Overview of Mississippian Depositional and Paleotectonic History of the Antler Foreland, Eastern Nevada and Western Utah

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ABSTRACT

Mississippian strata of the Antler foreland record depositional patterns resulting from the final episode of Roberts Mountains thrusting at the latitude of northern Nevada. They overstep continentward onto the distal parts of the earlier foreland related to Late Devonian-earliest Mississippian episodes of Roberts Mountains thrusting. Distal deposits of the Mississippian foreland are carbonate rocks and fine-grained siliciclastic rocks of continental provenance. Most of the Mississippian is embodied in the successive Morris, Sadlick, and Maughan sequences. The Morris sequence consists of less than 100 m of late Kinderhookian shallow-water carbonate rocks representing a transgressive-regressive cycle that blanketed the entire distal foreland to the distal edge of the foredeep, where it was erosionally truncated prior to deposition of the initial Mississippian foredeep deposits. The Sadlick sequence—of Osagean to Meramecian age and as much as 200 m thick—records in its lower part aggradational to retrogradational carbonate deposition, reflecting protracted subsidence of the distal foreland. In its upper part—during the Delle phosphatic event—low-relief topographic restriction led to deposition of compositionally unusual rocks, partly under the influence of hyposalinity. The Maughan sequence is latest Meramecian to late Chesterian in age and exceeds 1,000 m in thickness. It grades from mainly pelitic, partly paralic rocks in eastern Nevada eastward to shallow-water carbonate rocks in central Utah and reflects a marked change in the paleotectonic controls on deposition in the distal foreland.

The proximal Mississippian Antler foredeep is represented by the Diamond Range sequence and superjacent Newark Valley sequence. A regional unconformity beneath late Meramecian or early Chesterian strata of the Newark Valley sequence separates it from the Diamond Range sequence. Aside from the relatively thin, hemipelagic and gravity-flow, lower Osagean limestones that are part of the foredeep-lag deposits in the initial foredeep, the foredeep fill of the Diamond Range sequence is formed of siliciclastic sediment derived from the Roberts Mountains allochthon and is as much as 1,000 m thick beneath the angular unconformity that defines its upper limit. The Diamond Range sequence was deposited during mid-Mississippian Roberts Mountains thrusting, and its most proximal part was structurally imbricated and telescoped syndepositionally. The Newark Valley sequence—also as much as 1,000 m thick—consists in its lower part mostly of submarine to subaerial fan-delta and braid-delta deposits, which reflect erosion of an orogenic upland to the west, and which overlap structures related to Roberts Mountains thrusting. Rocks of the Newark Valley sequence merge

eastward into the more pelitic rocks of the Maughan sequence, so that these two sequences are essentially synonymous.

Foredeep deposition of the Diamond Range sequence was synchronous with Roberts Mountains thrusting in the proximal foreland and with low-relief topographically restricted deposition of the Sadlick sequence in the distal foreland. Thus, mid-Mississippian deposition within the Antler foreland is consistent with models invoking the foredeep, peripheral bulge, and backbulge pattern of lithospheric flexural-loading paleotectonics. Following emplacement of the Roberts Mountains allochthon, Chesterian paleotectonic controls on accommodation and deposition are not so easily explained. They may combine isostatic rebound of the proximal foreland region and intracratonic strain related to initiation of ancestral Rocky Mountain tectonism in the more distal foreland.

INTRODUCTION

Concepts about the Antler foreland and its orogenic hinterland are continuing to evolve, building on observations and interpretations dating back to the early geologic exploration of the Great Basin in Nevada and Utah. The term "Antler orogeny," however, was introduced less than 50 years ago by Roberts (1949) for the deformation responsible for an unconformity between Pennsylvanian strata and internally thrust faulted and deformed rocks now known to be early Paleozoic in age in the Antler Peak area (AP on Fig. 1) near Battle Mountain in north-central Nevada. Subsequently, the broader region trending north-northeasterly across central Nevada, where this unconformity was observed or inferred, came to be known as the "Antler orogenic belt" (Roberts et al., 1958).

Antler orogenesis was explicitly related by Roberts et al. (1958) to development of the Roberts Mountains thrust. This extensive fault was first mapped in the Roberts Mountains (GC and RC, Fig. 1) and vicinity by Merriam and Anderson (1942), who emphasized the abrupt contrast between the pervasively deformed Ordovician sandstone, mafic volcanic rocks, black shale, and bedded chert of its upper plate and the less deformed, age-equivalent, carbonate rocks in its lower plate. Thrust juxtaposition of these two distinct facies of Ordovician rocks had been anticipated previously by Kirk (1933). The age of the Roberts Mountains thrust was originally regarded as "later Cretaceous or early Tertiary" (Merriam and Anderson, 1942) but was reinterpreted as Late Devonian or Early Mississippian by Roberts et al. (1958) on the basis of regional relations and the apparent derivation from its allochthon of the Mississippian coarse-grained clastic deposits found in east-central and northern Nevada. However, because in its type area (GC and RC, fig. 1) the Roberts Mountains thrust is not depositionally overlapped by pre-Tertiary rocks, controversy about its regional age has persisted to recent times (e.g., Ketner et al., 1993).

The advent of plate tectonics generated a flurry of possible actualistic models for Antler orogenesis and kinematics of the Roberts Mountains allochthon, which was by then generally viewed as a mid-Paleozoic feature. In their report on the Penrose Conference held in 1979 on the Antler orogeny, Nilsen and Stewart (1980) summarized no less than ten possible models to explain the orogeny, seven of them explicitly based on plate-tectonic concepts and invoking subduction, dipping either east or west, at the continental margin. The original and enduring plate-tectonic synthesis of the Antler foreland was presented by Poole (1974) who followed Burchfiel and Davis (1972) in generating the Roberts Mountains thrust and Antler orogenic belt by collapse of a back-arc basin. In this paper, Poole applied to the foreland fill the popular, if somewhat misleading, term "Antler flysch," which along with "molasse" originally had been introduced in a geosynclinal context for the Mississippian strata of the foreland by Sadlick (1960).

Speed and Sleep (1982)—following Moores (1970)—attributed Antler orogenesis and foreland development to arc-continent collision. In their model, the Roberts Mountains allochthon is viewed as a large, east-facing accretionary prism that was underthrust by the continental margin. Although there are reasons (see below) not to apply such a model, at least not for Mississippian time, this paper by Speed and Sleep (1982) is nonetheless a landmark in that it first introduced the mechanism of vertical loading and lithospheric flexure to interpretation of the Antler foreland. The paleogeographic and depositional history of the foreland has been interpreted as being consistent with flexural-loading models (Fig. 2), calling for a continentward-migrating foredeep and forebulge (or peripheral bulge) and a discernible, low-gradient backbulge area at times during Antler orogenesis (Goebel, 1991; Silberling and Nichols, 1991; Giles [*née* Goebel] and Dickinson, 1995; Nichols and Silberling, 1995). These elements of a flexural-loading model are roughly equivalent to those recognized in earlier paleogeographic overviews of the mid-Mississippian in the eastern Great Basin part of the Antler foreland, such as the "flysch trough," "Joana bank"/"submarine rise," and "Deseret starved basin" of Gutschick et al. (1980) and the "flysch trough, rise, and starved basin" of Poole and Claypool (1984). Application of a flexural-loading model along the trend of the Antler foreland to the north of the eastern Great Basin in south-central Idaho and to the south in southern Nevada is, however, less apparent (Link et al., 1996; Trexler et al., 1996).

Further testing of the applicability of the flexural-loading model to the Antler foreland is of considerable interest, but for this task it is essential to separate observation from paleotectonic interpretation. Consequently, the geography of the Mississippian foreland record in eastern Nevada and western Utah is described herein in terms of five outcrop belts that range from most proximal to most distal with respect to the Roberts Mountains allochthon. These belts—designated A through E—are defined by the present-day geographic limits of particular mid-Mississippian rock units, as described in Figure 1. The shapes, relative proportions, and trends of these belts have been greatly affected by post-Paleozoic structural disruption.

Recent adoption of sequence-stratigraphic concepts for the Antler foreland has also proved useful in the eastern Great Basin (Trexler and Nichtman, 1990; Silberling et al., 1995; Giles and Dickinson, 1995). Because outcrop-based sequence stratigraphy focuses attention on the vertical changes in depositional environments, as recorded in the sedimentologic record, and on unconformities, it provides a means of cutting through the ambiguities inherent in attempting to express depositional histories in terms of established, formal lithostratigraphic units. It also provides a physical means of correlating strata and events among the different stratigraphic expressions within and between the proximal and distal parts of the foreland (Fig. 3).

