dismembered allochthonous assemblages (I, II, IV, V, VI in Figure 2) comprise the upper plate of the Luning-Fencemaker system. They are bordered to the south and east by two parautochthonous assemblages (III and VII) and on the west by coherent allochthonous terrane (VIII) that has structural affinities with the Sierran arc. The interested reader is referred to Oldow, 1984a.
FIGURE 2. Generalized map of pre-Tertiary rocks of the northwestern Great Basin. Bold lines are known faults, dashed where inferred; teeth are on the upper plate of thrust faults. Roman numerals delineate lithotectonic assemblages: I-Black Rock assemblage; II-Lovelock assemblage; III-Humboldt assemblage; IV-Sand Springs assemblage; V-Pamlico assemblage; VI-Luning assemblage; VII-Gold Range assemblage; and VIII-Pine Nut assemblage. Shaded areas represent Mesozoic and Paleozoic rocks constituting dismembered lithotectonic assemblages; horizontally ruled areas are autochthonous or paraautochthonous assemblages; and diagonally ruled areas are exposures of an internally coherent allochthonous assemblage (VIII). (From Oldow, 1984a, p. 248).
The E-W trending Luning thrust forms the southern boundary of upper plate (dismembered allochthonous) rocks in the Luning-Fencemaker system (Figure 2). The Pamlico (V) and Luning (VI) assemblages are juxtaposed with the Gold Range (VII) parautochthonous assemblage along the thrust. The Northern Cedar Mountains lie principally within the Luning assemblage. The Luning thrust traverses the southern periphery of the study area, where exposures of the Gold Range assemblage were mapped. Regional stratigraphic relations indicate that the Luning thrust trace changes to approximately N-S orientation a few kilometers east of the Northern Cedar Mountains. In its mapped and inferred positions the Luning thrust approximately parallels the 0.706 contour which lies to the south and east.

**Relationship with other pre-Tertiary fold and thrust belts**

Three other pre-Tertiary fold and thrust belts lie east of the Luning-Fencemaker system in Nevada and western Utah (Figure 1). Deformation in the mid-Paleozoic Roberts Mountain and Permo-Triassic Golconda thrust systems apparently was also controlled by the crustal boundary delineated by the 0.706 line (Silberling and Roberts, 1962; Speed, 1977a and 1979; Speed and Sleep, 1982; Oldow, 1984a and b). Following the interpretation of Ferguson and Muller (1949) and subsequent workers, this boundary is inferred to represent the Paleozoic sialic North American continental margin. The trend of the Sevier thrust system in western Utah parallels the Paleozoic miogeoclinal hinge line (Burchfiel and Davis, 1972 and 1975; Oldow, 1984a).

The Roberts Mountain system was emplaced during the Devonian-
Mississippian Antler orogeny. Lower continental slope and continental rise sedimentary and volcanic rocks of early Paleozoic age were thrust easterly over early Paleozoic carbonate and clastic continental shelf rocks, resulting in over 100 km of displacement (Roberts et al., 1958; Smith and Ketner, 1968; Oldow, 1984b). The system is proposed variously as the product of the collapse and thrusting of a marginal backarc basin (Burchfiel and Davis, 1972 and 1975) and as the result of accretionary prism obduction during westward subduction of a passive continental margin beneath a volcanic arc (Speed, 1977a and 1979; Speed and Sleep, 1982).

The Golconda thrust lies between the Roberts Mountain and the Luning-Fencemaker thrusts. It juxtaposes Pennsylvanian/Permian basinal rocks over temporally equivalent rocks with continental borderland affinities (Silberling and Roberts, 1962; Speed, 1977a). Timing of Golconda emplacement is poorly constrained but is thought to be during the Early Triassic (Silberling, 1975). Origin of the system is controversial. It is proposed as the second stage of backarc basin collapse (Burchfiel and Davis, 1972 and 1975) or as the result of accretionary prism obduction associated with the emplacement of simatic crust lying west of the 0.706 line (Speed, 1977a, 1979, and 1983). In either case, the fault system is accepted as being emplaced prior to deposition of rocks of the Mesozoic marine province.

The trace of the Sevier thrust tracks the depositional hingeline inferred for the Paleozoic miogeoclone (Oldow, 1984a). This suggests that crustal thickness changes at the hingeline may have localized the deformation that produced the Sevier system. Age of inception of
Sevier thrusting may be the same as that of the Luning-Fencemaker system (Oldow, 1984a; Allmendinger and Jordon, 1981). Both were active at the same time for at least parts of their histories.

The Sevier and Luning-Fencemaker systems are separated by a "quiet zone" which underwent relatively little deformation during this time. Speed (1978) and Oldow (1983, 1984a) propose that this "quiet zone" was detached at depth, suggesting that the Sevier and Luning-Fencemaker systems may have shared a common decollement. This implies that rocks termed "paraautochthonous" with respect to Luning thrusting were themselves part of the Sevier allochthon.
REGIONAL STRATIGRAPHIC FRAMEWORK

Rocks of the Luning and Gold Range assemblages (Oldow, 1983, 1984a) are juxtaposed by the Luning thrust in the Northern Cedar Mountains (Figure 2). The Luning assemblage is a structurally dismembered succession composed of numerous thrust sheets, whereas the Mesozoic rocks of the Gold Range assemblage overlie upper Paleozoic successions with angular unconformity, and as such are considered parautochthonous. As discussed earlier, it is proposed that a regional detachment, serving to connect the Luning-Fencemaker and Sevier thrust belts at depth, exists beneath the Gold Range assemblage and underlying Paleozoic rocks to the south and east (Oldow, 1984a). Since all pre-Tertiary rocks in the region probably are allochthonous, it is important to bear in mind that estimates of contraction for this region are based on stratigraphic relationships and do not incorporate displacements of the so-called parautochthon.

Without the benefit of subsurface data, regional stratigraphic relationships must be coupled with surface data derived from the Northern Cedar Mountains in order to assign decoupling surfaces for construction of a balanced cross-section. The stratigraphic framework of the rocks exposed in the Northern Cedar Mountains and vicinity is outlined below and in Figures 3, 4 and 5.

Luning Assemblage

The bulk of the rocks in the Luning assemblage are shallow marine carbonates and clastics of the Late Triassic Luning Formation and the
undifferentiated late Late Triassic to Early Jurassic Sunrise-Gabbs Formation (Muller and Ferguson, 1939; Ferguson and Muller, 1949; Oldow, 1981a). The shallow marine carbonate and clastic succession is overlain by the Early Jurassic to Early Cretaceous (?) Dunlap Formation, which consists of a heterogeneous sequence of fine to coarse clastic rocks locally interbedded with shallow-marine carbonates (Muller and Ferguson, 1939; Ferguson and Muller, 1949; Oldow, 1981a). The stratigraphic units are only rarely found in depositional contact, and in most cases they are juxtaposed by thrust faults.

Where exposed, the base of the Mesozoic sequence is generally a thrust contact with the Luning Formation constituting the oldest stratigraphic unit. An important exception to this regional relationship lies in the Shoshone Mountains (Figure 3). There the Luning Formation disconformably overlies lithologically similar rocks of the Middle Triassic Grantsville Formation (Silberling, 1959). This contact represents the only known lower depositional contact for the base of the Luning Formation. The Grantsville rests disconformably on a volcanic and volcanogenic unit which is enigmatic in its regional significance and age. Assessment of the regional significance of the volcanic unit, referred to here as the Shamrock succession in deference to the formal designation of Shamrock Formation proposed by Lankford et al. (in prep.), is hindered by uncertainty in its age. The Shamrock has had age assignments ranging from Pennsylvanian through Lower Triassic (Ferguson and Muller, 1949; Silberling, 1959 and 1973). The problem of the Shamrock's stratigraphic affinity may be resolved by
FIGURE 3. SW
Luning assemblage
comparative stratigraphic
columns.

KJd_u  Dunlap Formation, upper
KJd_l  Dunlap Formation, lower
JTrsg  Sunrise & Gabbs Formations

Tr_l u  Luning Fm., upper carbonate
Tr_l ms  Luning Fm., middle shaley
Tr_l mc  Luning Fm., middle clastic
Tr_l l  Luning Fm., lower carbonate
Trg  Grantsville Formation
PTrs  Shamrock assemblage

1From Muller & Ferguson, 1939;
Oldow, 1981a

2this study

3Silberling, 1959; Wetterauer, 1977; Oldow, unpublished data
radiometric age dating now in progress (Oldow et al., in prep).

Elsewhere in the Luning assemblage, the oldest rocks consist of thrust slices of mafic volcanic and volcanogenic units of the Permian Black Dyke Formation (Speed, 1977). The Black Dyke is interpreted to represent slices of the basement complex upon which the Mesozoic basinal sequence was deposited (Speed, 1978, 1979; Oldow, 1978, 1981a, 1983, 1984a; Seidensticker and Oldow, in press). The petrographies of the volcanic components of the Black Dyke and Shamrock are comparable (Seidensticker and Oldow, in press) but confirmation of a possible correlation must be deferred until better age constraints are available for the Shamrock.

A regional decoupling surface between the base of the Mesozoic sequence and the upper Paleozoic volcanic complex apparently exists and serves as the basal detachment for the Luning-Fencemaker thrust system in this region. Rocks of the Luning, Sunrise-Gabbs, and Dunlap Formations constitute numerous thrust sheets, several of which have documented displacements of tens of kilometers (Oldow, 1981a, 1981b), but rocks of the older volcanic units are seldom incorporated in the thrust stacks. Considered in light of the degree of shortening and imbrication of the Mesozoic sequence and the general lack of involvement of the older volcanic units, the existence of basal decollement beneath the Mesozoic sequence is required. This decoupling surface is of major importance in the construction of a balanced cross-section.

Zones of decoupling exist within the Mesozoic sequence, but the only regionally significant surfaces correspond with interformational boundaries. Within the Luning Formation, the regional distribution of
shallow marine carbonates and shallow marine to deltaic clastic rocks (Figure 3) does not define a single stratigraphic horizon that can be used as a regional detachment. Rather, imbrication within the Luning follows many different stratigraphic horizons and does not define a regionally significant system. The contact between the Luning and Sunrise-Gabbs Formations, on the other hand, is a regional feature. The conformable contact generally exhibits a sharp lithologic change from the interbedded carbonate and dolomite of the upper Luning to thin bedded distal shelf carbonates and fine grained clastics of the Sunrise-Gabbs Formation. The abrupt lithologic change and apparent lateral continuity of the contact regionally serve to make it an important decoupling surface. Unlike its lower contact, the upper boundary between the Sunrise-Gabbs and the overlying Dunlap Formation is less distinct. Where sufficient fossil control exists (Silberling, 1984; Oldow, 1984) it is apparent that the lithologic transition from the fine-grained carbonates and clastics of the Sunrise-Gabbs to the coarse clastics of the Dunlap is diachronous and quite variable in its character regionally. Nevertheless, a preferred zone of decoupling forms in the vicinity of the formational boundary.

**Gold Range Assemblage**

Stratigraphic relationships within the Mesozoic rocks of the Gold Range assemblage (Oldow, 1983, 1984a) are complex and details are poorly understood. The Mesozoic rocks consist of a heterogeneous pile of volcanic and coarse clastic rocks lying beneath the Luning thrust. Internal stratigraphic relationships are locally obscured by
FIGURE 4. Generalized distribution of pre-Tertiary layered rocks in the vicinity of the Northern Cedar Mountains, west-central Nevada.
minor thrust imbrication. Where lower depositional contacts are preserved, the sequence overlies the Permian Mina Formation (Speed, 1977b) with angular unconformity. The Mina Formation is a complex of imbricated and interbedded chert, argillite, and volcanogenic turbidites interpreted to be related to the volcanic Black Dyke Formation found within the Luning assemblage (Speed 1977b, 1979). The Mina Formation experienced internal deformation during the emplacement of the Permo-Triassic Golconda thrust system prior to the deposition of the Mesozoic sequence of the Gold Range assemblage (Speed, 1977b).

It is significant that several stratigraphic units overlie the Mina Formation (Figure 5). Within the Pilot Mountains the Dunlap Formation overlies the Mina with dramatic angular unconformity. Locally, however, an intervening unit, the informally designated Water Canyon sequence, unconformably underlies the Dunlap and overlies the Mina with angular unconformity. Elsewhere, in the eastern Excelsior Mountains (Figure 4) the Mina Formation is overlain in angular unconformity by both the Cretaceous Gold Range Formation (Speed 1977b, as revised in Speed and Kistler, 1980) and the Jurassic and Cretaceous (?) Dunlap Formation. In the Excelsior Mountains the lateral relationship between the Gold Range and Dunlap is not sufficiently well studied, but preliminary results of work in progress (Bartel and Oldow, unpublished data) suggest that they are separated by a major unconformity.

With the exception of the Dunlap Formation, stratigraphic units within the Gold Range assemblage are not laterally extensive. An additional unit, the Jurassic Telephone Canyon succession, which
feldspathic sand
volcanic rocks
chert breccia
volcanogenic sand
bedded chert
lenticular lmst.

Kgr  Gold Range Formation
KJd_u  Dunlap Fm., upper
KJd_l  Dunlap Fm., lower
JTrwc  Water Canyon assemblage
Pm  Mina Formation
Pbd  Black Dyke Fm.

EXCELSIOR MTNS.¹  S. PILOT MTNS.²  N. & S. CEDAR MTNS.³

¹ from Muller & Ferguson, 1939; Speed, 1977; Speed & Kistler, 1980
² from Wetterauer, 1977; Oldow, 1981
³ from Muller & Ferguson, 1939; Ferguson & Muller, 1949; Wetterauer, 1977; and this study.

FIGURE 5. Gold Range assemblage comparative stratigraphic columns
consists of volcanic and coarse clastic rocks, has an undetermined relationship with the other lithologic sequences (Oldow, 1981a). The lateral discontinuity of stratigraphic horizons probably has contributed to the lack of development of areally extensive decoupling surfaces within the Gold Range assemblage.

Assessment of the depositional history of the Gold Range assemblage sheds light on the tectonic development of the Luning thrust system. Deposition of the Mesozoic rocks is interpreted as having occurred along the tectonically active margin of basinal deposition associated with the marine province of the northwestern Great Basin. As previously indicated, the basin margin corresponds spatially with an old crustal boundary. The history of sedimentation of the younger successions, particularly the Gold Range Formation, may have been strongly influenced by contractional deformation associated with the emplacement of the Luning thrust. Syntectonic deposition of the Dunlap Formation has long been recognized (Ferguson and Muller, 1949) but the nature of the associated tectonism has been controversial. Ferguson and Muller (1949) originally proposed that Dunlap sedimentation was in response to regional contraction and the emplacement of the Luning thrust. Detailed study of the Dunlap Formation discounted this hypothesis (Stanley, 1972; Wetterauer, 1977) but did not propose a tectonic mechanism for the synorogenic deposition nor for the existence of profound lateral variations in thickness developed during sediment accumulation. Recent work (Bartel and Oldow, 1984, unpublished data) indicates that the Dunlap exposures of the Pilot Mountains accumulated
in shallow marine and subaerial environments strongly affected by extensional tectonism. Thick accumulations (in excess of 1,500 m) of coarse grained alluvial and shallow marine clastic rocks formed in response to the development of NNW striking growth faults, which in this area had a minimum throw in excess of 1.0 km.

The existence of major Jurassic extensional tectonism, presuming that growth faults were not spatially restricted to the Pilot Mountains, probably had a significant impact on subsequent thrusting. If, as seems likely, the extensional tectonism was a regional phenomenon, the stratigraphically controlled decoupling surfaces within the upper plate of the Luning thrust would have had substantial structural relief prior to contraction. During contraction it is predicted that the previously extended terrane would serve to invert the half-grabens, presuming that the extension and contraction were subcoaxial. In this region, the development of NNW striking growth faults is subparallel to the NW to SE axis of subsequent contraction (Oldow, 1981a, 1981b, 1983, 1984). Thus, it is more plausible that the extensional faults, where they were reactivated, assumed the role of tear faults. The existence of reactivated growth faults within the upper plate of the Luning thrust helps explain several previously poorly understood stratigraphic relationships, particularly among imbricate packages of the Dunlap Formation.
STRATIGRAPHY OF THE NORTHERN CEDAR MOUNTAINS

The Luning and Dunlap Formations of the Luning assemblage comprise the bulk of exposures in the Northern Cedar Mountains. These are structurally juxtaposed with rocks of the Dunlap Formation of the parautochthonous Gold Range assemblage by the east-west trending Luning thrust (Plate 1). The Mina Formation, not exposed in the study area, is found to unconformably underlie the Dunlap in exposures of the Gold Range assemblage in the Southern Cedar Mountains, approximately 6 km to the south. Detailed lithologic descriptions are contained in Appendix I.

Rocks of the Luning Assemblage

The Luning and Dunlap Formations, which comprise the pre-Tertiary rocks in the Luning assemblage in the Northern Cedar Mountains, were first identified there by Muller and Ferguson (1939) in their survey of the region. The Luning Formation was subdivided into lithologic members in the Northern Cedar Mountains by Mottern (1962). His nomenclature is modified in this paper. The internal lithologic divisions of the Dunlap Formtaion defined for the region by Wetterauer (1977) were identified and used in the Northern Cedar Mountains during this study. A stratigraphic column for the rocks constituting the Luning assemblage in the Northern Cedar Mountains is presented in Figure 6.

East-west trending, south-dipping faults form the contacts between the Luning and Dunlap Formations and are also recognized within the Luning Formation. Nappes are designated I through V from south to
FIGURE 6. Northern Cedar Mountains, Luning assemblage stratigraphy.

* Identifications and biostratigraphy by N.J. Silberling
north in structurally descending order. These thrust nappes are further subdivided into eastern and western domains for ease in discussion of internal stratigraphic and structural relationships.

The upper Triassic Luning Formation is the oldest unit identified in the Northern Cedar Mountains. In light of regional stratigraphic relationships, this implies that the decoupling level for the Luning thrust lies below or within the Luning Formation.

Rocks of the regionally extensive Sunrise and Gabbs Formations, present in most major Luning assemblage exposures, are not identified in the Northern Cedar Mountains.

**Luning Formation:** In its type locality in the Pilot Mountains, the Luning Formation is dated as Lower to Middle Norian in age and is divided into three members (Oldow, 1981a). Elsewhere in the region it is locally divided into four or five units (Silberling, 1959; Mottern, 1962; and this study). Regionally, the lower member is composed of interbedded carbonates and minor fine-grained terrigenous clastics, and is exposed in the Pilot Mountains, the Paradise Range and the Northern Cedar Mountains. It varies in thickness and proportion of clastic constituents between mountain ranges, but is relatively uniform in thickness and composition within individual mountain ranges. The middle member consists of interbedded fine-to-coarse grained terrigenous clastics and highly fossiliferous carbonates and calcareous mudstones. The clastics include interbedded chert pebble and boulder conglomerate, arenite, wacke, and sandy mudstone, each with sand constituents composed of chert, quartz, and, locally, feldspar grains. The middle
member is locally subdivided on the basis of lithology into two or three units (Figure 3). The clastic rocks of the middle member are laterally discontinuous over the region and probably are not time-correlative (Oldow, 1981a and 1984, personal communication). The upper member is a compositionally uniform, widely distributed platform carbonate/dolomite succession, and unlike the stratigraphically lower divisions may be essentially time-correlative throughout the region.

The Luning Formation forms the bulk of Mesozoic rocks in the Northern Cedar Mountains, with a total estimated stratigraphic thickness of approximately 2.2 km. Thickness estimates are complicated by internal mesoscopic folds which may have gone unrecognized due to the poor preservation of exposures. Two relatively complete sections are exposed in different thrust nappes south of the Mesozoic pluton which forms the central portion of the study area (Plate 1). Nappes I and II, separated by the South thrust which is thought to be of minor displacement, comprise one section. The other section composes nappe IV. Lower contacts of both are obscured by younger plutonic or volcanic rocks, while uppermost sections are truncated by faults.

In the study area, the Luning Formation is divided into four units. The lower carbonate unit is 500 m thick and is composed principally of carbonates which are intercalated with minor fine-grained siliciclastics in the upper 50 m. This unit appears to be analogous to the lower member of the type section in the Pilot Mountains to the west. Two lithologic units are recognized in the part of the section that is analogous to the middle carbonate-clastic member in the Pilot Mountains type section. In the Northern Cedar
Mountains, clastics (siliceous mudstones, chert pebble conglomerates and quartz/chert sandstones and wackes) comprise the lower 525 m and are here termed the middle clastic unit. Highly fossiliferous micrites, calcareous mudstones and wackes, totalling 800 m thickness, overlie the middle clastic unit and are termed the middle shaley unit. The upper carbonate unit, ranging from 350 to 375 m thick, is composed of massive, thick and medium-bedded fine-grained carbonates and dolomites of the widely distributed upper-member platform succession.

The most complete section of the lower carbonate unit of the Luning Formation is preserved in nappe IV, where it is approximately 500 m thick. The lower contact is obscured by the Mesozoic quartz pluton to the north. The member is composed principally of thin-bedded micrite. Fossils are poorly preserved. The contact with the overlying middle clastic member differs between thrust nappes. In nappe IV the upper contact is gradational over approximately 50 m. Here thin-bedded siliceous silty micrites are interbedded with carbonate clasts in a carbonate matrix, and siliceous mudstones. The top of the unit is arbitrarily placed at the last carbonate bed. In nappe I the contact of the lower carbonate unit with overlying chert pebble conglomerates of the middle clastic unit is conformable but sharp. Here a maximum thickness of approximately 50 m is exposed, the lower contact being obscured by Tertiary volcanics. Spirifirid brachiopods Septocardia sp. diagnostic of the upper Triassic (identified by N.J. Silberling, 1983) were collected from the lower carbonate unit in nappe I. Gervillia-like pelecypods were collected from highly deformed, poorly exposed carbonate at the northern boundary
of the Gold Range assemblage rocks in the study area. This fauna is found elsewhere in association with *Septocardia* in the Luning (N.J. Silberling, personal communication).

The middle clastic unit of the Luning Formation appears laterally uniform within, but variable between thrust nappes. A complete section is preserved only in nappe IV. The lower 150 m is comprised of 1- to 3-m beds of well-rounded small chert pebble (2 to 5 cm) conglomerate interbedded with siliceous mudstones which also form the conglomerate's matrix support. Conglomerate beds are laterally continuous over several tens of meters. The next 100 m is comprised of medium- to thick-bedded subangular small (1 mm to 4 cm) chert pebble conglomerate interbedded with siliceous mudstone and coarse chert arenite and wacke with a minor quartz component. The abundance of conglomerate beds decreases upsection. Mature green quartz/chert arenite and wacke in siliceous mudstone matrix (microcrystalline chert in thin section) dominate the next 150 m. Quartz content increases upsection from approximately 50% to 80% of the whole rock. Chert decreases from approximately 15% to negligible quantities. The upper 140 m consists of siliceous mudstone interbedded with lesser quartz arenite. Ridge-forming subrounded small (1 to 3 cm) chert pebble conglomerate occurs approximately 50 m from the top of the member. Contact with carbonates of the overlying middle member is gradational over 14 m.

In localities other than nappe IV, rocks of the middle clastic unit are poorly preserved, owing to extensive metamorphism of the siliceous clastics by the Mesozoic pluton and by Tertiary volcanics.
Evidence for this is best in exposures north of the pluton, where isolated outcrops of siliceous mudstone have gradational contacts with aphanitic to fine-grained plutonic rocks. This siliceous member appears to have been preferentially melted by the pluton, while contiguous carbonates were strongly metamorphosed but preserved. In nappe I, all but the basal 50 m of this unit have been obscured by siliceous Tertiary volcanics. Here the preserved exposure is composed of chert pebble conglomerate in siliceous mudstone matrix. Pebbles are generally larger than those in nappe IV, reaching diameters of up to 10 cm.

The middle shaley member of the Luning Formation is preserved in nappes I, II, and IV, and in nappe V, north of the pluton. A near-complete section, of approximately the same thickness as in nappe IV, is found in nappe I, but the lower contact is lost beneath Tertiary volcanics. The contact with the overlying upper carbonate unit of the Luning is conformable but sharp. In nappes IV and V the lower contact is transitional with the middle clastic unit over a few tens of meters. This unit is distinguished from the overlying carbonate unit by higher terrigenous clastic and fossil abundances, and by a decreased resistance to weathering. The lower 250 m in nappe IV is comprised of massive slope-forming cherty micrite with few preserved fossils. A greater preponderance of resistant rocks, lower chart content, and abundant fossils characterize the next 250 m. Megafossils are locally abundant and well-preserved at several horizons in nappe I. Large fluted clams *Trichites* in this nappe were identified as Lower to Middle Norian in age by N. J. Silberling.
(written communication, 1983). Coquinoïd limestone and calcareous siltstone comprise the upper 350 to 450 m in nappe I. In nappe V this unit contains the youngest rocks preserved.

Specimens of pelecypods Halobia cordillerana?, crinoids Halocrinus ornatissimus, and ammonites Mojsisovicsisi, collected from the middle shaley unit in nappe IV by H.G. Ferguson in 1928, were recently associated with the lower Lower Norian or possibly Karnian Kerri zone by N.J. Silberling (written communication, 1983). These fossils are possibly older than those identified in nappe I, but their position within the unit is not known and they may come from older horizons.

(NOTE: Fossils collected in the middle shaley member at the western boundary of nappe IV by N.J. Silberling (oral communication, 1985) after the preparation of this study are associated conclusively with the Karnian Kerri zone. This exposure is thus significantly older than its analogue in nappe II. This new data implies that the upper bound of the middle shaley unit in nappe IV is a fault which cuts out the upper strata, and possibly that this unit is significantly thicker than estimated previously. Offsetting this possible increase in thickness is the discovery of significant mesoscopic folding, recognized by N.J. Silberling on the basis of reversals and repetitions in paleontologic horizons within the unit (oral communication, 1985). The fault is thought to be of minor displacement. Preliminary investigation (Oldow and Brown, in prep.) of the implications
of this new data indicates that a greater amount of shortening is called for than proposed in this study. However, this information does not significantly alter the geometry or kinematics of the model discussed here.)

The upper carbonate unit of the Luning Formation is preserved only in nappes II and IV. Both upper contacts are truncated by faults. Nearly identical thicknesses of this section are preserved in the two nappes. The unit is composed of resistant, slightly recrystallized carbonate wacke, micrite, calacarenite, dolomite, and calcareous shale, with rare quartz sandstone/wacke lenses. The lower two-thirds of this unit is comprised of massive, resistant thick-bedded calcareous shales and mudstones. Beds vary laterally in thickness. Massive units decrease upsection, both in thickness and frequency of occurrence. The upper third of this part of the section is composed of sparsely fossiliferous, thin- to medium-bedded micrite and calcareous shale. Specimens of Alectryonia and "M. Whatelaya," collected by H.G. Ferguson from the upper section of this unit in nappe II in 1928, were tentatively associated by N.J. Silberling (written communication, 1983) with the Magnus zone of upper Lower Norian age. This member is uniform within, but somewhat varied between the two thrust sheets represented in nappes II and IV. Units in nappe IV are generally more massive and thicker-bedded than their equivalents in nappe II.

**Dunlap Formation:** The Dunlap Formation (Muller and Ferguson, 1939; Ferguson and Muller, 1949; Stanley, 1971; Wetterauer, 1977)
is found in the upper plate of the Luning thrust in the Excelsior Mountains, Paradise Range and the Shoshone Mountains (Figures 3 and 4). It overlies the Sunrise-Gabbs Formations with erosional disconformity and local conformity where depositional contacts are observed. The formation, which is the youngest unit preserved deposited prior to the emplacement of the Luning allochthon, consists of shallow marine to subaerial terrigenous and volcanogenic clastics and volcanics with minor, generally nonfossiliferous carbonates. The lack of fossils makes determination of the age range of this unit difficult. Fossils dated by Silberling (1981) from the Excelsior Mountains are Sinemurian in age, while fossils from the Garfield Hills were dated as Pliensbachian. There is however some uncertainty as to whether the formation from which the latter were derived was the Dunlap or the Sunrise-Gabbs. The upper age bound of this formation within the Luning assemblage is constrained only as being prior to emplacement of the Luning allochthon, and is possibly as young as Early to Middle Cretaceous.

The upper-plate Dunlap Formation is divided into two members based on the lithology of sand-sized components. The lower member is characterized by predominance of mature quartz sand, the upper by the influx of feldspathic sand components (Wetterauer, 1977). Basal limestone pebble conglomerates and breccias overlain by red quartz sandstones with occasional chert pebble layers and thin interbeds of limestone comprise the lower member. The upper member consists of quartz-, chert- and feldspar sandstones, wackes, silt- and mudstones with occasional pebble conglomerate lenses with chert, limestone and
lithic clast lithologies. Intermediate and siliceous volcanics form a part of the upper member in western exposures. Carbonates in the unit occur as generally nonfossiliferous lenses of shallow marine to fresh water affinity.

The Dunlap is regionally distinguished by extreme variability in lithology, thickness and depositional environment (Muller and Ferguson, 1939; Ferguson and Muller, 1949), clastic constituent lithology and its red color.