The time span over which the Antler foreland was developed and the definition of what constitutes the foreland have not met with general agreement for a variety of reasons, such as different scenarios for its paleotectonic control, different interpretations of observational geologic data, and even simple semantics. Onset of foreland-basin development was inferred by Speed and Sleep (1982) as Early Mississippian, the maximum age of deposition of clastic material clearly derived from the Roberts Mountains allochthon. Most authors, however, have related initiation of the foreland to the oldest deposits of non-miogeoclinal character and distribution, such as those of the West Range Limestone and Pilot Shale of late Frasnian (mid Late Devonian) age. Interpretation of these Upper Devonian rock units as backbulge deposits (Goebel, 1991) placed them in a flexural-loading, foreland context.

Establishing the younger age limit for those deposits of the Antler foreland that are the direct and contemporaneous consequence of Roberts Mountains thrusting has involved considerable confusion about how the Mississippian thrust system interacted with its foreland and about stratigraphic overlap relationships. The generally held perception of the foreland in the eastern Great Basin has been that overthrusting by the Roberts Mountains allochthon ceased during the Early Mississippian, after which foreland accommodation space was gradually filled during a prolonged post-tectonic phase taking up much of the rest of Mississippian time. This concept of first creating and then slowly filling a deep, persistent trough, however, is difficult to reconcile with reasonable values for the strength of the lithosphere, and it is contradicted by the structural and stratigraphic evidence. As argued below, we take the view that the final episode of thrusting and consequent foreland development took place during mid-Mississippian time and that ultimate stratigraphic overlap of the resulting deformed foredeep in northern Nevada occurred in latest Meramecian to earliest Chesterian (mid Late Mississippian) time.

Antler orogenesis and foreland sedimentation was viewed as having taken place in "several distinct pulses" in the pioneering paper by Roberts et al. (1958). Studies of the foreland record since then have reaffirmed the episodic nature of its paleotectonic evolution (Trexler et al., 1991; Nichols and Silberling, 1995). The record of older episodes of thrusting and foreland development, beginning in the mid-Late Devonian, has been disrupted and partly obliterated by the mid-Mississippian phase of Antler tectonism. Features resulting from this phase of Antler tectonism have in turn been considerably disrupted by post-Mississippian structures. Nevertheless, the resulting pattern of Mississippian foreland development is well displayed in scattered exposures from central Utah west to the break-out of the mid-Mississippian Roberts Mountains thrust in east-central Nevada. The object of this paper is to present a synthesis of this pattern.

DISTAL MISSISSIPPIAN FORELAND

The stratigraphic record in the more distal parts of the Antler foreland—within belts C to E on Figure 1—is more complete than that farther west and provides important paleotectonic insight on the foreland as a whole. This record is wholly composed of carbonate rocks and fine-grained clastic rocks, none of which were derived from the Roberts Mountains allochthon. Facies patterns within the part of the distal foreland shown on Figure 1 are juxtaposed across major faults of the Sevier thrust system, especially by the Canyon Range thrust and its lateral equivalents, but also by the Tintic (interpreted as in Silberling et al., 1995) and other structurally lower Sevier thrusts.

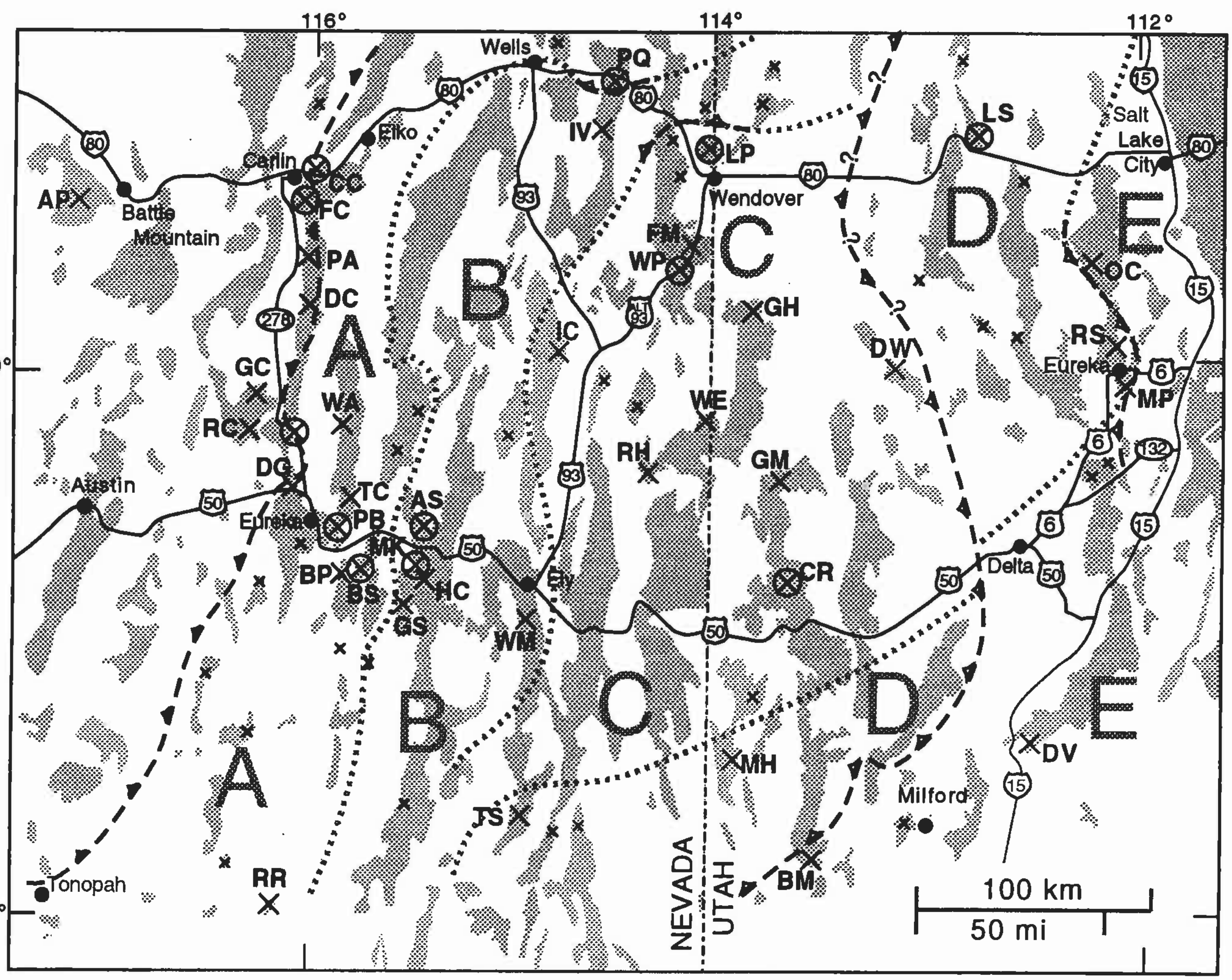
All but the oldest and youngest parts of the Mississippian section in the distal foreland are included in the stratigraphically successive Morris (upper Kinderhookian), Sadlick (Osagean to Meramecian), and Maughan (uppermost Meramecian to Chesterian) sequences of Silberling et al. (1995) (Fig. 3). Not given further consideration here are the earliest Mississippian strata of the Gutschick sequence and post-Maughan sequence Mississippian strata, which are of latest Chesterian age.

Morris sequence

The Morris sequence is composed almost entirely of carbonate rocks, which have a distinct geographic distribution and internal sequence-stratigraphic architecture as compared with those of the overlying Sadlick sequence (Silberling et al., 1995). The Morris is generally about 50 m thick and displays a simple transgressive-regressive depositional pattern like that characteristic of most shelf carbonate rocks. In most places, a relatively thin basal unit of deep-subtidal, rhythmically bedded limestone forms its transgressive systems tract, which is then overlain by a thicker succession of carbonate rocks representing successively shallower water deposits. Massive, cliff-forming, bioclastic (crinoidal) limestone characteristically forms most of its highstand systems tract, the uppermost part of which in belts E, D, and easternmost C is evaporitic penecontemporaneous dolomite or desiccated lime mudstone. The Morris sequence is represented in Utah by most of the strata included in the Fitchville Formation in exposures of belts D and E; farther west in belts C and B it is represented by the lower Joana Limestone (Fig. 3). As such, it forms a blanket over the entire distal foreland, extending westward across belts C and B to the distal side of belt A (Fig. 1), in which it is erosionally cut out towards the west beneath the mid-Mississippian initial foredeep deposits (Nichols and Silberling, 1995). Sequence 7 of Giles and Dickinson (1995) is approximately the same as the Morris sequence.

Sadlick sequence

The Sadlick sequence, overlying the Morris, has an unusual internal sequence stratigraphy in its lower part,



interpreted to have resulted from progressive localized subsidence. In its upper part it contains compositionally peculiar sedimentary rocks, evidently related to topographic restriction of the distal part of the mid-Mississippian Antler foreland. Throughout most of the distal foreland, the contact between the Morris and Sadlick sequences is transitional, marked by an abrupt upward change to more open-marine or relatively deeper water deposits. Where dated by fossils (primarily conodonts), the contact is approximately at the Kinderhookian-Osagean boundary. The Morris-Sadlick sequence boundary roughly corresponds to that between sequences 7 and 8 of Giles and Dickinson (1995), which, however, is apparently defined by the inferred age of conodont faunas rather than the rock record. The Sadlick sequence embraces not only rocks included in sequence 8 of Giles and Dickinson (1995) but also those in the lower part of their "post-tectonic phase" section.