In the Luning assemblage in the Northern Cedar Mountains the Dunlap Formation is approximately 1400 m thick. This unit is a single stratigraphic "stack", the basal contact of which is truncated by the Dunlap fault which bounds nappe III on the north. Both of the formation's members, defined by Wetterauer (1977) on the basis of sand-sized component lithology, were identified in the study area. The lower member, characterized by the predominance of quartz sand, is approximately 500 m thick. Feldspathic sand components characterize the upper member (approximately 850 m thick), although they form a smaller constituent in the study area than in localities to the west. Volcanic rock fragments and flows and tuffs characteristic of the upper member in western localities are absent in the Dunlap of the Northern Cedar Mountains. The Dunlap is exposed only in nappe III, which is comprised entirely of this formation. Outcrop is poor, with 60% to 70% of the unit obscured by float and alluvium. The red color characteristic of the Dunlap has multiple origins. Quartz grains appear to have been stained prior to final deposition, while hematite from decomposing mafic grains visible in thin section provides authigenic stain in mudstones and feldspathic units.
The lower quartzose member is found only east of a NNW-trending high-angle fault (C21, Plate 1), and is cut out to the west by the Dunlap fault. Predominantly calcareous breccia and conglomerate forms the basal 2 m of this member. Calcareous siltstone and minor chert clasts are angular to subrounded, 2 mm to 6 cm in diameter, and predominantly in matrix support. Matrix composition varies from micrite-quartz wacke to nearly pure quartz arenite. The matrix and carbonate clasts occasionally contain echinoid and other bioclast fragments visible only in thin section. The remainder of the lower member is composed of white, yellow and red quartz arenites, quartz-chert arenite and wacke, with rare cross-bedded lenses approximately 1 m thick of chert pebble conglomerate. Conglomerate clasts are subangular to subrounded and 3 mm to 3 cm in diameter. Where bedding is observed these units are cross-beded. Quartz grains are moderately well-sorted. The largest quartz grains are round, while smaller grains were visibly fractured in place. Sand-size components in lower units are almost entirely quartz, with chert sand abundance increasing upsection.

The upper feldspathic member of the Dunlap Formation is 850 m thick in nappe III, east of C21. West of the fault a section 500 m thick is preserved. The member is characterized by the influx of fine-grained detrital feldspar and volcanogenic grains; a decrease in quartz abundance, roundness and grain size; the appearance of carbonate and mudstone clasts in conglomerate lenses; and an overall upsection decrease in grain size. In thin-section mudstones are calcareous, red-stained, feldspathic mud containing grains of chert, quartz, occasional feldspar, and small (on the order of 1 to 3 mm)
clasts of mudstone and altered volcanics. Microcline and sanidine feldspars were identified. Red and brown staining derives from altered iron silicate grains, which may also have provided authigenic brown clay in the matrix as well as minor calcite for cement and observed occasional secondary grain replacement. Such products of iron silicate decomposition have been observed in arid and warm fluvial to alluvial environments (Walker, 1967; Stanley, 1971). Conglomerates occur occasionally in 1- to 2-m thick trough-bedded lenses which grade rapidly both laterally and vertically into sandstone, wacke and siltstone. Two red mudstone samples from the upper part of this member, when examined in thin section, were found to be composed of very fine (less than 1 mm diameter) dolomite rhombs in dark red matrix. The uppermost 10 m of the member are red and green shaley mudstones interbedded with thin-bedded non-fossiliferous silty micritic shale.

The above-described units are correlated with the Dunlap Formation based on lithologic similarity. No fossils were found in the Northern Cedar Mountains. The only direct age constraint on this unit is that it was deposited and lithified prior to contractional deformation which in turn predated the emplacement of the Mesozoic quartz pluton. The pluton was not dated in this study, but similar plutons in the region were dated at 90 to 100 my (see discussion in the regional section of this paper). Correlation of this unit with the Dunlap Formation appears to be valid in light of its distinctive lithology and depositional style, and because the Dunlap Formation
is the only unit in the region with this lithology which was deposited prior to contractional deformation.

**Facies Relationships:** Facies relationships between nappes in the study area carry implications for the nature of the Luning Formation's depositional environment. Relationships between Luning facies in the Northern Cedar Mountains and those elsewhere in the region provide constraints for the significance of the absence of the Sunrise-Gabbs Formations in the study area.

Comparison of analogous Luning Formation members in nappes I, II and IV shows an overall thinning in bedding and an increase in the sharpness of contacts within and between members, as well as a north-to-south increase in conglomerate clast size in the middle clastic member. These factors indicate a more near-shore affinity for nappes I and II than for nappe IV, consistent with previous interpretations as to the shape of the depositional basin.

The upper carbonate member of the Luning Formation is of the same platform facies as its regionally uniform analogues elsewhere in the upper-plate assemblage. Where complete sequences are preserved, this unit is everywhere overlain by significant thicknesses of the Sunrise-Gabbs Formations (Muller and Ferguson, 1939; Ferguson and Muller, 1949; Silberling, 1959; Oldow, 1981a). The regional uniformity of both the Luning's upper carbonate member and the Sunrise-Gabbs Formations implies that Luning section in the Northern Cedar Mountains was overlain by the Sunrise-Gabbs prior to contraction. This implies that the Sunrise-Gabbs was either removed prior to deformation by erosion or tectonically obscured by faulting during
contraction. Evidence for the first case would be abundant Sunrise-
Gabbs detritus in the Dunlap Formation. This is absent in the study
area, which by its position between Luning nappes II and IV is
inferred to have been deposited at or near the same site as the
Luning in these nappes. The presence of preferred decoupling surfaces
near the lower and upper bounds of the Sunrise-Gabbs elsewhere in the
region provides further support for its having been tectonically
obscured in the Northern Cedar Mountains.

Rocks of the Gold Range Assemblage

The Dunlap Formation is the only unit of the Gold Range
assemblage which is exposed in the Northern Cedar Mountains. It
unconformably overlies the Permian (?) Mina Formation in the Southern
Cedar Mountains, 4 km to the south (Muller and Ferguson, 1939;
Ferguson and Muller, 1949; Wetterauer, 1977). A stratigraphic column
for the Gold Range assemblage in the Northern and Southern Cedar
Mountains is given in Figure 7.

Mina Formation: The Mina Formation was defined by Speed (1977b)
in his reappraisal of rocks originally assigned to the Excelsior
Formation (Muller and Ferguson, 1939; Ferguson and Muller, 1949). The
formation consists of bedded chert, volcanogenic turbidites, and
mudstones, of marginal basin affinity. It is found in lower-plate
exposures in the Excelsior Mountains, Pilot Mountains and Southern
Cedar Mountains. Faunal and K-Ar dating have constrained the Mina
Formation to be Permian (Speed, 1977b, 1979, and 1983) or Mississippian
in age (D.L. Jones, 1984, personal communication). The base of the
FIGURE 7. Northern and Southern Cedar Mountains, Gold Range assemblage stratigraphy.
formation is not exposed. Although it occurs only in the lower plate of the Luning thrust and is therefore paraautochthonous with respect to it, the Mina Formation is locally extensively thrust-faulted and folded and may itself be a part of the Sevier allochthon previously discussed.

In the Southern Cedar Mountains the Mina Formation consists of bedded chert and silty mudstone in its northernmost exposure, investigated during this study. Lithologic correlation with the distinctive bedded chert facies of the Mina elsewhere in the region identifies these rocks in the Southern Cedar Mountains. No date constraining the age of this unit were collected.

**Dunlap Formation:** The Jurassic-Cretaceous (?) Dunlap Formation occurs both in the lower and upper plates of the Luning thrust. Age constraints for the formation are discussed elsewhere in the "Stratigraphic Framework" section. Outside of the study area, where the base of the formation is observed, the Dunlap generally overlies the Permian Mina Formation with pronounced angular unconformity. Locally, in the westernmost Pilot Mountains the Dunlap disconformably overlies the Water Canyon assemblage which unconformably overlies the Mina.

The Dunlap is characterized by a pronounced spatial variability in thickness and composition both between and within mountain ranges, with the thickest and most coarse-grained lithologies exposed in the western Pilot Mountains, the formation's type locality. Regionally, the Dunlap of the lower plate, like its upper-plate equivalents, is
divided into two members by the lithology of sand-sized components (Wetterauer, 1977). The chief difference between upper- and lower-plate rocks constituting the Dunlap is in the composition, abundance, and clast size of conglomerates. In the lower plate, the lower member comprises basal locally-derived chert pebble, cobble and boulder conglomerate and breccia overlain by quartz arenite and sandstone and lenses of red mudstone. The upper member comprises chert breccia and red feldspathic wacke with limestone lenses. Sand-size and smaller conglomerates are less abundant and much smaller in clast size outside the Pilot Mountains. Feldspathic components increase in abundance toward the west, and andesitic volcanics are found intercalated in the upper member in westernmost localities, suggesting a westerly volcanic source (Wetterauer, 1977).

Approximately 100 m of the Dunlap Formation is exposed on the southern margin of the study area immediately below the Luning thrust. The lower contact is lost in alluvium and Tertiary volcanics. The lowest observed units are red cherty mudstone and siltstone with occasional thin unfossiliferous carbonate lenses. These are overlain by cherty red siltstone and mudstone interbedded with minor feldspathic chert-quartz wacke and sandstone, grading upward into approximately 50 m of pebble conglomerate interbedded with red chert + feldspathic wacke and mudstone. Conglomerates are composed of clasts of red mudstone, chert, and minor (up to 20%) volcanic clasts in red mudstone matrix. The uppermost units are chert pebble conglomerate in chert matrix. Clasts are subrounded and poorly sorted, from 2 cm to 10 cm in diameter. Through cross-bedding in conglomerate indicates
stratigraphic tops to be up. This section overall appears to be of upper-member lithology.

Wetterauer (1977, p. 154) described Dunlap rocks from the Southern Cedar Mountains as "very fine-grained red quartz sandstone; few floating chert fragments; red siltstone; sparse chert breccia; medium-grained quartz-chert sandstone; interbeds of limestone with some chert breccia layers near top of section." Wetterauer's section appears to be comprised of both members and no thickness was specified. The Dunlap Formation in the lower plate exposures in the study area appear correlative with the upper part of the section in the Southern Cedar Mountains 6 km to the south.

The age of the Dunlap Formation in the Gold Range assemblage is not constrained in the study area. It is identified on the basis of lithologic similarity to the type locality and by its stratigraphic position, unconformably overlying the Mina Formation.
STRUCTURE OF THE NORTHERN CEDAR MOUNTAINS

The rocks of the Luning assemblage comprise the bulk of pre-Tertiary exposures in the Northern Cedar Mountains, and are bound on the south by the east-west striking, north-dipping Luning thrust which separates the allochthon from the subjacent Gold Range assemblage. Five thrust nappes are recognized in the Luning assemblage and are separated by low-angle faults which trend east-west but, unlike the Luning thrust, dip to the south. The nappes are further divided into eastern and western domains by northerly-striking Cenozoic high-angle faults. The displacements on these high-angle faults result in exposure of different structural levels within the thrust nappes, and have been instrumental in deciphering the geometry of pre-Tertiary structures.

Cenozoic high-angle faults

The Northern Cedar Mountains are largely bounded to the east and west by Cenozoic high-angle faults, and the range itself is dissected by them. Abrupt topographic changes and dramatic juxtaposition of pre-Tertiary and Tertiary rocks along the western boundary indicate that major faults on this side of the range are down to the west. Cenozoic faults dissecting the range displace earlier structures and thus juxtapose different structural levels within them. One fault of particular interest in this study, here named Cz1 (see map, Plate 1), serves as the domainal boundary within the thrust nappes.

Cenozoic faults within and bounding the study area have variable orientations from nearly N-S to E-W. Major faults strike from N10E to N45W and dip from 60° to 90° both east and west. N- to NW-striking
faults near the western boundary of the range have predominantly steep westerly dips. Slickensides measured on these faults showed predominantly right-oblique motions (see Figure 8). Studies of Cenozoic faults at and west of the western boundary of the range (Gianella and Callaghan, 1934) characterized three groups of faults, trending respectively approximately N11E, N61E, and N11W. These faults showed predominantly strike-slip right-lateral motion.

Timing of Cenozoic faulting is not well constrained in the study area. Tertiary igneous rocks (basalts, andesites and rhyolite tuffs -- not differentiated in this study) are, in some locations, cut by these faults, while in other locations they have utilized fault surfaces during deposition. Volcanics forming the northern boundary of the study area and truncated to the west by high-angle faults have been identified as the Singatse Tuff, dated as 27 my (Ekren et al., 1980). Tertiary intrusives at least locally postdate activity on the Cz1 fault as they surface along its southern trace.

The known timing of phases of Cenozoic faulting in the surrounding area has implications for that of faults in the Northern Cedar Mountains. In the Bettles Well valley between the Pilot Mountains and the Gabbs Valley Range, approximately 10 km west of the study area, N-trending faulting began less than 17 my bp, in all likelihood between 10 and 7 my (Meinwald, 1981; Oldow and Meinwald, in prep.; Oldow, 1984, personal communication). Faults in the Stewart Valley west of the study area were active as late as 1932 (Gianella and Callaghan, 1934).

The Cz1 fault is described because of its importance in deciphering Mesozoic structures in the study area. The fault strikes nearly north-
FIGURE 8. Lower hemisphere equal-area projections of measurements of slickensides on Cenozoic high-angle fault surfaces. Arrows indicate lateral sense of latest motion on fault surfaces, where these were measurable. "D" and "U" indicate sense of vertical latest motion on faults: "D" is down, and "U" is up.
south. No fault surfaces were directly measurable in the field, but the fault's insensitivity to topography indicates steep to vertical dips. Displacement of Mesozoic depositional contacts can be used to deduce motion on the fault. The south-dipping contact between the Luning Formation's middle clastic and shaley members in nappe IV is displaced to the north on the east side of Cz1. In nappe II, the contact between these two members dips north, and is displaced to the south on the east side of Cz1. Predominant strike-slip motion would displace both contacts in the same direction. Observations therefore indicate that motion on Cz1 was predominantly dip-slip, east side down, with a minimum throw of 350 m in the area described.

**Upper plate structures**

Beds within the Luning assemblage are deformed in one and locally two superimposed fold sets designated D1 and D2 in order of decreasing age. First-phase structures consist of gentle (1b to 1c, Ramsay classification) to tight (1c) major (greater than 30 m half-wavelength) and minor (less than 30 m half-wavelength) folds, with associated penetrative axial plane cleavage. First folds are upright to overturned, predominantly to the north, and have axial planes which generally strike NE but range $140^\circ$ from due east to N50W. Second phase structures consist of locally developed, gentle to tight, minor folds with associated axial plane cleavage, generally approximately perpendicular to D1. Second folds have axial planes which strike predominantly N50W but range from N5W to N78W, and are upright to overturned as much as $40^\circ$ in either direction.
Minor folds of both phases were found only in the carbonates of the Luning Formation, which were better preserved than the clastics in the Luning and the Dunlap formations. In the fine-grained clastics of the Dunlap, D1 produces penetrative cleavage which was axial planar to major first folds. A spaced, locally penetrative cleavage in these clastics crosscuts D1 cleavage and is consistent in orientation with D2 structures in carbonate units. Major and minor first fold axial planes strike approximately parallel to low-angle faults, and minor D1 fold axial planes proximal to these faults dip consistently (although more steeply) to the south. Second-phase minor folds have axial planes at high angles to these faults. Minor folds of both phases increase in occurrence and limb oppression with increasing proximity to low-angle faults.

**Geometric analysis:** Lower hemisphere equal-area projections of D1 fabric data are summarized in Figure 9. Planar features (axial planes and axial plane cleavage) have predominant NE strikes, with most between N10E and N65E. They dip moderately to steeply both to the north and south. Variations are systematic and related to reorientation by subsequent deformations and to spatial relationships to low-angle faults. Folds in the upper plates of nappe-bounding low-angle faults are increasingly overturned to the north as the fault contact is approached. Poles to axial planes of these folds form a partial great circle girdle consistent with rotation about an ENE (approximately N70E) axial plane and are thought to be the product of shear associated with thrust displacement.