The lower part of the Sadlick sequence is composed of open-marine, shelf limestones, which commonly exceed 100 m in thickness and undergo pronounced east-west facies changes across the original west-sloping depositional ramp from belts E to B (Figs. 1 and 4). In most of belts D and E (Fig. 1) in Utah, the lower part of the Sadlick sequence is the Gardison Limestone, most of which is a thick aggradational succession of distinctively laminated (low truncation angle cross-stratified), fine-grained, pellet-crinoid packstone tidalites, containing lenticular units of crinoidal grainstone, and shelly tempestites. These limestones are then regionally overlain by a retrogradational uppermost unit of deeper subtidal, wave-influenced, notably cherty limestone (Silberling and Nichols, 1992; Silberling et al., 1995). Oolitic crinoidal grainstone, as a unit up to a few meters thick, is characteristically intercalated in the lower part of the laminated tidalite section, and it forms the major share of the section in the southern Wah Wah Range (BM on Fig. 1), beneath the Wah-Wah (=Canyon Range) thrust. Farther north, on the upper plate of this major structure, the lower part of the Sadlick sequence in

Figure 1. Index map of part of eastern Nevada and western Utah showing the location of belts A–E of the Mississippian Antler foreland and the localities designated by letter symbols in the text. Shaded areas are exposures of pre-Tertiary strata and(or) plutonic rocks from Stewart and Carlson (1978) and Hintze (1980).

*Belts are labeled with large shaded letters and bounded by heavy shaded dotted lines where not coincident with faults. East limit of **belt A** is eastern extent of initial foredeep deposits of Island Mountain Formation and Tripon Pass Limestone. East limit of **belt B** is western extent of lithosome SA.LaDo.St ("Sadlick sequence laminated dolomitic siltstone" lithosome of Silberling et al., 1995), characteristic of the Needle Siltstone. East limit of **belt C** is western extent of lithosome SA.La.Pk ("Sadlick sequence laminated packstone"), characteristic of the Gardison Limestone. East limit of **belt D** and west limit of **belt E** is eastern extent of lithosome SA.LaDo.St.*

***Heavy dashed line having solid thrust-fault barbs** marks the eastern extent of the Roberts Mountains allochthon; **heavy dashed lines having open barbs** are known or inferred Mesozoic thrust faults.*

***Large crosses**, localities referred to in text; **encircled large crosses**, localities that are field-trip stops (see Road Log); **small crosses**, other significant localities.*

***Locality symbols** listed alphabetically:*
AP (west of belt A), Antler Peak, Battle Mountain.
AS (central belt B), Antelope Spring, White Pine Range.
BM (south belt D), Blawn Mountain, southern Wah Wah Range.
BP (central belt A), Black Point, northern Pancake Range.
BS (central belt A), Black Shade well, northern Pancake Range.
CC (north belt A), Carlin Canyon.
CR (central belt C), Confusion Range.
DC (north belt A), Dry Canyon, Piñon Range.
DG (central belt A), Devils Gate.
DV (south belt E), Dog Valley, Pahvant Range.
DW (north belt C), northern Dugway Range.
FC (north belt A), Ferdelford Canyon, Piñon Range.
FM (north belt C), Ferguson Mountain, Goshute Mountains.
GC (central belt A), Gabel Canyon, Roberts Mountains.
GH (north belt C), Gold Hill, Deep Creek Mountains.
GM (central belt C), Granite Mountain.
GS (central belt B), Green Springs, White Pine Range.
HC (central belt B), Harris Canyon, White Pine Range.
IC (north belt C), Indian Creek, Cherry Creek Range.
IV (north belt B), Independence Valley, Pequop Mountains.
LP (north belt C), Leppy Hills.
LS (north belt D), southern Lakeside Mountains.
MH (south belt D), Mountain Home Range.
MK (central belt B), Mokomoke Mountains, White Pine Range.
MP (central belt E), Mammoth Peak, East Tintic Mountains.
OC (north belt E), Ophir Canyon, Oquirrh Mountains.
PA (north belt A), Papoose Canyon, Piñon Range.
PB (central belt A), Packer Basin, Diamond Mountains.
PQ (northeast belt A), Pequop Summit, Pequop Mountains.
RC (central belt A), Roberts Creek, Roberts Mountains.
RH (central belt C), Red Hills.
RR (south belt A), northern Reveille Range.
RS (central belt D), Rattlesnake Spur, East Tintic Mountains.
TC (central belt A), Tollhouse Canyon, Diamond Mountains.
TS (southwest belt D), Trough Spring Canyon, Egan Range.
WA (central belt A), Walters Canyon, Diamond Mountains.
WE (central belt C), Weaver Canyon, Deep Creek Mountains.
WM (central belt B), Ward Mountain, Egan Range.
WP (north belt C), White Horse Pass, Goshute Mountains.

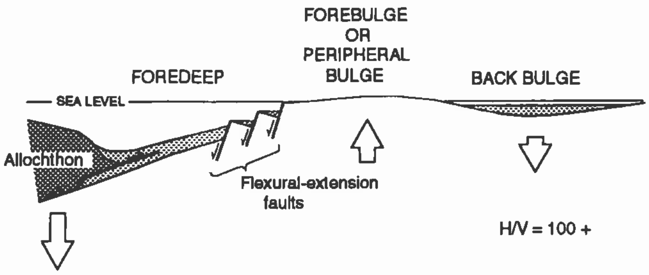


Figure 2. Diagrammatic cross section of a flexural-loading peripheral foreland system. Horizontal and vertical relative dimensions of paleotectonic features are not to scale.

the Mountain Home (or Needle) Range (MH on Fig. 1) forms the upper part of the "Joana Limestone" (Hintze, 1988) in a facies distinct from that of the correlative part of the typical Joana, which is still farther west in the vicinity of Ely, Nevada. Massive, oolitic grainstone units make up much of the lower half of the upper "Joana" in the Mountain Home Range, and the uppermost oolitic units intertongue with fine-grained, laminated, pellet-crinoid packstones like those of the Gardison Limestone, which are interpreted as tidalites in sections farther northeast in Utah.

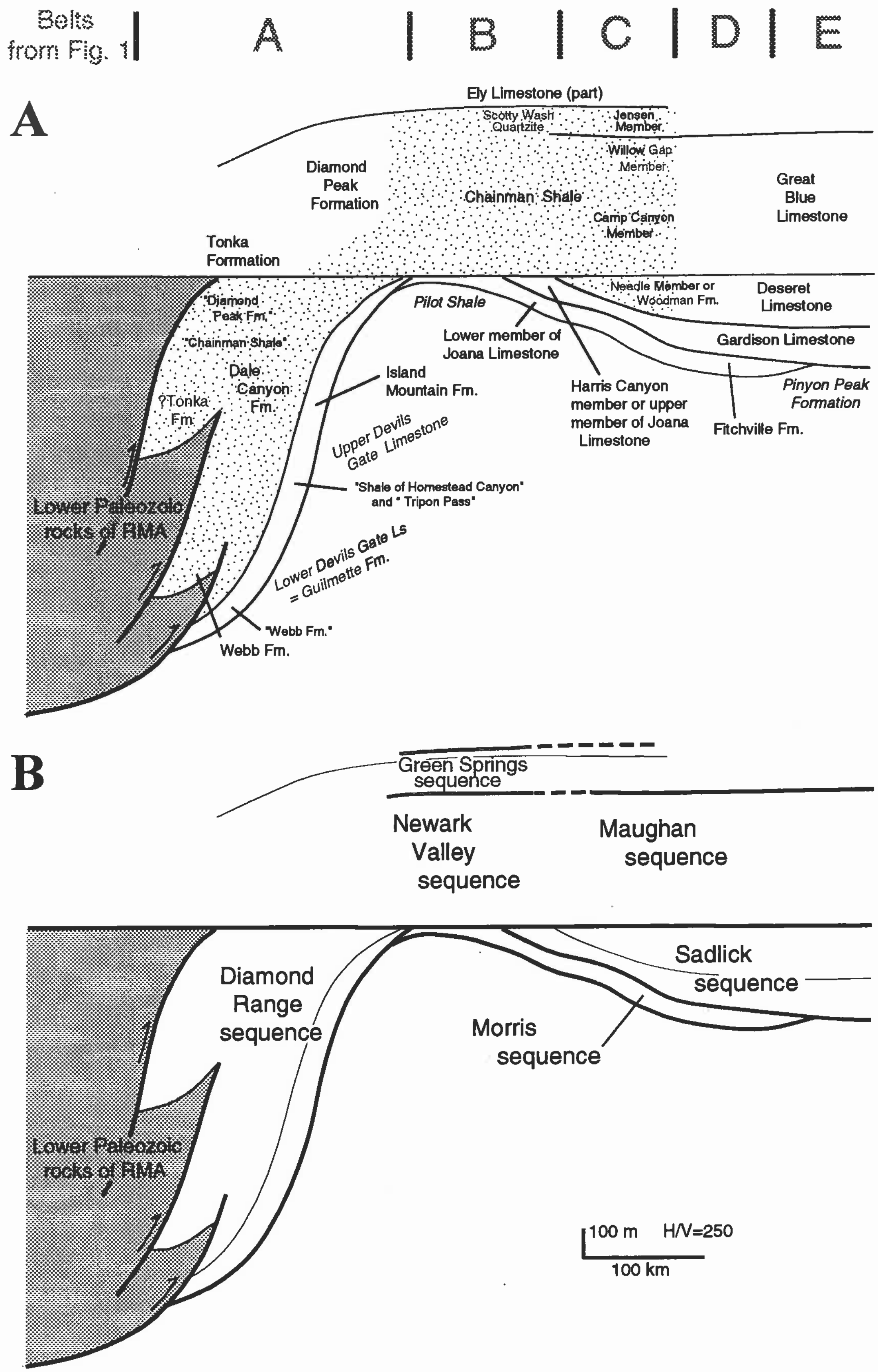
In the Confusion Range (CR on Fig. 1), in westernmost central Utah, the upper "Joana Limestone" (Hintze, 1988) (Fig. 4) is distinctly cyclic (Goebel, 1991), forming a retrogradational parasequence set (Fig. 5), composed of four parasequences. Rhythmically thin- to medium-bedded or nodular wackestone and subordinate crinoidal packstone form the lower parts of each of these parasequences and are interpreted as relatively deep-subtidal, but above storm wave base, deposits. In the lowest of these parasequences the basal, rhythmically bedded limestone grades upward into cross-stratified crinoidal and then oolitic grainstone; in the two next higher parasequences it is overlain by crinoidal grainstone and packstone, some of which exhibits wispy wave sorting; and at the top of the "Joana" the rhythmically bedded limestone is directly overlain by deposits of the Delle phosphatic event, interpreted to represent maximum flooding within the Sadlick sequence (Silberling et al., 1995). The parasequences of the upper "Joana" in the Confusion Range thus record progressive drowning and form a retrogradational parasequence set.

In easternmost Nevada within belt B, the lower part of the Sadlick sequence is entirely composed of deep subtidal limestone. This facies forms the upper part of the Joana Limestone at Ward Mountain (WM, Fig. 1; Fig. 4) near the poorly preserved type section of the Joana in the Ely mining district. It has been named the Harris Canyon member of the Joana Limestone (Crosbie, 1997), based on its exposures in the White Pine Range (HC, fig. 1). The Harris Canyon is evenly thin- to medium-bedded, commonly graded, lime mudstone, wackestone, and packstone. Fine-grained beds generally have films of siliciclastic impurities or lags of bioclastic grains such as pelmatozoan columnals on their bedding surfaces. Rare occurrences of isolated cushion-shaped coral colonies, up to 0.5 m in diameter and in growth position within some beds, indicate that water depths were within the photic zone. Limestones of the Harris Canyon are interpreted to be tempestites deposited mainly below normal wave base on the distal ramp (Goebel, 1991). In addition to the White Pine Range (at GS, HC, and AS, Fig. 1) and the Ward Mountain area (WM, Fig. 1) of the Egan Range, the Harris Canyon member of the Joana occurs in the Pequop Mountains (IV, Fig. 1), Goshute Mountains (FM, Fig. 1), and northern Deep Creek Mountains (GH, Fig. 1).

Throughout belt B the Harris Canyon member is the only rock unit representative of the Sadlick sequence; higher parts of the sequence, as represented east of belt B by the distinctive deposits of the Delle phosphatic event, are absent. Instead, Chainman Shale, representing the next younger Maughan sequence, directly overlies the Harris Canyon. In the White Pine Range (AS, MK, HC, and GS, Fig. 1) the base of the Chainman Shale is a pre-Tertiary, bedding-subparallel, attenuation fault (Crosbie, 1997). It seems improbable, however, that this is the case everywhere throughout the over 200 km-long belt in which the Chainman Shale directly overlies the Harris Canyon, so the Chainman-Harris Canyon contact is inferred to have been an unconformity.

Other "upper Joana" sections in belt C of easternmost Nevada and northwestern Utah (such as the sections at TS, RH, WE, IC, and GH, Fig. 1) are lithologically transitional between the deep-subtidal Harris Canyon member of the Joana and the cyclic deep- to shallow-subtidal upper part of the "Joana" in the Mountain Home and Confusion Ranges (CR and MH, Fig. 1). The pronounced widening of

*Figure 3. Conceptional palinspastic cross sections of the Mississippian Antler foreland for 39–40° north latitude. Cross sections are palinspastically corrected for estimated horizontal shortening on the Mississippian Roberts Mountains thrust system and the Mesozoic Central Nevada thrust belt in map belt A (Fig. 1) and the Sevier thrust system in belts D and E, and for horizontal extension related to the Tertiary Snake Range decollement in western belt C. **RMA**, Roberts Mountains allochthon. **(A)** Lithostratigraphic units. **Stipple pattern**, rocks units embraced by the Chainman Shale in a group sense. The Jensen, Willow Gap, Camp Canyon, and Needle Members were originally defined as members of the Chainman Shale. **(B)** Stratigraphic sequences (Trexler and Nichtman, 1990; Silberling at al., 1995).*


Belts from Fig. 1
A
B
C
D
E
A
Ely Limestone (part)
Scotty Wash Quartzite
Jensen Member
Willow Gap Member
Diamond Peak Formation
Chainman Shale
Camp Canyon Member
Great Blue Limestone
Tonka Formation
Needle Member or Woodman Fm.
Deseret Limestone
"Diamond Peak Fm."
Pilot Shale
"Chainman Shale"
Lower member of Joana Limestone
Gardison Limestone
Dale Canyon Fm.
Island Mountain Fm.
Harris Canyon member or upper member of Joana Limestone
Pinyon Peak Formation
?Tonka Fm.
Upper Devils Gate Limestone
Fitchville Fm.
"Shale of Homestead Canyon" and " Tripon Pass"
Lower Paleozoic rocks of RMA
Lower Devils Gate Ls = Guilmette Fm.
"Webb Fm."
Webb Fm.
B
Green Springs sequence
Newark Valley sequence
Maughan sequence
Sadlick sequence
Diamond Range sequence
Morris sequence
Lower Paleozoic rocks of RMA
100 m
H/V=250
100 km