D1 lineations include fold axes for D1 folds and bedding-cleavage intersections in clastic rocks. Fold axes are doubly plunging with
maxima at approximately N65E, 30° and S80W, 30°. Bedding-cleavage intersections are more variable but may be grouped into two sets. One set is NE-plunging 10° to 20°, with trends varying from N22E to N74E, while the other plunges SW, 20° to 50°, with trends from S20W to S64W.

Second-phase fabric data is summarized in Figure 10. Axial planes strike from N10W to N80W with a maximum at approximately N50W, and dip from 20° to vertically to both north and south. Spaced cleavage strikes generally NW but is variable over 100°, and dips 40° to vertically to both north and south. D2 lineations (fold axes and bedding-cleavage intersections) also exhibit great orientational variability. Second-phase, generally NW-trending structures elsewhere in the region exhibit primary variability (Oldow, 1981b), and this may also be the case in the study area.

Poles to D1 cleavage, shown in Figure 11, are generally in great circle distributions. Great-circle arc traces and poles for individual localities are also shown in Figure 11. Poles to great-circle distributions plunge 60° to 90° to the north and northwest.

D1/D2 cleavage intersections in clastics produce pencil structures. Pencil structures have been found to be compaction features in areas where one "foliation" is parallel to bedding (Kligfield et al., 1983), but in the study area both foliations are at moderate to high angles to bedding where measurements of all three were possible. Pencil structures in the study area are, therefore, inferred to be tectonic in origin. Orientations for these and other D1/D2 lineations are shown in Figure 12. Pencil lineations in Dunlap clastic rocks trend NE to SE and plunge 30° to 70°. Other D1/D2 intersections trend NE to SE to SW and plunge
FIGURE 11. Lower hemisphere equal-area projections of D1/D2 crosscutting relationships in structures where D1 cleavage was folded about a D2 axis. D1 cleavage is distributed in great circles with D2 cleavage which is axial planar to D1 (great circles shown as dashed arc traces).
FIGURE 12. Lower hemisphere equal-area projections of D1 x D2 lineations in the Luning assemblage.
moderately to steeply.

**Low-angle faults in the Luning assemblage:** Three WSW-ENE trending, south-dipping faults are mapped in the upper plate assemblage. From south to north, they are: the South thrust, which forms the northern boundary of nappe I; the Cedar Mountain thrust, north of nappe II; and the Dunlap thrust, north of nappe III. Fold-fault relations and older-over-younger relationships for the South and Cedar Mountain thrusts are typical of thrust faulting. As explained later, the Dunlap fault is also interpreted as a thrust in spite of the younger-over-older relationship it produces.

The shallowly south-dipping **South thrust** is located approximately 1 km north of the Luning thrust. It is truncated to the east and west by Cenozoic high-angle faults and is not observed east of Cz1, where its inferred trace is obscured by Tertiary volcanics. The upper plate of the thrust is an ENE-trending first-phase anticline which is strongly overturned to the north. Portions of the upper plate are locally obscured by Tertiary volcanics. At the contact, the lower plate of the thrust is comprised of moderately to steeply north-dipping units of the Luning's middle shaley carbonate member, stratigraphically continuous with the Luning upper carbonate member to the north.

Minor D1 folds in the upper plate increase in limb oppression and degree of northward recumbency as the thrust contact is approached. D2 deformation produced only gentle flexures resulting in doubly-plunging D1 fold axes.

The parallelism of D1 axial planes in the upper plate with the fault surface, localization of D1 structures along the thrust trace, and the
fact that it cuts a major D1 fold, suggest that the South thrust formed
during D1 deformation. The lack of localized D2 structures renders it
uncertain whether or not the South thrust was reactivated during D2.
The vergence of D1 minor and major folds and the orientation of the
thrust surface indicate that relative motion on the upper plate of the
South thrust was updip and to the north.

The Cedar Mountain thrust is truncated to east and west by range-
bounding Tertiary faults. Within the range, it is displaced at several
points by Tertiary faults that result in the exposure of differing
structural levels. The effect of Cz1 in exposing different structural
levels to east and west is of particular importance in later structural
interpretations.

The leading edge of the thrust's upper plate is composed of the
upper carbonate member of the Luning Formation. Stratigraphic tops are
to the north. Beds which dip moderately to the north become steep to
overturned as the fault trace is approached from the south. Minor D1
and D2 folds are localized along the fault trace. Upper carbonate
member section is cut out increasingly from east to west. East-side
down motion on Cz1 exposes higher structural levels on the east.

Several exposures of dolomitized carbonate breccia truncate lower
plate (nappe III) bedding to the north on subhorizontal ground surfaces
east of Cz1. These are inferred to be klippen of the upper plate of the
Cedar Mountain thrust.

The lower plate (nappe III) north of the thrust is composed of
Dunlap Formation in which stratigraphic tops, indicated by tangential
crossbeds and truncations in trough crossbedding, are to the south.
Beds are moderately to steeply south-dipping, locally overturned in western exposures.

Measurements of the exposed fault surface indicate that at the far west it dips approximately 40°S. Displacement along Cz1 indicates approximately 60° southerly dips in the central part of the fault trace. The subhorizontal contact between nappe III and overlying nappe II klippen east of Cz1 indicates that the Cedar Mountain thrust shallows rapidly upsection to the north.

Minor D1 and D2 folds are localized in the upper plate near the fault trace, and are tighter and more overturned to the north as the contact is approached. D1 axial planes are subparallel to the thrust near the contact, and the thrust cuts minor first-phase folds. Exposures were not good enough to clarify whether D2 folds deform the fault surface or are cut by it. The localized nature of D1 and D2 folds along the thrust trace suggests that the thrust was active during both phases, and thus formed during D1 deformation. As with the South thrust, all indications are that motion on the Cedar Mountain thrust was updip and to the north.

The Dunlap thrust trace, approximately parallel to and north of the Cedar Mountain thrust, is traced across the range, and is truncated to east and west by range-bounding high-angle faults. Its upper plate is the southward-younging Dunlap Formation (nappe III) which forms the lower plate of the Cedar Mountain thrust. The Dunlap is juxtaposed over the likewise southward-younging Luning Formation of nappe IV. Bedding planes in the lower plate, like those in the upper plate, dip steeply to the south (see Figure 15 for bedding plane orientations from each nappe).
The large north-south offset of the fault trace (Plate 1) by displacement along Cz1 indicates that the fault dips to the south more shallowly than the Cedar Mountain thrust. The Dunlap thrust truncates upper-plate bedding at a steep angle, as shown by the fact that 500 m of the lower member and 300 m of the upper member are cut out in the lower structural levels exposed west of Cz1.

Unlike the Cedar Mountain and South thrusts, both of which have classic thrust fault upper/lower plate relations and geometries, the Dunlap thrust is characterized by the younger-over-older relationship of its upper and lower plates and thus its designation as a thrust is suspect. The interpretation of the Dunlap thrust is discussed later in this chapter. The Dunlap thrust has a shallow southerly dip and presumably intersects the Cedar Mountain thrust at depth.

Nappe III constitutes the limb of a major D1 fold. Minor D1 folds, perhaps obscured by the poor state of preservation of the fine-grained Dunlap Formation clastics which constitute the bulk of nappe III, were not found in this area. However, first-phase cleavage is pervasive in these rocks and is predominantly south-dipping. Stratigraphic "tops" indicators suggest that nappe III constitutes part of the southern limb of a major first-phase anticline. Bedding-cleavage intersections of $20^\circ-40^\circ$ indicate that this unit may represent a portion of the limb proximal to the hinge. Second-phase cleavage is strongly north-dipping in eastern localities, less strongly so to the west.

Direct evidence for the timing and nature of Dunlap faulting is poorly constrained. Indications that it may have formed during D1 are: its parallelism with thrust faults, especially the Cedar Mountain thrust;
and its coplanar relationship with D1 axial surfaces. D1 cleavage in 
the Dunlap Formation composing the fault's upper plate strikes parallel 
to the fault trace and dips predominantly to the south, as does the 
fault surface. However the fault cuts downsection, juxtaposing younger 
over older rocks. The possibility of extensional motion on this fault 
will be discussed in a later section. It is at least inferrable that 
this fault acted as a thrust for some part of its history, and that it 
is most likely to have done so during D1 deformation.

Lower plate structures

Rocks lithologically correlated with the Dunlap Formation comprise 
the Gold Range assemblage in the study area. No minor folds were 
observed in this unit, but D1 and D2 cleavages are identified in fine-
grained clastic rocks. Bedding planes are subhorizontal to moderately 
dipping (Figure 13). Major folds are open flexures with both 
approximately north-south and east-west striking axial planes. Bedding 
near the Luning thrust trace strikes approximately N80W to E-W and dips 
10° to 16° to the north. Topography follows bedding surfaces, sloping 
up and shallowing as one moves southward from the thrust trace.

Two sets of subvertical fractures occur in chert conglomerates. 
They are approximately perpendicular to one another and to bedding. 
Both sets have approximately 1 cm spacing. The two orientations are 
approximately coincident with the strikes of D1 and D2 cleavages in 
lower-plate fine-grained clastics (Figure 14). One set is NE-trending 
(approximately N28E), the other NW-trending (approximately N62W). No 
evidence of dissolution along these surfaces was observed.
FIGURE 14. Lower hemisphere equal-area projections of fabric data in the Gold Range assemblage, comparing D1 and D2 cleavages with orthogonal fracture sets measured in chert conglomerate. "First" fractures are demonstrably crosscut by "second" fractures.
FIGURE 15 (on following pages). Lower hemisphere equal-area projections of bedding plane measurements in the five nappes comprising the Luning assemblage in the Northern Cedar Mountains.
FIGURE 15a. Lower hemisphere equal area projections of bedding plane measurements in nappe I of the Luning assemblage.
FIGURE 15b. Lower hemisphere equal-area projections of bedding plane measurements in nappe II of the Luning assemblage.
FIGURE 15c. Lower hemisphere equal-area projections of bedding plane measurements in nappe III of the Luning assemblage.
FIGURE 15d. Lower hemisphere equal-area projections of poles to bedding planes in nappe IV of the Luning assemblage.
FIGURE 15e. Lower hemisphere equal-area projections of poles to bedding planes in nappe V of the Luning assemblage.
The origin of the fractures is equivocal. Spacing and orientation imply that these are two spaced cleavages, but chert pebble conglomerate is an unusual substrate for such structures. Paired perpendicular, subvertical joint sets have been observed in foreland fold and thrust belts where one set was parallel to tectonic transport (Julian, 1983). There they are inferred possibly to be effects of post-thrusting relaxation. Whatever the origin of the fractures in the study area, they appear to be related by orientation to D1 and D2 structures.

**Luning thrust**

The east-west striking, north-dipping Luning thrust crosses the southeasternmost portion of the study area. Upper-plate rocks along the contact are comprised of a thin belt of poorly preserved carbonates, faunally correlated with the Luning Formation. The contact relationship between the carbonates and the Luning Formation to the north and east is obscured by Tertiary volcanics. Carbonate rocks are recrystallized and highly deformed as indicated by flattened fossils, in a crude foliation. The thrust fault is not observed west of the inferred trace of high-angle fault Cz1, but is thought to be obscured by Tertiary volcanics south of a small exposure of rocks of the lower and middle members of the Luning Formation.

This fault is correlated with the Luning thrust in the region because it forms the contact between Luning and Gold Range assemblage rocks in the study area. In the Southern Cedar Mountains a few kilometers to the south, the only pre-Tertiary units present are the Mina Formation, overlain unconformably with Dunlap Formation which appears to have affinity with the Gold Range assemblage Dunlap in the
study area.

Direct evidence for the timing of Luning thrusting is lacking because of the poor preservation of rocks at the contact. Emplacement during D1 deformation is inferred for several reasons. Elsewhere in the study area, D1 deformation produces tighter folds and better-developed cleavage than D2, and D1 structures appear to be associated with the emplacement of the other thrusts. Elsewhere in the region, the Luning thrust was emplaced during first-phase deformation producing structures oriented similarly to those of D1 in the study area (Oldow, 1981a, 1981b and 1984a). Reactivation during D2 deformation is not constrained by evidence in the study area but is reported elsewhere (Oldow, 1981a).

Transport direction on the Luning thrust is inferred from the orientation of the thrust surface and by stratigraphic relations with the Southern Cedar Mountains. Bedding orientations and topography in the underlying Gold Range assemblage indicate that the Luning thrust dips northward. The stratigraphic omission of the Luning Formation in the Southern Cedar Mountains precludes the possibility that the Luning Formation of the upper plate had its origin south of the Luning thrust in the Northern Cedar Mountains. These two factors suggest that transport on the Luning thrust in the Northern Cedar Mountains had a significant southward component. Studies elsewhere in the region (Oldow, 1981a, 1981b and 1984a) determine a tectonic transport in a southeasterly direction.

Deformation timing relations

Timing relations are not directly constrained in the Northern Cedar
Mountains. D1 and D2 deformations postdate deposition and lithification of the Dunlap Formation, elsewhere dated as Jurassic-Cretaceous (?) in age, and predate the intrusion of the quartz monzonite pluton which forms the central portion of the range. The pluton is composed of quartz-bearing monzonite with peripheral trachyte and latite porphyries. In its southern exposure the pluton is overlain by volcanics comprised of welded tuffs, ranging in composition from latite to quartz-trachyte to rhyolite/rhyodacite, containing occasional sedimentary rock (chert siltstone) xenoliths. The plutonic and volcanic rocks are interfingered at their contact, and cooling rims in both indicate that the volcanics were extruded contemporarily with the emplacement of the pluton. Lithologies of both the pluton and the volcanic unit are described in more detail in the appendix. Crosscutting plutons elsewhere in the region, of similar composition, are dated as 90-100 my bp (Speed and Kistler, 1980; Oldow, 1984a).

**Backthrusting in the Northern Cedar Mountains**

The Luning thrust's upper plate includes the Cedar Mountain and South thrusts. Although relative transport on the Cedar Mountain and South thrusts was to the north, stratigraphic omission in the Southern Cedar Mountains of the Luning assemblage rocks in their upper plates precludes absolute northward transport within the overall southward-directed transport regime of Luning thrusting.

Various models have been proposed for the inception of backthrusting, but all are related to proximity to a ramping forward-moving thrust (Butler, 1982). Backthrusting of this type in all models involves only relative transport over small distances. All models propose that
backthrusting occurs to accommodate the bending of units travelling up a steepening or ramping thrust decollement. Proximity of the Cedar Mountain and South thrusts to the Luning thrust provides a likely setting for this cause-effect relationship. A proposal for the genesis and evolution of these faults is discussed in the forward model section of this paper.

**Motion on the Dunlap thrust**

Since direct evidence for motion on the Dunlap thrust is inconclusive and its configuration renders it suspect as an extensional fault, it becomes necessary to consider all models which appear consistent with its present geometry.