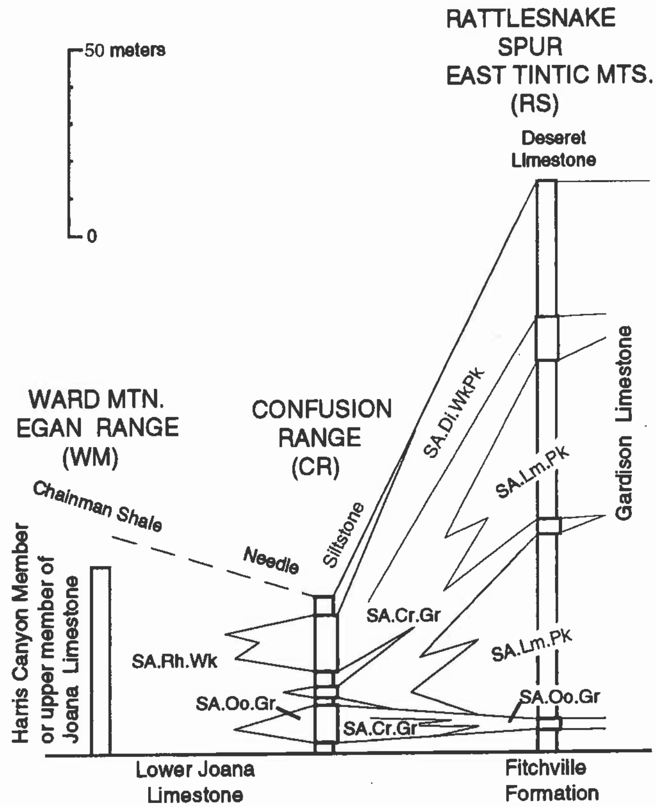


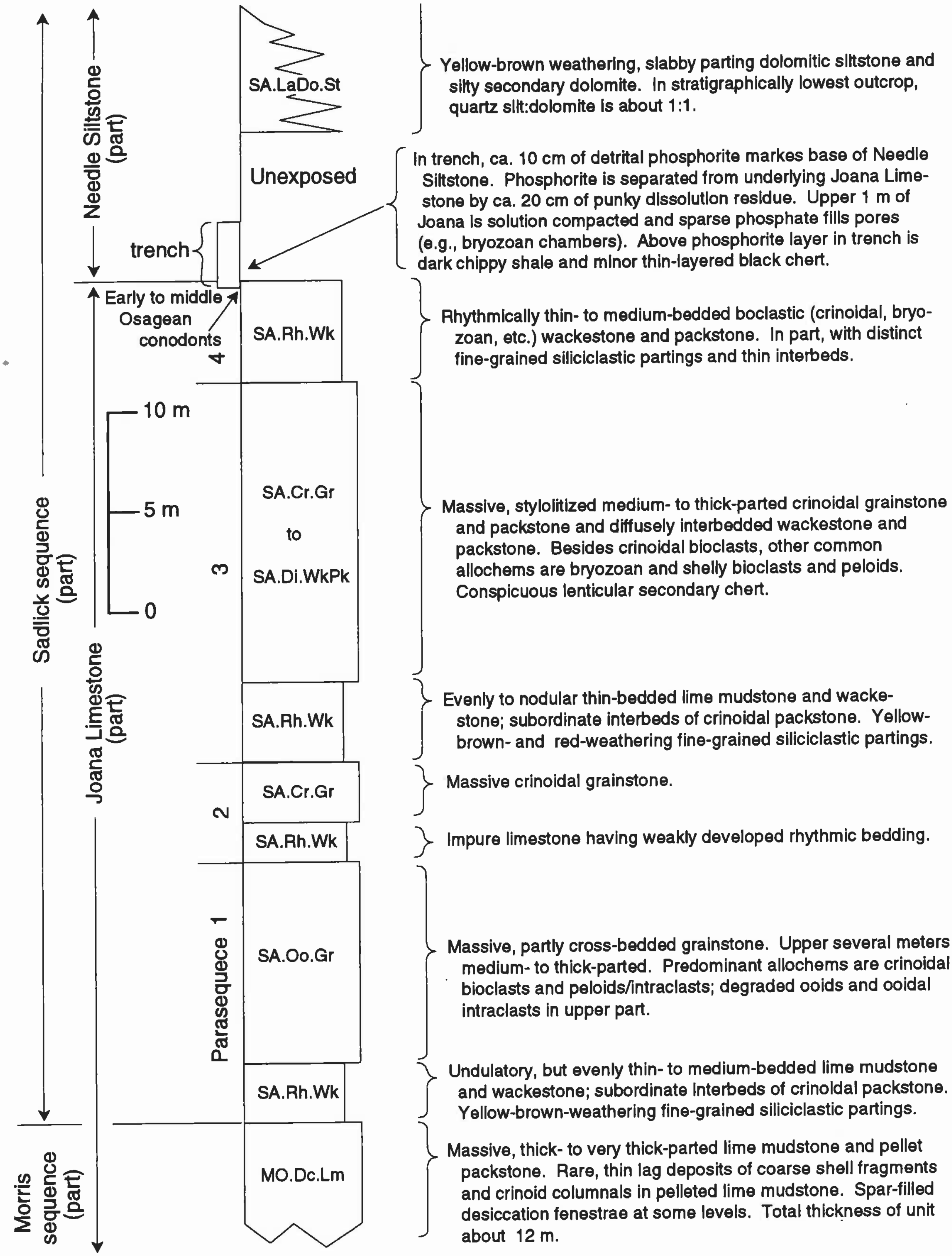
Figure 4. Stratigraphic columns of the lower Sadlick sequence showing the inferred lithic correlation of the lithosomes. Lithosome designations are from Silberling et al. (1995): SA.Rh.Wk, Rhythmically bedded wackestone; SA.Cr.Gr, Crinoidal grainstone; SA.Oo.Gr, Oolitic grainstone; SA.Di.WkPk, Diffusely interbedded wackestone and packstone; SA.Lm.Pk, Laminated packstone.

the middle part of belt C, and the stretching out of the lateral stratigraphic transition between the Cherry Creek (IC) and Confusion Range (CR) Mississippian sections between these localities, agrees well with the estimate by Gans and Miller (1983) of 250% structural extension in this part of the region related to the mid-Tertiary Northern Snake Range decollement.

The paleotectonic significance of the east-west change in the character of the lower Sadlick sequence limestones is open to interpretation. The structurally coherent outcrop belt of upper "Joana" limestone, extending from the Mountain Home to the Confusion Range, has been taken to represent the location of the Antler forebulge in early Osagean (as well as late Kinderhookian) time (Goebel, 1991; Carpenter et al., 1994; Giles and Dickinson, 1995). Evidence in support of this paleotectonic inference is not only the relative prevalence of shoal-water oolitic or crinoidal grainstones in the upper "Joana" but also the lesser thickness of this unit compared to its equivalents both to the east and west (Fig. 4). On the other hand, the overlying upper part of the Sadlick sequence reflects deposition in a topographically restricted setting (Silberling and Nichols, 1991), and points to the existence by mid-Osagean time of a topographically emergent belt located far to the west of the Confusion Range. This same emergent belt apparently also separated the depositional site of the Harris Canyon member of the Joana Limestone in belt B from that of age equivalent rocks in the foredeep of belt A (Fig. 3), and it is therefore a more dynamic candidate for a primary Osagean peripheral bulge. Consequently, although the Confusion Range section of the upper "Joana" reflects some kind of paleotectonic control on accommodation space, we favor viewing this section as primarily representing successive ramp-margin build-ups separating shallow subtidal limestones to the east from the relatively deep-subtidal limestones that were deposited to the west and farther down an irregularly subsiding depositional ramp.

Thickness changes of the "Joana" Limestone of the lower Sadlick sequence have been used by Silberling et al. (1995) to define the extent of the Early Mississippian positive area named the "Wendover high" by Sadlick (1965). However, the depositional significance of some of these changes is uncertain. The "Joana" Limestone apparently thins northward within the Confusion Range (Hintze, 1988), but it has not been determined which part of the unit—that belonging to the Morris or to the Sadlick sequence—is being thinned. Farther north, the "Joana" is not represented in the Devonian-Mississippian section at Granite Mountain (GM, Fig. 1), and it is extraordinarily thin in the Dugway Range (DW, Fig. 1), but the role of attenuation faults at these localities is uncertain. These two sections along with that in the southern Goshute Range (WP, Fig. 1) were used by Silberling et al. (1995) as part of the paleogeographic control for the Wendover high. Recent work in the southern Goshute Range (and unpublished mapping there by J.E. Welsh) demonstrates that the Joana Limestone is present locally and of normal thickness, even though it is missing over most of the area because of Late Jurassic attenuation faulting. In the Leppy Hills (LP, Fig. 1) near Wendover,

Figure 5. Stratigraphic column of the lower Sadlick sequence in the Confusion Range (SE1/4 sec. 17 and NE1/4 sec. 20, T.18 S., R.16 W.). Lithosome designations are from Silberling et al. (1995). Morris sequence: MO.Dc.Lm, Desiccated lime mudstone. Sadlick sequence: SA.Rh.Wk, Rhythmically bedded wackestone; SA.Cr.Gr, Crinoidal grainstone; SA.Di.WkPk, Diffusely interbedded wackestone and packstone; SA.Oo.Gr, Oolitic grainstone; SA.LaDo.St, Laminated dolomitic siltstone. Needle Siltstone was originally defined as a member of the Chainman Shale (Sadlick, 1965).


Sadlick sequence (part)
Needle Siltstone (part)
trench
Early to middle Osagean conodonts
Joana Limestone (part)
10 m
5 m
0
Morris sequence (part)
Parasequece 1
2
3
4
SA.LaDo.St
Unexposed
SA.Rh.Wk
SA.Cr.Gr to SA.Di.WkPk
SA.Rh.Wk
SA.Cr.Gr
SA.Rh.Wk
SA.Oo.Gr
SA.Rh.Wk
MO.Dc.Lm
Yellow-brown weathering, slabby parting dolomitic siltstone and silty secondary dolomite. In stratigraphically lowest outcrop, quartz silt:dolomite is about 1:1.
In trench, ca. 10 cm of detrital phosphorite markes base of Needle Siltstone. Phosphorite is separated from underlying Joana Limestone by ca. 20 cm of punky dissolution residue. Upper 1 m of Joana is solution compacted and sparse phosphate fills pores (e.g., bryozoan chambers). Above phosphorite layer in trench is dark chippy shale and minor thin-layered black chert.
Rhythmically thin- to medium-bedded boclastic (crinoidal, bryozoan, etc.) wackestone and packstone. In part, with distinct fine-grained siliciclastic partings and thin interbeds.
Massive, stylolitized medium- to thick-parted crinoidal grainstone and packstone and diffusely interbedded wackestone and packstone. Besides crinoidal bioclasts, other common allochems are bryozoan and shelly bioclasts and peloids. Conspicuous lenticular secondary chert.
Evenly to nodular thin-bedded lime mudstone and wackestone; subordinate interbeds of crinoidal packstone. Yellow-brown- and red-weathering fine-grained siliciclastic partings.
Massive crinoidal grainstone.
Impure limestone having weakly developed rhythmic bedding.
Massive, partly cross-bedded grainstone. Upper several meters medium- to thick-parted. Predominant allochems are crinoidal bioclasts and peloids/intraclasts; degraded ooids and ooidal intraclasts in upper part.
Undulatory, but evenly thin- to medium-bedded lime mudstone and wackestone; subordinate interbeds of crinoidal packstone. Yellow-brown-weathering fine-grained siliciclastic partings.
Massive, thick- to very thick-parted lime mudstone and pellet packstone. Rare, thin lag deposits of coarse shell fragments and crinoid columnals in pelleted lime mudstone. Spar-filled desiccation fenestrae at some levels. Total thickness of unit about 12 m.

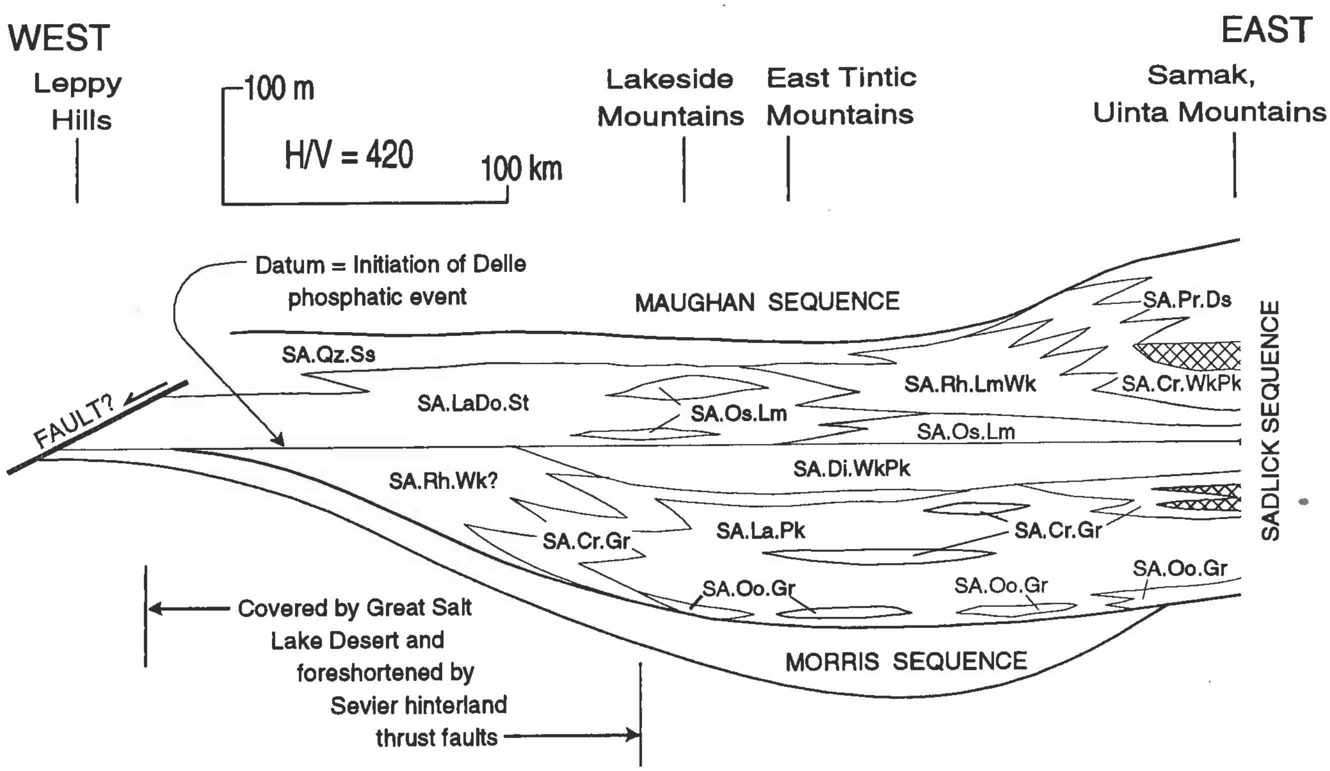


Figure 6. Diagrammatic east-west cross section of the Sadlick, Morris, and lower Maughan sequences from the Leppy Hills (LP) to the west end of the Uinta Mountains (SA) about 60 km east of Salt Lake City, with control from the southern Lakeside Mountains (LS) and East Tintic Mountains (ET, = RS and MP on Fig. 1) projected into the line of section. Palinspastically corrected for the estimated horizontal shortening across the Mesozoic Sevier thrust system. Lithosome designations for the Sadlick sequence are from Silberling et al. (1995): SA.Qz.Ss, Quartzose sandstone; SA.LaDo.St, Laminated dolomitic siltstone; SA.Rh.Wk, Rhythmically bedded wackestone; SA.Os.Lm, Ostracode lime mudstone; SA.La.Pk, Laminated packstone; SA.Cr.Gr, Crinoidal grainstone; SA.Di.WkPk, Diffusely interbedded wackestone and packstone; SA.Oo.Gr, Oolitic grainstone; SA.Rh.LmWk, Rhythmically bedded lime mudstone and packstone; SA.Cr.WkPk. Crinoidal wackestone and packstone; SA.Pr.Ds, Primary dolostone. Cross-hatch pattern, Eogenetic secondary dolostone.

however, the Wendover high is still a valid relationship, as indicated by overstepping of the lower Sadlick sequence by the upper part of the sequence (Nichols and Silberling, 1993).

Deposition of open-marine, partly skeletal limestone that forms the protracted transgressive systems tract and thick lower part of the Sadlick sequence was abruptly terminated by onset of the Delle phosphatic event, deposits of which characterize the upper part of the Sadlick sequence. Onset of the Delle phosphatic event is marked by the abrupt appearance of a thin layer of detrital phosphate. In belts C and D the upper few meters of limestone underlying this layer are conspicuously solution compacted, contain sparse pore-filling phosphate, and are separated from the basal Delle phosphate layer by up to a meter of limestone-dissolution residue (Nichols and Silberling, 1990; Silberling and Nichols, 1991; Silberling et al., 1995). Farther east, in belt E and beyond the east edge of Figure 1, the phosphate layer marking onset of the Delle event overlies and infills a hardground dissolution surface on the underlying limestone (e.g., Sandberg and Gutschick, 1984, fig. B, plate 1).

As in the lower part of the sequence, east-west lateral change is also conspicuous in the upper part of the Sadlick sequence (Fig. 6). Within belts C and D of the distal foreland the upper part of the Sadlick sequence is mainly dolomitic or calcareous siltstone and subordinate sandstone of cratonic provenance, whereas farther east in belt E, and eastward, it is mainly carbonate rock. In belts C and D, dolomitic siltstone (lithosome SA.LaDo.St of Silberling et al., 1995; Fig. 6) forms most of the Woodman Formation and the Needle Siltstone Member of the Chainman Shale, which as originally defined are essentially synonyms. Lithostratigraphic terminology has been further complicated by introduction of the name Delle Phosphatic Member for the lower Woodman and treatment of the Needle Siltstone as the next younger member of the Woodman (Sandberg and Gutschick, 1984). Although the Delle Phosphatic Member has been applied widely (e.g., Hintze, 1988) away from its type locality in the Lakeside Mountains (LS on Fig. 1), there is no consistent way to recognize its upper limit regionally, and perpetuation of this member name is not recommended. By whatever formational name, the dolomitic siltstone (lithosome SA.LaDo.St) locally is overlain by shoreface quartz and crinoid-columnal sandstone forming the top of the Sadlick sequence, and it grades eastward along the boundary between belts D and E—as at MP (Fig. 1)—into rhythmically bedded, deep-subtidal limestone (SA.Rh.LmWk) of the Tetro Member of the Deseret Limestone. Lithosome SA.Rh.LmWk is in turn overlain by shallow-subtidal to supratidal carbonate rocks (Lithosomes SA.Cr.WkPk and SA.Pr.Ds, Fig. 6) that prograded from the region to the east of Figure 1.