The following is a summary of direct evidence about the present geometry of the Dunlap fault. The fault strikes approximately E-W to ENE and, based on reconstructions made possible by displacement along Cenozoic faults, dips 20° to 25° to the south. First-phase penetrative cleavage in nappe III, the fault's upper plate, shows a preference for steep southward dips. Spaced D2 cleavage dips both north and south. No minor folds were found in nappe III. The fault cuts downsection, producing a younger (Dunlap Formation, nappe III) over older (Luning Formation, nappe IV) relationship. The Gabbs and Sunrise formations, which lie stratigraphically between the Luning and Dunlap formations where depositional contacts are preserved in the region, are not found in the area under discussion. Bedding planes in both nappes dip 45° to 75° to the south. The shallowly-dipping fault thus cuts bedding at a high angle (approximately 50° for nappe III) and also cuts major first-phase folds.
Based on its geometry, three possibilities are suggested for the development of the Dunlap fault. The models discussed are: that it developed as a low-angle normal fault; as a high-angle normal fault reoriented to its present position by folding; and finally as a backthrust. The test for the possible validity of each model is the necessary pre-fault configurations of the units involved and the proto-fault surface. Discussion in this chapter forms the framework for final testing of these models in the chapter on forward modeling.

If the Dunlap fault is a low-angle normal fault, faulting must postdate D1 because the structure truncates major first-phase folds. The implication is that there was a southerly component of motion for the fault's upper plate. Although bedding attitudes and southward-younging stratigraphies in nappes III and IV are consistent with the possibility that these units originated in the same major fold limb, extensional (i.e. southerly) motion on the fault would not juxtapose units from the same fold in the geometry observed, which truncates Luning Formation and obscures the Sunrise-Gabbs formations which presumably intervene stratigraphically between the Luning and Dunlap Formation. The low-angle normal fault case thus requires that nappes III and IV originated in different folds, and that the upper plate of the Dunlap fault underwent southerly transport of a distance which was dependent on intervening structures but at a minimum is several kilometers. Figure 16 shows a cross-section of a hypothetical "fold set" and fault surface which illustrates the difficulties associated with modelling the Dunlap fault as a low-angle normal fault.

It is important to note that, since the Cedar Mountain thrust formed
FIGURE 16. Hypothetical reconstruction of the Dunlap fault as a low-angle normal fault. Figure 16a shows present-day geometry (bedding plane and fault attitudes) as derived from field mapping. Box ABCD is dissected by the Dunlap fault. Figure 16b shows the pre-fault positions of the upper- and lower-plate components of ABCD as necessitated under the low-angle normal fault model.
during first-phase deformation, the low-angle normal fault case requires the emplacement of the Cedar Mountain thrust prior to Dunlap faulting. As backthrusting does not generally involve transport along great distances (Butler, 1982), it is inferrable that the Dunlap and Luning sections juxtaposed by the Cedar Mountain thrust were deposited in or near the same site. This fact will become important in the forward model section of this paper, where further discussion of this alternative will show that the low-angle normal fault is not feasible in light of its relationship to other structures in the study area.

The folded high-angle normal fault case requires a two-stage development which is illustrated schematically in Figure 17. In the first stage a high-angle, planar, north-dipping, E-W striking fault forms at approximately 50° (as derived from present fault-bedding plane intersections) to horizontal bedding sometime before contractional deformation. Present juxtaposition of units requires a minimum throw on this fault of 1.5 km down to the north. Subsequent D1 and D2 deformation folds bedding and the fault, eventually overturning it to its present orientation.

This case calls for a period of extension forming E-W trending faults prior to D1. No structures representing such a phase have been observed in the region. Although extensional high-angle faults have been observed, they are NW to NNW-trending (Bartel and Oldow, 1984). It is also difficult to see why such a fault, oriented advantageously for reactivation in D1 contraction, would not become a thrust instead of folding in place as required by the model. If reactivation did occur, this fault would rapidly produce older-over-younger relationships unless
FIGURE 17. Hypothetical reconstruction of the Dunlap fault as a folded high-angle normal fault. Figure 17a shows extensional motion on the fault, detailed in box ABCD. Figure 17b shows folding of the fault to its present position as derived from surface data. Note that this case requires folding of a fault ideally oriented for reactivation during Luning thrusting.
the original normal fault slip was much greater than 1.5 km. Arguments
presented in the forward model section will rule out this case.

If the Dunlap thrust formed as a backthrust, surface data indicate
that it formed during D1 deformation, after the folding that tilted the
Luning section in its lower plate. The southerly vergence of D2
structures in nappe III argues against its reactivation during D2. As
previously discussed, relations between upper and lower plate rocks
indicate that both are from analogous limbs of different folds or from
the same fold. The northerly transport required in this case would
produce the observed relationships with a minimum amount of shortening,
by means of telescoping a single fold limb. A hypothetical "fold set"
illustrating the reconstruction of this case is shown in Figure 18.

Although the younger-over-older, downcutting fault relationship may
be used as an indicator of extensional faulting in terranes which have
not been folded, such relationships may readily be produced by
backthrusting of previously folded terranes. The Cedar Mountain
backthrust illustrates this within the study area. Although it
juxtaposes older over younger section, the backthrust cuts downsection
over the folded Dunlap Formation of nappe III. Hypothetical further
northward movement on the Cedar Mountain backthrust would cut further
downsection through progressively older units and easily produce younger-
over-older relationships. This can readily be imagined by treating
nappes III and IV as one limb of a major fold and projecting motion of
nappe II northward until the contact truncates middle-member Luning
Formation in nappe IV.
FIGURE 18. Hypothetical reconstruction of the Dunlap fault as a backthrust. Figure 18a shows present-day geometry as derived from surface data. Figure 18b shows the pre-fault positions of the components of box ABCD as required by this model. Comparison with figure 16 illustrates the relative simplicity of this model.
A balanced cross-section of the study area was prepared on the basis of surface data and regional stratigraphy. This approach was used to test the feasibility of various models which initially appeared to be consistent with surface data, with the objective of producing a forward kinematic model or models of deformation in which total shortening was minimized. The balanced cross-section and model were constructed along A-A' (Plate 1). An additional cross-section (B-B', Plate 1), exposing different structural levels, was used in conjunction with A-A'. Cross-sections along A-A' and B-B' were combined to produce a shallow-surface cross-section which provided important constraints for the model.

**Constraints for the balanced cross-section**

**The Rocky Mountain model:** The Rocky Mountain model of deformation in fold and thrust belts (Bally et al., 1966; Dahlstrom, 1969) was applied in order to simplify the construction of the balanced cross-section. The Rocky Mountain model assumes plane strain deformation by concentric (flexural slip) folding and by faulting in a previously lithified section. It thus allows conservation of bed length and thickness as well as conservation of area in the deformed and restored cross-sections.

Use of the Rocky Mountain model implies the following conditions:

-- the absence of primary thickness changes in units in the restored section;

-- that units behave competently at the pressure and temperature conditions of deformation; and
-- that the area be involved in only a single phase of deformation (in order to justify the assumption of plane strain).

Previous discussion has established that these conditions do not strictly apply in the study area. In view of the unique usefulness of the technique in providing an evaluative framework for proposed models, balanced cross-sections of the Northern Cedar Mountains were nevertheless prepared using the Rocky Mountain model. The results were then evaluated with respect to actual conditions in the study area to see if departure from the ideal boundary conditions could invalidate the model.

**Minimizing estimates of displacement:** As evidence for the actual amount of shortening in fold and thrust belts is generally not retrievable, the only dependable estimates which may be made are for the smallest possible displacements which could have produced the structures observed. Assumptions inherent to the Rocky Mountain model (Bally et al., 1966; Dahlstrom, 1969) listed above, work to minimize estimates of displacement. Use of this model minimizes shortening because it assumes no bed-length shortening and no volume loss (reduction in area of the cross-section) during deformation. Thus the restored cross-section shows only the minimum length of section possible.

Conservation of displacement estimates also determined many of the choices made during modelling. Surface data sometimes required that the stratigraphic section be doubled by faulting. In these cases subsurface positions of ramps and decoupling surfaces were chosen to minimize the length of doubled section. Models were also consistently prepared to minimize the distance to the inferred depositional site of allochthonous units. The stratigraphic omission of the Luning Formation in the
Southern Cedar Mountains 6 km south of the study area, where Dunlap Formation unconformably overlies the Mina Formation, makes it impossible for the Luning Formation in the upper plate (which has a minimum thickness of 2.2 km) to have been deposited in the Northern Cedar Mountains beneath the lower-plate Dunlap exposure in the study area. However, to minimize estimates of displacement, the upper plate was removed in the restored section only as far north as was necessary to bring upper-plate strata to a horizontal depositional orientation. Southward extrapolation of these units to the modelled unconformity at the top of the Mina Formation produces improbable onlap configurations but minimizes displacement estimates. Finally, although previous discussion has shown evidence that the Gold Range assemblage was at least locally displaced by Luning thrusting, it was assumed for the purposes of modelling that it was autochthonous. Possible displacements of the assemblage in the study area were not constrained by surface data and only serve to increase shortening estimates. This final assumption also provides a fixed point in cross-sections at the Luning thrust trace.

Surface data and the principle of minimizing displacements allow the establishment of two "fixed points" in the parautochthon which provide important constraints for the cross-section. These two points are the intersections of the sole thrust ramp and the Paleozoic/Mesozoic boundary with the present-day surface. These points remain fixed in all phases of the forward model. The modelled surface intersection of the sole thrust ramp is the Luning thrust trace in the study area. This is the southernmost point at which the sole thrust ramp could intersect the present-day surface. The basal detachment surface in the Luning
thrust system is likely to be the Paleozoic/Mesozoic boundary, as previously discussed. In the Northern Cedar Mountains the basement is inferred to be the Permian Mina Formation. The point where the basement unconformity intersects the surface is constrained within narrow limits by the northernmost exposure of Mina Formation in the Southern Cedar Mountains. Here the top of the Mina Formation is not preserved, the younger units having been eroded away. The northernmost surface trace of the unconformity, lost beneath Tertiary volcanics, can therefore be no further south, although it could be north of this exposure, which was chosen to represent the surface expression of the unconformity in the model.

Forward model

The kinematic forward model presented in this section (Plate 2) developed as a consequence of preparation of the balanced cross-section. Effects of each stage of the deformation were removed in reverse chronological order of occurrence. Each removal was accomplished by balancing the partially restored section to its deformed equivalent, and was constrained by surface data, the considerations discussed previously in this section, and by established models for deformation in other fold and thrust belts. The resultant cross-sections, when viewed in forward time sequence, form the kinematic model presented here.

Methods used in modelling: Balanced cross-sections are generally constructed using the configuration of the thrust belt's lower plate as a fixed framework upon which to model the deformation in the allochthon (Dahlstrom, 1969). As direct information about the subsurface is
unavailable for the study area, two important surfaces in the lower plate, the basement surface and the sole thrust ramp, were constructed based on the requirements for allochthon deformation in each model tested during this study. These surfaces, represented by the two "fixed points" on the Luning thrust trace and the Paleozoic/Mesozoic boundary, were drawn to intersect these points and to dip northward at the minimum angles required by the model for the deformed allochthon. Minimum angles were chosen because of data which indicate that basement and ramp angles in foreland thrust belts are usually shallow, the former on the order of 5 to $10^\circ$ or less, and the latter rarely up to $30^\circ$ (Bally et al., 1966; Oldow, 1984, personal communication). The first test for each model, thus, became the required configuration for the lower plate. Models which required angles greatly in excess of those usual for fold and thrust belts were discarded in favor of those which conformed to such constraints.

Time-sequencing of events represented by steps in the forward model was deduced from surface data when possible and otherwise inferred from observations and models in other fold and thrust belts. Generally in other fold and thrust belts where timing constraints exist, such as the Sevier and Appenines, sequential development proceeds from hinterland to foreland (Bally et al., 1966; Eliott, 1982 and 1983). This principle was used in model sequencing of forward-directed thrusting in this study. Backthrusts were modelled as late-stage accommodations of forward-directed thrusting at the Luning sole thrust ramp, as discussed in the structure section of this paper.

Decoupling surfaces used in modelling were those indicated by
structural and stratigraphic relations in the region. Consistent decoupling levels are observed: at the base of and within the lower member of the Luning Formation, the allochthon's lower boundary; at or near the base of the combined Sunrise-Gabbs formations; and at or near the base of the Dunlap Formation (see Figure 3 and discussion in the stratigraphy section). A decoupling surface is inferred at the top of preserved Dunlap section. Fabrics within the formation preserved in nappe III indicate that it was buried and lithified prior to, and buried during, emplacement of the Luning allochthon. Since overlying units are nowhere preserved, it is inferred that a decoupling surface formed at or above the upper boundary of exposed section.

The model employs uniform thicknesses for all upper-plate units except the lower carbonate member of the Luning Formation, and includes extra thicknesses of units not preserved in the study area but inferred to have been present during deformation. The principle of uniform thickness puts the maximum number of constraints on the cross-section in that variations may not be introduced in order to accommodate the model. Extra section thicknesses were added not only to conform to regional relations but also to insure that the validity of the model was not limited by accounting only for preserved section. Modelled thicknesses and the data used in arriving at them are summarized in Figure 19. Preserved thicknesses of upper-plate formations in the study area total approximately 3.6 km. The thickness of the upper plate used in modelling was 5 km.

Measured thicknesses of the upper two of the four members of the Luning Formation in the study area were comparable between nappes I, II
THICKNESS ESTIMATES USED IN PREPARATION OF BALANCED CROSS-SECTION AND FORWARD MODEL (in meters)

<table>
<thead>
<tr>
<th>FORMATION</th>
<th>Regional Variation</th>
<th>Nappes II &amp; III</th>
<th>Nappe IV</th>
<th>Assumed for model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dunlap</td>
<td>200-2530(^1)</td>
<td>/1370/</td>
<td>--</td>
<td>2400*</td>
</tr>
<tr>
<td>Sunrise-Gabbs</td>
<td>300-490(^2)</td>
<td>--</td>
<td>--</td>
<td>400</td>
</tr>
<tr>
<td>Luning:</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tr(l_u)</td>
<td>300-750(^3)</td>
<td>375/</td>
<td>/377/</td>
<td>500</td>
</tr>
<tr>
<td>Tr(l_ms)</td>
<td>900(^3)</td>
<td>825</td>
<td>790-825/</td>
<td>825</td>
</tr>
<tr>
<td>Tr(l_mc)</td>
<td>360(^3)</td>
<td>18(^i)</td>
<td>525</td>
<td>525</td>
</tr>
<tr>
<td>Tr(l_l)</td>
<td>/540(^3)</td>
<td>/50</td>
<td>1(^{500}) shoals with modeled basement geometry</td>
<td></td>
</tr>
</tbody>
</table>

KEY:
/ indicates thrust contact. To left = basal fault; to right = upper contact fault.
\(i\) indicates that contact is obscured by subsequent emplacement/deposition of igneous rocks. To left = basal contact; to right = upper contact.

*estimate includes the unit labelled "KJd?" in model

Sources for regional data:
1Wetterauer, 1977
2Taylor et al., 1983; Silberling, 1959; Ferguson & Muller, 1949
3Oldow, 1981

FIGURE 19. Thickness estimates used in preparation of the balanced cross-section and forward model (see Plate 2).
and IV, where the most complete sections were preserved. Only one complete section of the middle clastic member is preserved, in nappe IV. Lithologic comparisons do not indicate important facies changes between nappes for any member of the Luning. These observations allow and provide possible support for the assumption of uniform thickness used in modelling.