Lithosome SA.LaDo.St of the Sadlick sequence, composed of dolomitic siltstone and the other rock types associated with it, exemplifies the compositionally unusual rocks representing the Delle phosphatic event. In general, these rocks (1) mostly lack body-part fossils and bioclasts of benthic organisms, (2) underwent early diagenesis under suboxic to anoxic conditions (as indicated by the development of secondary phosphate, dolomite, and ammonium feldspar), (3) bear evidence of extensive diagenetic dissolution as well as silicification of some associated limestones, and (4) were partly deposited from and altered by hyposaline water (Sandberg and Gutschick, 1984; Nichols and Silberling, 1990; Silberling and Nichols, 1991; Jewell et al., 1996). These characteristics are most fully developed in belt D, as in the Lakeside Mountains (LS on Fig. 1; Fig. 7). The siltstone itself varies from being finely planar laminated, to having abundant feeding tracks on the stratification, to more rarely being current cross stratified, reflecting different levels of low oxygenation and bottom-current velocity during deposition. Compositionally, it varies from silty secondary dolomite, in which the silt is dolomite supported, to secondarily dolomitic siltstone and calcareous siltstone. In context with the evidence for episodic anoxic conditions, the secondary dolomite matrix or cement apparently resulted from reducing-environment dolomitization (Baker and Kastner, 1981) of an original, organic-rich lime mud.

Limestones of distinctly different kinds form a minor part of the siltstone lithosome. Best represented among these are the petrographically distinctive "ostracode lime mudstones" (Silberling and Nichols, 1991) in units as much as 8 m thick (lithosomes SA.Os.Lm, Fig. 6; units 13 and 16, Fig. 7). Conodonts have never been recovered from these limestones, and megafossils are effectively absent. As seen in thin section, however, ostracodes are prevalent and the only recognizable bioclasts, indicating deposition of the ostracode lime mudstones from water of abnormal salinity. Besides occurring as intercalated units within the siltstone lithosome in belts C and D, ostracode lime mudstone forms the initial deposits of the Delle phosphatic event in belt E beyond the eastern extent of the siltstone lithosome (Fig. 6). Effects of dissolution are inconspicuous in the ostracode lime mudstones, whereas other limestones, such as those forming "pinch and swell" structures in the Lakeside Mountain section (level 15, Fig. 7), are dissolution remnants of more normal marine limestone containing conodonts, radiolarians, and goniatites (Sandberg and Gutschick, 1984, who regarded these structures as "concretions"; Nichols and Silberling, 1990).

Geochemical comparison of limestones affected by the Delle event with the limestone of the widespread, pre-Delle

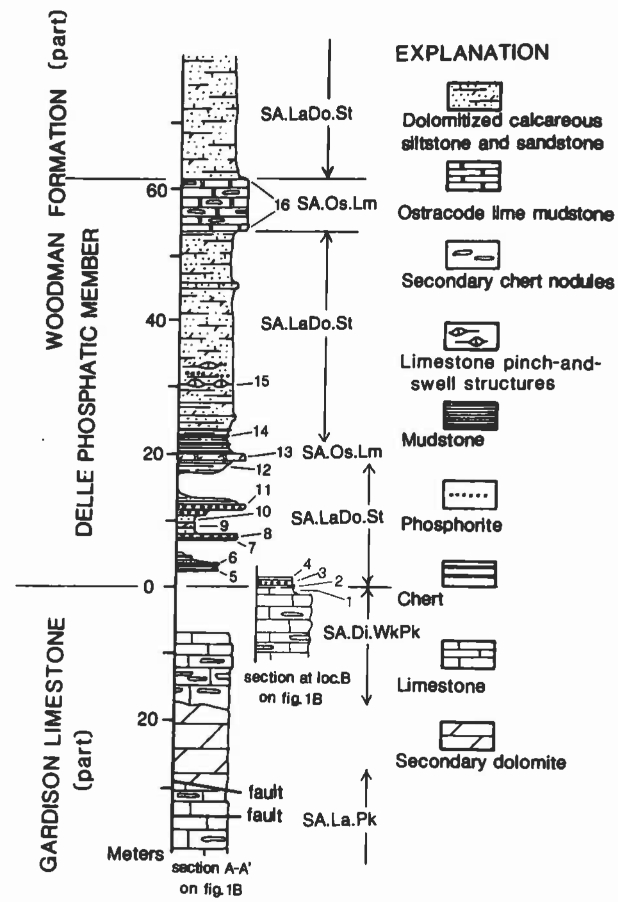


Figure 7. Stratigraphic column of the type section of the Delle Phosphatic Member and adjacent strata. Terminology of the rocks above the Gardison Limestone follows that of Sandberg and Gutschick (1984). Rocks of the Woodman Formation could equally well be assigned to the Needle Siltstone. Designations for lithosomes of the Sadlick sequence are from Silberling et al. (1995): SA.LaDo.ST, Laminated dolomitic siltstone; SA.Os.Lm, Ostracode lime mudstone; SA.Di.WkPk, Diffusely interbedded wackestone and packstone; SA.La.Pk, Laminated packstone. Numbered beds and units are those of Nichols and Silberling (1990).

event, normal marine, uppermost unit of the Gardison Limestone (lithosome SA.Di.WkPk, Figs. 6, 7) shows significantly different patterns (Fig. 8; Jewell et al., 1996). Compared regionally, the Delle-event ostracode lime mudstone (SA.Os.Lm) has substantially lighter oxygen isotope values and mostly lighter carbon isotope values than does the limestone of lithosome SA.Di.WkPk (Fig. 8). The dissolution pinch and swell limestone remnants within the siltstone lithosome (SA.LaDo.St) are similarly negative in $\delta^{18}O$ and even more negative in $\delta^{13}C$ as compared with lithosome SA.Os.Lm. Isotope values for the solution-compacted uppermost part of lithosome SA.Di.WkPk are intermediate between those of lithosomes SA.Di.WkPk and SA.Os.Lm (Fig. 8).

Covariance of carbon and oxygen isotope pairs in a manner similar to that of Figure 8 is often interpreted as an indication of postdepositional diagenesis rather than a change in depositional environment. Evidence for a significant shift in water chemistry between deposition of the SA.Di.WkPk and SA.Os.Lm lithosomes comes from Sr concentrations (200–550 ppm) which are indistinguishable between the two limestones. Unaltered normal marine limestones have significantly higher Sr concentrations. The comparable Sr concentrations of both lithosomes indicate that they have undergone similar postdepositional geochemical modification and that the different carbon and oxygen isotopic compositions represent differing depositional environments for the two units.

The change of oxygen isotope values for limestones associated with the Delle event, as compared with the immediately underlying open-marine limestone, indicates deposition of the ostracode lime mudstones from isotopically light, brackish surface water at times of decrease in water depth (Jewell et al., 1996) or possibly in response to a change in climatic conditions. At such times this hyposaline water was flushed through the normal marine limestones that occur as dissolution remnants within the siltstone lithosome or directly below it, causing their dissolution and alteration of their carbonate chemistry. The tendency towards light carbon isotope values in limestones associated with the Delle event can be explained by their incorporation of isotopically light organic carbon during diagenesis.

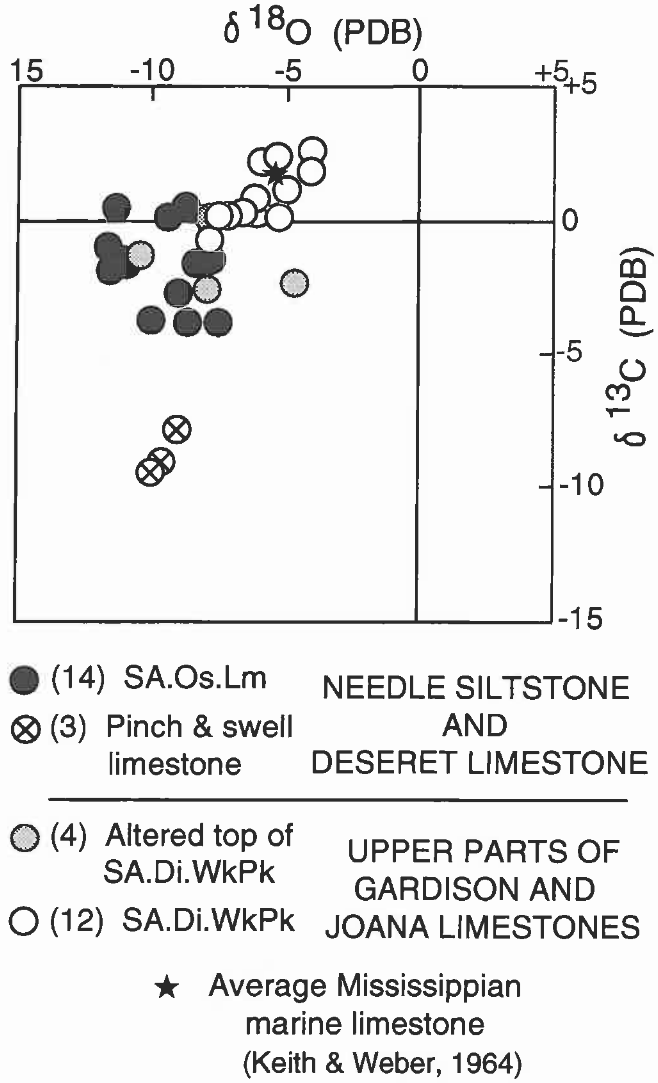


Figure 8. $\delta^{18}O$–$\delta^{13}C$ plot of Sadlick sequence limestones above and below the level marking the onset of the Delle phosphatic event. Lithosome designations are from Silberling et al. (1995): SA.Os.Lm, Ostracode lime mudstone; SA.Di.WkPk, Diffusely interbedded wackestone and packstone. The level of the "pinch & swell limestone" structures within lithosome SA.LaDo.St is plotted on figure 7.