As only one section of the Dunlap Formation, fault bounded at both upper and lower contacts, is preserved in the study area, no inferences as to possible thickness variations may be made from direct evidence in the study area. Regional relations indicate marked lateral variations in lower-plate thicknesses of this unit, but lesser variability in upper-plate sections. Upper-plate Dunlap in the southern Gabbs Valley Range is approximately 1500 m thick (Ferguson and Muller, 1949), comparable to the 1400 m thickness mapped in the Northern Cedar Mountains. Lower and upper member thicknesses also appear to be comparable between the two ranges (Ferguson and Muller, 1949; Wetterauer, 1977). Syntectonic deposition of the Dunlap in a subsiding extensional basin has been demonstrated in the Gold Range assemblage in the Pilot Mountains (Bartel and Oldow, 1984), where the formation exhibits marked east-west lateral variability in thickness. Regional relations thus indicate the possibility of marked spatial thickness variability for the Dunlap Formation, certainly in the lower plate and quite possibly in the upper plate as well.

Indirect evidence for Dunlap thickness in the study area is preserved in the structures of the Luning Formation's upper carbonate member, which indicate that the Luning was buried at some depth during deformation.
Although development of structures varies systematically with proximity to thrust faults, uniformity in structural style between nappes indicates that no great differences existed in depth of burial at the time of deformation. This implies that, whatever the thickness of Sunrise-Gabbs and Dunlap formations overlying one nappe, a similar thickness probably overlaid the other.

In summary, regional and local relations neither indicate nor preclude that the upper-plate Dunlap Formation was uniform in thickness over the depositional area represented by the Northern Cedar Mountains. The model, prepared using the assumption of uniform thickness of this unit, will be evaluated to determine whether variability in this parameter could invalidate or substantially change its results.

Additional thicknesses were added to the upper members of the Luning Formation and the Dunlap Formation, both of which are everywhere truncated by thrust faults, and an extra thickness was also added to represent the undifferentiated Gabbs and Sunrise Formations. Upper-member Luning section is thicker by approximately 200 m both in the Shoshone Mountains and in the Pilot Mountains (Silberling, 1959; Oldow, 1981) than in the study area. In accordance with regional thicknesses, an additional 125 m thickness was added to the model. Preserved thicknesses of the Sunrise-Gabbs formations vary in the region from 300 to 500 m. A thickness of 400 m was chosen for use in the model. Fabrics in the Dunlap Formation indicate that it was buried at some depth during deformation. A thickness of 1000 m was added to account for missing Dunlap and presumably younger units present at the time of deformation.
Cryptic faults and other major structures: The modelling process revealed that observed surface structures could not account for the present geometry of the study area. A cryptic low-angle normal fault, here called the LNF fault, was required in modelling subsurface relations on the Luning thrust ramp. "Balancing" of units back to their presumed position prior to motion on this fault, which would have formed after emplacement of the Luning allochthon, also elucidated the nature of major structures previously obscured by the Mesozoic pluton which occupies the central part of the Northern Cedar Mountain range.

Units in backthrust nappes I and II together appear to represent a nearly complete section of the Luning Formation, telescoped slightly by the South thrust and forming the upper plate of the Cedar Mountain backthrust. These units presumably originated in the upper plate of the ramped sole thrust, whose surface expression is represented by the trace of the Luning thrust, and their northward-facing stratigraphy is consistent with this assertion. Units in nappes III and IV north of the Cedar Mountain backthrust have stratigraphic tops to the south and are therefore in the reverse orientation to the ramped units from which the Cedar Mountain thrust nappe originated. These ramped units, which presumably included considerable thicknesses of the Dunlap and Sunrise-Gabbs formations as previously discussed, must therefore underlie the Cedar Mountain backthrust's upper plate, and models must accommodate the total original ramped section and allow it to "surface" between nappe III and the Luning thrust. Section thickness is modelled as 5 km, but preserved sections alone account for approximately 3.5 km. The distance between the traces of the Cedar Mountain thrust and the Luning thrust,
measured along B-B' (Plate 1), is 4 km. Simple geometric calculations show that accommodation of ramped units between nappe III and the Luning thrust require a minimum ramp angle of 63°, assuming that only 3.5 km of section is present on the ramp. Decreasing the angle of the ramp to a more realistic geometry leaves insufficient room for ramped units between nappe III and the Luning thrust. The same problem arises if thickness beyond that for preserved units is added to the section. Both realistic ramp angles and reconstructed section thicknesses thus place nappes III and IV in a downcutting relationship over ramped units. Models for deformation in fold and thrust belts call for latest forward-directed motion on the leading foreland thrust; no such downcutting relationships should be produced by forward-directed thrusting. A more likely explanation is the existence of a low-angle normal fault beneath nappes I through IV which formed after thrusting, displacing these units from the north to their present position and thus reducing the distance between nappe III and the Luning thrust. Plate 2H shows the present-day surface geology and the modelled subsurface. The postulated geometry of the LNF fault is termed fault 5 in this figure.

Several constraints determined the modelled geometry and motion on the LNF fault. The problem which the LNF fault resolves is the present lack of space for ramped units between nappe III and the Luning thrust, as discussed above. The LNF fault is thus required to pass through or north of the Luning thrust trace by the position of the Gold Range assemblage Dunlap Formation in the Luning thrust's lower plate. Since the LNF fault also passes beneath nappe IV, this requires a shallowly south-dipping fault angle. This also implies that the "Luning thrust"
may actually be the southern LNF fault trace, and the Luning ramp may therefore be north of its modelled location in the subsurface. This possibility will be further discussed later in this section.

Comparison of plates 2G and 2F shows the modelled geometry of, and motion on, the LNF fault. Both were modelled using the minimum displacement and fault angle which would produce the "best fit" between the upper and lower plates. In the reconstruction, the fault passes beneath the Mesozoic volcanics in the pluton, placing the volcanics in the LNF upper plate, and above the carbonate roof pendants overlying the pluton north of the volcanics. Surface data supported this morphology, as follows.

The mapped configuration of the carbonate roof pendants, which were recrystallized to such an extent that their protoliths were impossible to identify, substantially duplicates that of carbonate members in the Luning section in nappe IV (see Plate 1). The shape of the area between them is similar to the configuration of the middle clastic member of the Luning Formation in nappe IV, which as previously discussed was preferentially melted during emplacement of the pluton. Steeply dipping bedding planes and foliations in the roof pendants also substantially duplicate those in nappe IV. Based on this relationship it was inferred that the roof pendants might be metamorphosed equivalents of nappe IV, structurally continuous with them prior to motion on the LNF fault. Restoration of nappe IV above the roof pendants shows that the LNF fault angle must continue to be shallow to this point in its northward projection.

The reconstruction was also constrained by the position and timing of
emplacement of the Mesozoic pluton and consanguinous volcanics. The LNF's northward projection requires that it unroof the area now occupied by the pluton at no great distance above the present surface, but the coarse-grained texture of the pluton requires that it was emplaced at depth. Inception of LNF faulting, therefore, postdated pluton emplacement. The Mesozoic volcanics were extruded near the paleosurface at the same time the pluton was emplaced, and must therefore rest in the upper plate of the LNF fault.

Restoration of the LNF's upper plate to its inferred pre-fault position clarifies the structure of the Northern Cedar Mountains as a whole. Nappe IV and the roof pendants are seen to be the southern limb of an east-west trending anticline whose northern limb is preserved in the rocks of the Luning exposed north of the pluton. The pluton occupies the hinge of the anticline, with the Mesozoic volcanics extruded in the apex of the hinge.

Timing of LNF faulting is constrained to postdate pluton emplacement. Relations between roof pendant rocks and their equivalents in nappe IV also have implications for the age of this fault. The exposure pattern of the middle clastic member in nappe IV, produced by Cenozoic high-angle faulting, is duplicated in the configuration of the roof pendants (see Plate 1). This suggests the possibility that displacement on the LNF fault took place after Cenozoic high-angle faulting, perhaps less than 27 my ago (Ekren et al., 1980; Oldow and Meinwald, in prep.). As high-angle faulting continues to the present, no upper age bound is established.

The southern trace of the LNF fault is most likely to lie either
within Tertiary volcanics between the Luning thrust and the Cedar Mountain thrust or to coincide with the Luning thrust trace. The relationship of the LNF fault to Tertiary volcanism was not investigated by us, but work by R.F. Hardyman and D. Whitehead (personal commun., 1986) indicates that low-angle faults in fact do cut the Tertiary volcanics.

As shown in the forward model and supported in further discussion, the Cedar Mountain and Dunlap faults are constrained to sole into a single surface at shallow depth and to join the Luning thrust. The LNF fault would be likely to take advantage of this previously existing surface.

Reconstruction of major folds and thrusts in their presumed pre-LNF fault geometries indicate that the fault acted to separate a major D1 anticline. In the LNF fault's lower plate, this anticline is preserved in nappe V, with the Mesozoic pluton's northern exposure occupying the hinge, and the carbonate roof pendants in its southern limb. Nappes III and IV, in the LNF fault's upper plate, are also from the anticline's southern limb, having been separated from nappe V by LNF faulting. In the reconstruction nappe IV is the less-metamorphosed equivalent of the carbonate roof pendants, while the siliciclastics of nappe III rest above the southern exposure of the Mesozoic plutonics. As discussed in the stratigraphy section of this study, the pluton's emplacement was accompanied by metamorphism of carbonate rocks and preferential melting of siliciclastics in the Luning Formation. It is inferrable that the pluton also preferentially melted the Dunlap Formation, which suggests a reason for the superimposition of nappe III and the pluton in the
reconstruction.

Based on the constraints of surface data, placing the lower carbonate member of the Luning Formation in the core of the anticline, this structure is modelled as a ramped anticline which is detached at relatively shallow depth. The fault which forms this detachment is at considerably more shallow depth than constructions will allow for the basal decoupling surface of the Luning thrust. Fault 1 in the forward model is the postulated detachment surface for this structure. As the model shows, this fault has no surface expression because both backthrusting and subsequent extensional faulting have obscured it. Fault 2 is postulated to account for the shape of the anticline and will be discussed in the presentation of the forward model later in this section.

Dunlap thrust -- final discussion and conclusions

Modelling of cryptic structures provides more data for evaluation of the nature of displacement on the Dunlap thrust. Models previously discussed were that of a folded high-angle normal fault, a low-angle normal fault, and a backthrust. The folded high-angle normal fault case, as shown in Figure 17, requires that, in order to rotate the fault to its present geometry, units on both sides of the fault deform as a unit with little or no reactivation of the fault during contraction, producing a syncline in units to the south. The minimum depth to the basement needed to accommodate the complete section folded in the syncline would be more than 5 km, even if only preserved units (Luning and Dunlap) were included in the fold. Such a structure cannot be accommodated in the subsurface immediately north of the Luning thrust trace. In any case it seems
unlikely that a previously existing fault, ideally oriented for reactivation during contraction, would not reactivate in preference to deformation of surrounding units.

The low-angle normal fault model, as previously discussed, requires that the Dunlap fault's upper plate, which includes nappes I, II and III, originate in the southern limb of an anticline north of nappe V. The Luning Formation preserved north of the pluton (nappe V) represents the northern limb of an anticline. Nappe IV is from the southern limb of such an anticline, but in the low-angle normal fault case cannot be the limb from which the Dunlap fault's upper plate originated (see discussion in structure section and Figure 16). The problem then becomes finding a possible site of origin for the fault's upper plate. Were this site south of nappe V, another syncline/anticline pair would be required between nappes IV and V; however, this area is too small to accommodate such a structure. The next possible site of origin would be in an anticline north of nappe V, outside the Cedar Mountains and therefore at least 16 km north of the Dunlap fault. Northward projection of the preserved fault surface to this position places the site several kilometers above the present ground surface.

As the Dunlap fault's upper plate includes the Cedar Mountain and South backthrusts, a model for this case must also accommodate the space required for their formation, and to explain the kinetics by which they formed at this site. The Cedar Mountain and South thrust nappes measure 5.5 km in the plane of the cross-section, so at least this distance must be added to the northward projection of movement on the Dunlap fault in the low-angle normal fault case.
A model for the low-angle normal fault case thus requires a number of hypothetical structures and a minimum of 21.5 km horizontal component of motion on the fault. While it might be possible to construct such a model, it seems an unnecessarily complicated solution to the problem in view of the evidence in support of the backthrust model.

As previously discussed, the backthrust model requires minimum transport on the Dunlap thrust. South-to-north backthrusting of the anticline's southern limb, thrusting the Dunlap section from the anticline itself back over inferred Sunrise-Gabbs and preserved Luning section (demonstrably truncated by the fault), requires only approximately 1.5 km of transport. D1 cleavages in the Dunlap Formation in nappe III of the fault's upper plate are predominantly south-dipping and support modelling south-to-north motion on this fault.

**Presentation of forward model**

A forward model for the kinematics of deformation in the Northern Cedar Mountains (constructed along A-A', Plate 1) is presented in Plate 2. Steps in the model were chosen to represent each stage of deformation at its inception and at the point it ceased. The sequence of deformation for forward-directed thrust faults (faults 1, 2, and 3) proceeds from hinterland to foreland (from north to south) in the thrust belt. Forward-directed thrust faults uniformly cut upsection. Backthrusts (faults 4) are modelled as a late-stage accommodation of ramping at the Luning sole thrust ramp. Although they cut upwards, because they cut structures formed previously by faults 1 and 2 they locally cut down stratigraphic section. Fault 5 is the LNF fault discussed above.
Numbering of faults in the cross-sections corresponds to the time-sequence of motion on the faults, with fault 1 the earliest and fault 5 the latest formed.

Faults 1, 2, and 3 are all thrust faults with the SE-directed transport which was the normal forward direction in the Luning allochthon emplacement regime. Fault 3 is the Luning thrust. Faults 1 and 2 have no surface expression in the study area, being obscured by subsequent deformation, but are required for modelling of the ramped anticline.

The Cedar Mountain, South and Dunlap backthrusts are grouped as faults 4. Apparent complexities in these fault surfaces in the restored section (Plate 2A) are artifacts of their deformation of structures formed in previous (forward-directed) thrusting.

Fault 5 is the low-angle normal fault, or LNF fault, whose inception postdated thrusting. It undercuts backthrusts and is modelled as joining with them at depth. Motion on the low-angle normal fault was from north to south. It is modelled as displacing Mesozoic volcanics, and all pre-Tertiary rocks south of them in the Luning thrust upper plate, a distance of approximately 4 km.

Plate 2B shows modelled configuration at the end of motion on fault 1, which forms the detachment beneath the ramped anticline in the study area. The basal detachment level within the Luning's lower carbonate member is required by units preserved in the anticline. The position of the ramp was chosen to minimize the length of doubled section. Deformation in the modelled additional preserved uppermost section (KJd? in the figure) is simplified as it is of no importance to the model.
Plate 2C shows modelled configuration at the end of motion on fault 2, folding fault 1 and deforming the ramped anticline to its present morphology. Fault 2 is a geometric construct, proposed here to account for the observed morphology of the ramped anticline within the constraints of balanced cross-section preparation. It is proposed only as one possibility for this deformation.

Fault 2 is modelled as a splay of fault 1 with a detachment surface at the base of the Dunlap Formation which eventually ramps to and rejoins fault 1. This forms a duplex structure with a "horse" of Dunlap Formation beneath the core of the ramped anticline. The model as presented requires differential bedding-plane slip at multiple levels in the Dunlap "horse".

In Plate 2D motion on fault 3, the Luning thrust, is shown up until the moment of inception of motion on backthrusts. The most important feature of this cross-section is the modelled position of the backthrusts at their inception. They are modelled here as accomodations of ramping, with their cutoff ramp beginning at the base of the Luning ramp. Wiltschko (1981) presents several models of ramping thrusts which show stress patterns productive of backthrusting of this morphology. Motion on fault 3 continues concurrent with motion on faults 4, but actual southward transport of units south of the ramp is modelled as decreasing because of the accomodation of ramping units by backthrusting.

Plate 2E shows modelled configuration at the end of motion on faults 3 and 4. Ramping of the backthrusts' lower plates has rotated and shallowed the backthrust fault surface angles. Contraction of the Luning thrust upper plate during the period between this and the previous
figure occurred principally by relative northward motion on the backthrusts as the ramping plate underthrusted them. The backthrusts have cut north through the ramped anticline, locally producing younger-over-older relationships as they deformed this earlier structure.

Plate 2F represents the period of time between contractional and extensional deformation. It models erosion and subsequent emplacement of Mesozoic plutonic and volcanic rocks in the hinge of the ramped anticline. The pluton is modelled as preferentially melting siliceous clastic units in accordance with field observations. The figure also shows the proto-LNF fault surface.

Modelled configuration after extensional motion on fault 5 (LNF fault) is shown in Plate 2G. The geometry of the LNF fault requires that it join the backthrusts at depth. The position and geometry of this fault indicate the possibility that it may have originated as a backthrust, reactivated here by extension. The fault surface shared by the LNF fault and the backthrusts is also shown as intersecting the Luning thrust at the Luning thrust trace. Geometric balancing requires that this surface pass near or through the trace. The "Luning thrust" upper plate exposure may in fact be part of the Cedar Mountain backthrust plate, displaced to its present location by LNF faulting. If so, the Luning thrust ramp may actually be north of its modelled location. The position of the anticline in the lower plate of the LNF fault will not allow the ramp to be more than 2 km north of its modelled location, as the anticline would begin to cut downsection over the units on the ramp were it any further northward.
Evaluation of forward model

The only variations of significance for the evaluation of the forward model are those which might change the salient features of the model. Actual conditions which might vary sufficiently from hypothetical constraints to cast doubt on the model include:

-- primary variability in the thickness of the Dunlap Formation;
-- inaccuracy of estimates of extra thicknesses of upper-plate section; and
-- variations in deformational style from the Rocky Mountain model, particularly as regards the conditions of plane strain and concentric folding.

It is possible that the Dunlap Formation did vary markedly in depositional thickness within the environs of the study area, and it is probable that estimates of additional unit thicknesses are incorrect. Major structures in the model which are sufficiently model-dependent to be called into question by variations in these parameters are the LNF fault and the ramp and basement surfaces. The point of departure for arguments for the existence of the LNF fault was the presumed thickness of section in the ramping Luning thrust's upper plate. The units of Luning Formation which comprise nappes I and II are the only preserved section from these ramped units. A model which assumed that only these units were present on the ramp during thrusting could be accommodated without postulating the LNF fault. Such a model would, however, fail to take into account the fabrics in preserved section, which indicate that the section had been buried and lithified prior to, and remained buried during, emplacement of the fold and thrust belt.
Regional stratigraphic relations also call for the presence of the Sunrise and Gabbs formations, which overlie the Luning everywhere depositional contacts are preserved. The aggregate thickness required by these two factors is too great to be accommodated on a ramp with a realistic geometry, without the presence of the LNF fault. If the real section were thinner than modelled, the chief effect on the model would be to reduce the angle on the Luning thrust ramp and the depth to the basal thrust surface.

The plane strain condition of the Rocky Mountain model is invalidated by the presence of multiply-oriented phases of deformation, notable D1 and D2. However, all major contractional structures in the model were formed during D1, and there is no evidence that D2 made more than minor modifications in these structures. As the cross-section was constructed along a NNW-trending line, approximately parallel to D1 transport, D2 transport is approximately perpendicular to the plane of the cross-section. This minimizes the effect of D2 transport on the model, and this serves also to minimize estimates of shortening.

Localized deformation such as the development of mesoscopic structures at the leading edges of backthrusts locally thickens section at these locations and invalidates the constraint of concentric folding. However, variation on the map scale from this cause is slight. Only the presence of units experiencing ductile flow would allow significant thickening of the hinges of map-scale folds, and thinning of the limbs, sufficient to invalidate the model. None of the units preserved or inferred in the study area are of appropriate lithologies to deform in
this fashion at the pressure and temperature conditions indicated by the fabrics and generally very low metamorphic grade of the rocks.

Although the forward model is a useful tool for analyzing the kinematics of deformation in the study area, there are a number of limitations on its use as a predictor of subsurface structure. It predicts only the minimum angle on and approximate position of the Luning thrust ramp rather than the actual angle, position and detailed morphology of the ramp surface. The depth to the basement is accurate only within a 1 to 2 km range as it is dependent on the actual thickness of the upper plate section overlying it. The model can predict only the southernmost possible surface trace of the basement unconformity rather than its actual position, angle or detailed morphology. The position and ramp cutoff angle of fault 1 are both geometric constructs, the actual length of doubled section being unconstrained by field data. Fault 2 is a geometric construct for the existence of which there is no direct evidence. Finally, the model in the subsurface is highly schematic for the sake of simplicity in construction. Faults are depicted as planes or smooth curves when in all likelihood there is a great deal more variability in fault surfaces.
TECTONIC IMPLICATIONS

Results of this study have a number of implications for the tectonics of the region as well as for the evaluation of structures in fold and thrust belts.

Relationship of the Luning thrust to preexisting features

The Luning thrust, mapped in the Northern Cedar Mountains for the first time during this study, has been traced through several other localities to the west (Oldow, 1981a and 1983a) (Figure 4). The spatial correspondence of the Luning thrust trace to the $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ contour, and the existence of the ramp in the Northern Cedar Mountains, are provocative. These observations support earlier interpretations (Ferguson and Muller, 1949; Oldow and Geissman, 1982; Oldow, 1984a) that the Paleozoic sialic/simatic crustal boundary corresponding to the 0.706 line controlled deformation in the Luning-Fencemaker fold and thrust belt.

Degree of shortening

This study results in a prediction of minimum shortening in the Luning allochthon section preserved in the study area and also has implications for shortening in the region.

The kinematic forward model, prepared throughout to minimize shortening, indicates a minimum of 40% shortening with overall southward transport of approximately 40 km. Several factors probably contribute to significantly increase actual shortening in the area.
In that the line of the cross-section is approximately parallel to D1 transport and perpendicular to D2, the model minimizes shortening from D2 deformation. Direct evidence from the field area for contraction during D2 is principally contained in the north-vergent D2 folds localized in nappe II, the Cedar Mountain thrust plate. The absence of such structures in the Dunlap and South Thrust plates imply that they were not reactivated during D2 deformation. Poor preservation of structures at the "Luning thrust" trace, and the possibility that this trace is actually the southern terminus of the Cedar Mountain thrust plate emplaced in that position by LNF faulting, obscure evidence of Luning thrust reactivation during D2. As forward-directed (south-vergent) D2 structures are found in several localities in the upper plate of the Luning thrust, it is likely that some reactivation of the Luning thrust, accompanied by backthrusting on the Cedar Mountain thrust plate, occurred during D2.

The actual depositional site for the upper plate of the Luning thrust was probably significantly northward of its modelled location, as unrealistic onlap relationships in the restored section imply.

The at least partial synchronicity of Luning and Sevier thrusting indicates the possibility that a regional decoupling surface connects the two systems. If so, it must lie beneath lower-plate rocks of the Luning system, implying that they too are allochthonous with respect to Luning-Fencemaker thrusting.

Tectonic shortening and thickening of units is ignored by the model, which assumes that deformed and restored thicknesses are the same.
No direct predictions regarding shortening in other areas in the region are possible based on the results of this study. Only the preparation of balanced cross-sections and forward models for individual areas will provide this information. However, the study does have implications for the magnitude of shortening in other localities.

Overall, there are no significant facies changes between Luning sections in different thrust nappes in the study area. Abrupt facies changes in the Luning between nappes in other localities such as the Pilot Mountains (Oldow, 1981a) may imply significantly greater amounts of shortening than the 40% represented in the Cedar Mountains.

Structures in the Northern Cedar Mountains are indicative of less contraction than has been inferred in other areas. In the study area, minor folds are open to tight and are localized in backthrust plates, whereas in other areas in the Luning allochthon (Oldow, 1978; 1981a; 1984a) D1 deformation is associated with tight to isoclinal folds and D2 with close to tight folds, generally well-developed throughout thrust nappes.

Based on these observations it might be expected that balanced cross-sections and forward models for other areas in the allochthon will show significantly higher amounts of shortening than that estimated in the study area.

Reversals of transport direction in fold and thrust belts

The study shows that local reversals of transport direction do not necessarily have implications for the overall transport regime. Backthrusting in the study area produced little absolute northward
transport. Backthrust upper plates essentially maintained their position as forward-moving units underthrust them, moving slightly north as the ramping plate beneath them rotated and shallowed the backthrust fault surfaces.

**Upper/lower plate fault relationships in areas with multiple deformations**

Relationships considered diagnostic of contractional or extensional faulting must be viewed with caution in areas experiencing multiple episodes of deformation. Faulting of previously folded units, whether extensional or contractional, may locally produce relationships previously considered anomalous. This has been demonstrated by the younger-over-older, section-downcutting relationships the Dunlap and Cedar Mountain thrusts produce, but is also true for extensional faulting. The LNF fault provides an example. The southern exposure of the Mesozoic pluton replaces the younger units of the southern limb of the anticline it intrudes. Removing the effects of pluton emplacement indicates that the LNF fault would have juxtaposed the Luning section in domain IV over the Dunlap Formation in the anticline's southern limb, producing older-over-younger relationships by extensional faulting. Careful analysis of surface data, and the use of balanced cross-section techniques to reverse deformational effects and check proposed models, are important tools in determining the actual nature of faulting areas with multiple episodes of deformation.
CONCLUSIONS

Summary of results in the study area

Mapping of the Luning thrust trace, not previously observed in the Northern Cedar Mountains, provided constraints for the preparation of a kinematic forward model which showed the following sequence of deformational events in the area.

Emplacement of the Luning allochthon in the study area during NW-SE (D1 deformation) shortening on the order of 40% was preceded during this phase by the emplacement of a ramped anticline. Late-stage backthrusting resulting from ramping of the allochthon produced three backthrusts, mapped for the first time during this study, with relative northwesterly transport within the overall southeasterly-directed regime. These backthrusts locally produced younger-over-older relationships by faulting of the previously formed anticline. Lesser NE-SW (D2 deformation) contraction produced no map-scale structures but locally reactivated at least one of the backthrusts and probably the Luning thrust as well. Units in the Luning allochthon in the study area underwent a minimum of approximately 40 km of southerly-directed transport during D1. Relationships established elsewhere in the region tentatively constrain the timing of these two events between the Middle to Late Jurassic and the Late Cretaceous.

Contraction was followed by a period of no significant deformation, sufficiently long to allow substantial erosion of the anticline prior to emplacement of the pluton and coterminous volcanics in its hinge.
Regional relations indicate that this pluton might have been emplaced from 100 to 70 my ago.

Subsequent low-angle normal faulting, possibly reactivating an older backthrust surface, displaced structures from the anticline and backthrusts to the south. Inception of this event is constrained to postdate pluton emplacement and may have postdated the onset of NW-trending high-angle Basin and Range faulting which began in the region less than 27 my ago. High angle faulting was preceded and accompanied by siliceous Tertiary volcanism.

The use of balanced cross-section techniques with surface data

The results obtained in this study demonstrate the power of the balanced cross-section technique used in conjunction with surface data. Traditional analyses of surface data (mapping, geometric fabric analysis and neutral cross-sections) were vital in providing the framework for modeling deformation, but only the application of balanced cross-section principles in the stepwise removal of multiple episodes of deformation allowed the preparation of the forward tectonic model. The model elucidates the tectonics of the area to a degree impossible with the use of traditional techniques alone.

The balanced cross-section/stepwise forward model technique is an inexpensive tool, the use of which may be especially important in areas such as the Basin and Range where subsurface seismic data on Mesozoic structures is not available.

The study also indicates that surface fabric data supplies constraints for subsurface analysis which may be important where
seismic sections are available, especially in areas which have undergone multiple episodes of deformation. Fabric data analysis, in constraining fault motions, has important implications for determining the tectonic regime in which they form. These regimes are not necessarily obvious from the geometries of seismic sections alone.

**Hazards of pattern-recognition in faulted and folded terranes**

In areas which have undergone multiple episodes of deformation, care must be taken in characterizing faults by the age or stratigraphic relationships of the units they juxtapose. Faulting of folds, whether by contraction or extension, may readily produce either older-over-younger or younger-over-older relationships. This again points out the importance of surface data in determining the nature of tectonic regimes.
APPENDIX I

MESOZOIC FORMATIONS IN THE NORTHERN CEDAR MOUNTAINS: ROCK DESCRIPTIONS

LUNING FORMATION:

-- Lower carbonate member (Tr1):

Data was insufficient to require description here.

-- Middle clastic member (Tr1mc):

A. Rounded-pebble chert pebble conglomerate (140m):

Ridge-forming chert pebble conglomerates grading laterally and vertically into siliceous mudstones with minor sand components. Chert pebbles are white, moderately well-sorted (2 to 5 cm diameter), round, and in matrix and occasional clast support. Matrix is beige to white and aphanitic in hand sample, in thin section identified as microcrystalline chert with occasional possible relict grains. Minor sand components are predominantly chert with minor quartz.

B. Angular chert pebble conglomerate (100m):

Conglomerates grade laterally and vertically into coarse chert sandstones and sandy mudstones. Conglomerate matrix is gray aphanitic siliceous mudstone; clasts are subangular to subrounded, poorly sorted (1mm to 4 cm), gray and red chert (to 55%) and milky to translucent quartz and chert (to 25%) in both clast and matrix support.

C. Quartz/chert arenite (145m):

Slope-forming green quartz/chert arenite in siliceous mudstone matrix. Quartz content increases upsection. Quartz grains are rounded and well-sorted (from 0.2 to 0.7 mm), with frequent bubble trails. Quartz comprises 50 to 70% of the rock. Chert grains are the same size but are less round. Chert abundance decreases upsection from a maximum of 15% to negligible quantities. Chert grains appear slightly to moderately compressed between adjacent quartz grains.

D. Siliceous mudstone and quartz arenite (140m):

Yellow to brown siliceous mudstone interbedded with occasional gray quartz arenite. One lens of ridge-forming chert pebble conglomerate occurs approximately 50m from the top of the member. Clasts are round to oblong rounded chert pebbles up to 3 cm in length, in coarse chert sand matrix support. Conglomerate grades laterally and vertically into
coarse chert sand. The uppermost 15m are gradational with overlying Tr1ms member. Yellow-brown thin- to medium-bedded siliceous mudstone is interbedded with thin-bedded gray to brown siliceous silty micrite.

-- middle shaley carbonate member (Tr1ms):

A. Gray to black silty micrite (75m):

Massive, gray to black, predominantly slope-forming silty micrite (minor fine-grained chert and quartz), grading into thick-bedded, resistant gray and black micrite with irregularly shaped brown siliceous nodules. Elsewhere such nodules are identifiable as fossils.

B. Fossiliferous silty micrite (60m):

Basal 2-meter bed of vermicular black, finely fossiliferous calcareous silty micrite is overlain by slope-forming silty micrite interbedded with 1- to 2-m thick beds of massive to nodular black to gray micrite + fossils in matrix support. In thin section fossils are identified as pelecypod and echinoid fragments with occasional brachiopods, gastropods and micritized ovals (possibly pellets). A dome-shaped stromatolite was found in float.

C. Slope-forming micrite (70m):

The lower half of this section is of variably composed slope-forming units: black, massive, non-fossiliferous micrite, interbedded with thick-bedded gray fossiliferous calcareous siltstone. Fossils are brown siliceous nodules identifiable as gastropods and bivalves to 2 cm in diameter, in matrix support. The upper half of this section is obscured by float.

D. Resistant silty micrite (30m):

Massive resistant gray to black, very thick-bedded (1 to 2m), sparsely fossiliferous; silt is calcareous.

E. Fossiliferous silty micrite (20m):

Slope-forming fossiliferous silty micrite. The lower 15 m are black, wavy-bedded and nodular. The upper 5m are gray and thick-bedded (0.75m). Megafossils in matrix support include large (to 20 cm long) high-spired gastropods, nautiloids Grypoceras sp., pelecypods Trichites and star-shaped crinoid stems tentatively identified as Pentacrinus. Matrix contains pelecypod and gastropod fragments to 1 cm in diameter.

F. Slope-forming calcareous siltstone (10m):

Rusty brown slope-forming calcareous siltstone; no fossils preserved.
G. Fossiliferous silty micrite (25m):

Ridge-forming thick-bedded gray silty micrite (calcareous silt), variably fossiliferous. Fossils: pelecypod and gastropod fragments in matrix support.

H. Fossiliferous calcareous sandstone/siltstone (25m):

Slope-forming nodular gray carbonate, highly fossiliferous.

I. Ridge-forming silty micrite (25m):

Thick-bedded gray ridge-forming silty micrite (calcareous silt) with occasional brown siliceous shell fragments.

J. Coquinoid calcareous siltstone (450m):

Slope-forming thin-bedded yellow gray calcareous siltstone, interbedded with slope-forming nodular and fissile black to brown thin-be-ded coquinoid calcareous siltstone. Fossils consist almost entirely of pelecypod fragments with occasional echinoids. Bioclasts are black in hand sample and appear in approximately 50% of the section. At 1- to 20m intervals are resistant, 1- to 2-m thick units of black to gray silty micrite, occasionally with pelecypod fragments in matrix support. The upper 15m of this section consists of nodular yellowed slope-forming thin- to thick-bedded silty micrite matrix, supporting abundant fluted bivalves 4 to 8 cm in diameter.

--- Upper carbonate member(Tγ₁u):

A. Massive nodular micrite (5m):

Gray massive 2-m thick beds of ledge forming micrite, interbedded with slope-forming calcareous siltstone. Micrite contains brown siliceous irregularly shaped nodules and occasional chert nodules. The section coarsens upward to calcwacke.

B. Calcareous shale/calcareous mudstone (20m):

Planar, medium-bedded (5 to 10 cm) fissile calcareous shale, interbedded with calcareous mudstone.

C. Massive nodular slightly fossiliferous coralline micrite (50m):

Gray to black micrite; lenticular chert nodules locally define bedding planes in basal beds. Corals resembling those described as Thecosmilia sp. found elsewhere in this unit by Mottern (1962) occur as poorly preserved carbonate and brown siliceous nodules.

D. Slope-forming carbonate (30m):
Medium- to thick-beded (0.5 to 1m thick) slope-forming calcareous mudstone.

E. Massive fossiliferous micrite interbedded with silty micrite (35m):

Gray to black resistant massive (0.5 to 2m thick micrite interbedded with ledgy medium-beded (15 to 20 cm) calcareous silty micrite. Massive micrite is occasionally finely fossiliferous with pelecypod fragments in matrix support. At the top of this unit are 2m of 0.5m thick beds of nodular micrite in which fossil content increases upsection. Large gastropods (1 to 8 cm diameter); micrite intraclasts or pellets, ovoid and 1-3cm long.

F. Massive nodular coralline micrite (30m):

Gray to black massive (4m) resistant micrite. Pelecypod fragments; gastropods to 2cm diameter. Corals preserved as carbonate and brown siliceous nodules.

G. Thin- to medium-beded nodular micrite (10m):

Black, wavy-beded and ledge-forming, ± lenticular chert nodules.

H. Fossiliferous ± sandy micrite (25m)

Gray to black medium-beded (25-30 cm), ledgy, slightly wavy-beded ± sandy micrite; fossil hash of pelecypod and gastropod fragments.

I. Calcereous wacke (5m):

Black, laminar to massive calcereous wacke.

J. Micrite, ± fossiliferous (55m):

Gray medium-beded calcereous shales and mudstones (75m):

Thin-beded (10 to 20cm) ledge-forming gray and black calcereous shales and mudstones, ± small chert nodules, interbedded with fissile calcereous shales.

DUNLAP FORMATION -- LUNING ASSEMBLAGE:

-- Lower quartzose member:

A. Basal conglomerate and breccia (1.5 to 5m):

LATERALLY VARIABLE. Several sections have a three-layer lithology with carbonate-rich layers bracketing a quartzose layer. Basal
carbonate varies from limestone conglomerate and breccia to massive limestone to calcareous siltstone; clasts 1mm-6cm diameter, subangular to subrounded. ± chert nodules. The central quartzose layer is also variable, from quartzose micrite to quartz arenite to limestone/chert pebble conglomerate and breccia in quartz sand matrix. In thin section the matrix is composed of quartz grains 0.1mm-0.3mm in diameter (25 to 60% of the matrix) and chert grains (5-10%), in matrix support in echinoid-rich biomicrite. Larger quartz grains are round; smaller grains are angular to subangular. Conglomerate and breccia clasts are angular microcrystalline chert and echinoid-rich biomicrite, 2mm-2cm in diameter. Clasts in matrix support. Secondary minerals include pyrite and tourmaline. The uppermost carbonate layer is comprised of quartzose sandy micrite ± chert pebbles, brecciated in subangular clasts 1-3cm in diameter. Locally abundant concentrically banded cherts, also brecciated, and irregular hollow chert nodules filled with quartzose micrite. Extensive carbonate recrystallization, dolomitization, and copper (?) remobilization. Secondary calcite and silica cements.

B. Quartz arenites (2m):

White to yellow-brown quartz arenite. Quartz grains 0.2-0.7mm in diameter, 50 to 70%. Matrix: brown opaque-stained microcrystalline silica. Larger quartz grains are well-sorted and round; smaller grains are subangular to angular. Rare chert grains are subrounded. Bedding not observed; units are 1-2m thick where not obscured by alluvium and float.

C. Quartz arenite/wacke and float, ± chert pebble conglomerate (450m):

Red quartz/chert grainstones, chert = 10-20%. Poor outcrop; 80% of this area is obscured by float and alluvium. Chert pebble conglomerate contains elongate, subangular to subrounded clasts, 3mm-3cm long. Conglomerate beds approximately 0.5m thick, truncating crossbedded quartz arenite.

D. Lithic/chert coarse sand arenite (16m):

Grains to 1mm diameter; well-sorted red and black mudstone (25%) and white chert/quartz (25%).

-- Upper feldspathic member:

A. Red/brown siltstone and mudstone (490m; 85% obscured by alluvium and float):

Red and brown silty calcareous wackes with fine-grained components of chert, volcanogenic, feldspathic and quartz grains.

B. Limestone/mudstone conglomerate, chert wacke and mudstone (260m):
Red to purple fissile mudstone, chert silt wacke, sand and minor conglomerate. Conglomerate clasts: gray limestone (45%); red mudstone (15-25%); subrounded, 2mm-1cm diameter. Red mudstone matrix. In thin section sandy wacke is composed of 15-40% quartz, 10-50% chert, 0-30% red mudstone and 0-10% volcanogenic grains with occasional altered feldspar laths in red-stained calcareous clay matrix. Mafic volcanogenic grains visibly altering to red/brown opaque clays and calcite.

C. Siltstone/mudstone, grading upward into limestone (160m):

Red to purple silty wackestone; in thin section grains comprise up to 25%; subrounded to subangular, very fine-grained quartz (0-10%) and chert (0-10%) in purple-stained calcareous mud matrix. Occasional deposits of fine-grained authigenic?/depositional? dolomite rhombs in mud matrix and grain support are only recognizable in thin section. Very fine-grained angular feldspar. Red semi-opaque clasts, locally elongated by soft-sediment deformation, to 1mm long, ± feldspar laths to 125 microns long, locally comprise up to 30% of the sediment. Lithic clasts comprised of feldspar laths "aflot" in microcrystalline calcite matrix locally comprise 5-10%.

In the uppermost 20m of this section, red and green silt/mudstone is interbedded with leggy medium-bedded (20 - 40cm) gray to yellow silty unfossiliferous micrite.

DUNLAP FORMATION -- GOLD RANGE ASSEMBLAGE:

A. Red mudstone/cherty sand wacke, ± unfossiliferous limestone lenses (50m):

Red/brown chert silty wacke, grading laterally and vertically to occasional gray-green chert/quartz sandstone and wacke. Chert grains are subangular, moderately well-sorted (0.3-1mm) and comprise approximately 50% of the rock. Quartz grains are rounded, slightly smaller than chert grains (0.1-0.4mm), and comprise 25-35% of the rock. Smaller grains are more angular than larger grains.

B. Lithic pebble conglomerate and red mudstone (30m):

Clasts: red-brown mudstone, subrounded, poorly sorted (2mm-10 cm diameter); gray and red-brown chert, subrounded; gray to red mudstone; purple/brown/black mafic volcanic clasts with white feldspar laths; and gray aphanitic volcanic clasts with weathered hornblende laths. Matrix: red mudstone and/or fine-grained equivalents of clasts. Conglomerate is interbedded with red mudstone and blue-gray chert sand wacke.

C. Chert pebble conglomerate (20m):
Subrounded, poorly sorted (2mm-10cm) chert clasts in grain support. Red, black and blue-gray clasts in buff to blue-gray chert matrix. Partially silicified clasts in lower sections of this unit evidently have sandstone, volcanic and limestone protoliths; silicification increases upsection.
APPENDIX II

MESOZOIC TUFS AND TUFF BRECCIAS IN THE NORTHERN CEDAR MOUNTAINS:
ROCK DESCRIPTIONS

The following sample descriptions are arranged in location order from east to west.

S23, S32, S33: Rhyolite/rhyodacite crystal-lithic welded tuff. Phenocrysts, in grain support, include:
   -- Quartz: rounded, translucent, 0.5-1.5mm diameter; 30-50%
   -- Alkali feldspar: subangular (anhedral to lathlike), white to pink, unzoned, 1-5mm diameter; 15-50%

Subangular clasts of gray chert and fine-grained red sandstone from 5mm to 2cm in diameter comprise less than 5% of the rock. Groundmass: light to medium gray, aphanitic, vesicular. Cavities 0.5-1mm in diameter comprise less than 5% of the rock's volume.

S38: Rhyolite/rhyodacite crystal-lithic welded tuff. Phenocrysts, in groundmass support, include:
   -- Quartz: transparent to translucent, rounded, 1-3mm in diameter; 20-30%
   -- Feldspar: white to pink laths and anhedral to 2mm long; 5-10%
   -- Biotite: rounded plates to 2mm in diameter. Rare.

The groundmass is light brownish-gray and aphanitic, and exhibits conchoidal fracture. Lathlike cavities appear to be voids left by weathering of feldspars.

S37: Quartz-trachyte/quartz-latite crystal welded tuff. Phenocrysts in groundmass support include:
   -- Feldspars: translucent fresh, white weathered, subhedral, striated, 0.5-2mm diameter; 30%
   -- Quartz: rounded, to 2mm diameter; 15%

Groundmass is black and finely crystalline with irregular fracture. Clasts of fine-grained sandstone and voids with remnants of these clasts, 0.5-1.5cm in diameter.

10118302 and 10118303: Latite-quartz trachyte crystal tuff/breccia; quartz-trachyte tuff with clasts of latite. Quartz-trachyte phenocrysts are in groundmass support and are composed of:
   -- Feldspar: translucent white to pink; both sanidine and zoned plagioclase, slightly rounded laths and subhedral crystals 0.5-3mm long; 15-25%
-- Quartz: translucent, rounded, 1-5mm in diameter; 5-10%
-- Biotite: brown, broken plates; 0.2-2mm in diameter; 1-5%
-- Hornblende: green to brown, broken crystals, to 1.5mm long; 1-5%

Groundmass of the quartz-trachyte is black to brown and aphanitic. In thin section it appears composed of microcrystalline quartz(?) with identifiable fragments of quartz to 0.4mm in diameter, altered and broken feldspar laths, and secondary magnetite.

Latite clasts are amorphous in shape. Phenocrysts are white zoned plagioclase showing albite twinning in thin section, to 0.05mm long and comprising 50-60% of the latite. The groundmass is black in hand sample and generally opaque in thin section, with occasional identifiable feldspar laths and magnetite subhedral.

10118301: Latite crystal welded tuff, similar in composition to clasts in 10118302 and 10118303, although here phenocrysts are larger. They are:
-- Feldspars: zoned and twinned white plagioclase laths and glomerocrysts, 0.3mm-3mm long, altering to sericite; 40-50%
-- Hornblende, altering to chlorite; rare

The groundmass is black in hand sample and is composed of:
-- Feldspar laths: 0.03-0.1mm long; 25-30%
-- Magnetite: subhedral cubes, 0.01mm on a side; 10%
-- Secondary chlorite: 10%
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\[ R_{\text{m}} \] middle clastic member

\[ F_{\text{li}} \] lower carbonate member

**MESOZOIC IGNEOUS ROCKS**

\[ M_{\text{zv}} \] undifferentiated tuffs & tuff-breccias

\[ M_{\text{zj}} \] quartz-bearing monzonite

\[ rP \] carbonate roof pendant

**CONTACTS:** dashed where uncertain, dotted where projected beneath mapped units

- depositional contact
- thrust fault-teeth on upper plate
- low-angle normal fault-box on upper plate
- high-angle normal fault- \( \) on downthrown side

**OTHER SYMBOLS**

- \( \) strike & dip of bedding
- \( \) D1 cleavage
- \( \) D2 cleavage
- \( \) pencil lineation

**PLATE 1**
NORTHERN CEDAR MOUNTAINS

WEST-CENTRAL NEVADA

L. S. Brown