A model of Delle-event deposition is shown in Figure 9. Deposition of silty siliciclastics was limited to the western part of the distal foreland as represented in belts C and D. Therefore, both the fresh water and silt that distinguish Delle-event sedimentation evidently had the same fluviatile sources from the west or north within an emergent belt B or an original northern extension of belt B. In this model, the strong pycnocline beneath the hyposaline upper water layer enhances episodes of oxygen-deficiency in the bottom waters, explaining the paucity of shelly fossils in these rocks and the production of diagenetic secondary dolomite, phosphate, and ammonium feldspar. Times of increased runoff of fresh water and silty sediment resulted in both thickening of the hyposaline wedge of water and decreasing the water depth by increasing the rate of silt deposition. In effect, this brought the bottom up into the hyposaline water mass, resulting in deposition of the ostracode lime mudstone units. Evidence for this is the occurrence of ostracode lime mudstone units (such as unit 13, Fig. 7, in the Lakeside Mountains section) at the tops of

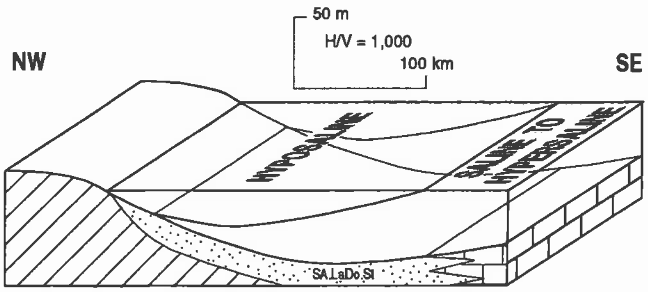


Figure 9. Model of depositional conditions during the Delle phosphatic event in the distal Antler foreland, as discussed in text.

shallowing-upward parasequences. Such parasequences show a shallowing and freshening upward transition from chert and phosphate, to planar laminated siltstone, to sparsely fossiliferous siltstone, and finally to ostracode lime mudstone. Some of the fine-grained siliciclastics and original lime mud may have run out onto the pycnocline surface by detached flow or been episodically flushed into the depositional site during periods of high precipitation. This would allow them to have been widely distributed and then deposited from suspension, which could account for the planar lamination of some of the siltstone lithosome. The silt of the lithosome SA.LmDo.St is compositionally the same as that of siltstones in the Upper Devonian Pilot Shale (Nichols and Silberling, unpublished petrographic data), which at least locally was emergent and exposed to erosion during the mid-Mississippian along belt B and perhaps also along an original northern extension of it. Along with the evidence for hyposalinity, the retention of fine-grained, ultimately craton-sourced siliciclastic sediment along the western side of the distal foreland is indicative of topographic restriction of the distal foreland during deposition of the upper Sadlick sequence.

Maughan sequence

Throughout the distal Antler foreland the contact between the Sadlick and the overlying Maughan sequence is abrupt and marks upward change to relatively deeper water or more open-marine deposition. The Maughan sequence includes one or more subordinate stratigraphic sequences of 3rd-order rank, and it is thus a 2nd-order sequence. Goniatite faunas from near the base of the Maughan in different places are equivalent to the upper part of Mamet foraminiferal zone 15 or zone 16i and are thus uppermost Meramecian to earliest Chesterian in age (Alan Titus, personal communication, 1995). The top of the Maughan sequence approximates the foraminiferal zone 18–19 boundary within the upper Chesterian, again based on goniatite correlations by Alan Titus (personal communication, 1995). Both the bottom and top of the sequence thus correlate with major eustatic lows in the long-term sea-level record of Ross and Ross (1987).

In general, the Maughan sequence grades westward across the distal foreland from sections in the northern parts of belt E, D, and C that are predominantly of shelf limestone (such as the Great Blue Limestone), to sections in belt B that are primarily fine-grained siliciclastic rocks (such as the Chainman Shale). Intermediate sections are mixtures of bioclastic limestone, impure, commonly abundantly spiculitic limestone, pelitic rocks, and subordinate quartzose sandstones. Owing to the relatively great thicknesses of the Maughan sequence, ranging from several hundred to well over 1,000 m in belt E, and to structural complexities in most of its outcrop areas, its regional sedimentology and internal sequence stratigraphy in the most distal parts of the Antler foreland are as yet poorly known. The eastward thickness increase in belts D and E may presage abnormal late Paleozoic subsidence rates, associated with deposition of the Oquirrh Formation in central Utah and not necessarily be related to tectonism along the Pacific margin.

One of the most coherent and instructive sections of the Maughan sequence is that in the Confusion Range (CR, Fig. 1). This section, which is nearly 450 m thick, is shown on Figure 10. It includes in its lower 100 m a well-expressed subordinate (3rd-order) sequence that progrades upward from restricted black shale or mudstone, through lagoonal calcisphere-endothyrid limestone, to open-marine, crinoid-rich massive limestone that forms its top. Above this subordinate sequence, the thick upper part of the Maughan sequence in the Confusion Range progrades again from black shale and spiculitic, evenly-bedded limestone up to thick-bedded, open-marine, bioclastic limestone at the top of the sequence. The occurrence of a subordinate-rank sequence within the lower part of the Maughan sequence in this well-dated section helps explain the complicated stratigraphic relations and the varying usages of rock-unit names such as the Uncle Joe and Tetro Members of the Deseret Limestone and the Humbug Formation in sections farther east in Utah. There, correlations and placement of the base of the Maughan sequence are uncertain. Recognition of the base of the Maughan sequence on the basis of physical stratigraphy in these sections is complicated by the expression of lower-order sequences or parasequences in both the lower Maughan and upper Sadlick sequences.

West of the Confusion Range, in the more southern parts of belts C and D, the Maughan sequence (or its protolith where metamorphosed) is mainly dark shale or mudstone and subordinate amounts of cross-stratified quartzose sandstone. Depositional environments range from marine to paralic, parts of the section being coal-bearing. Pelitic rocks of the Maughan sequence are poorly exposed at the type locality of the Chainman Shale in the Ely mining district. Here, and at most other localities, poor exposure and known or probable structural disruption of the predominantly pelitic section make detailed stratigraphic study difficult.

Above the Maughan sequence in the Confusion Range is a distinct third-order sequence that is evidently the equivalent of the Green Spring sequence (Fig. 3B), as typified in the White Pine Range (GS and MK, Fig. 1). In the Confusion Range the Green Springs sequence extends upward into the lower part of the Pennsylvanian-Mississippian Ely Limestone and is of latest Chesterian age (Mamet foraminiferal zone 19). Beneath the Ely Limestone, strata of the Green Springs sequence in the Confusion Range are mainly terrigenous-clastic rocks and fossiliferous impure limestone of the Jensen Member, originally described as the uppermost member of the Chainman Shale (Sadlick, 1965). The Maughan-Green Springs sequence boundary is marked by limestone dissolution and a layer of phosphorite (Fig. 10). The Scotty Wash Quartzite, recognized at localities in eastern Nevada (such as WP, IC, WM, and TS, Fig. 1) by the appearance of thick, nearshore quartzite units of cratonic provenance in black shale successions, is mainly part of the Green Springs sequence. Some "Scotty Wash" quartzites, however, may occur in the upper part of Maughan sequence.

PROXIMAL MISSISSIPPIAN FORELAND

Proximal terrigenous-clastic deposits of the Mississippian foreland occur in belt A, which extends from the break-out of the Mississippian Roberts Mountains thrust system eastward to the eastern limit of the distinctive initial Mississippian foredeep deposits (Fig. 1). In the break-out zone of the thrust, imbricate foredeep and "wedge-top" (terminology of DeCelles and Giles, 1996) proximal foredeep deposits are represented from the Roberts Mountains (RC, Fig. 1) northward through the Sulphur Spring Range to the Piñon Range (FC and CC, Fig. 1).

Historically, all of these rocks have been designated the "Chainman Shale" and the Diamond Peak Formation (Fig 3A), which generally correspond respectively to the Antler "flysch" and "molasse" of some authors. Coarse clastics in both units are clearly derived from the Roberts Mountains Allochthon (RMA). Because the Diamond Peak at its type locality in the Diamond Mountains south of locality WA (Fig. 1) is notably more sandy and conglomeratic than are the underlying strata referred to the Chainman Shale, the base of the Diamond Peak in other exposures has commonly been drawn beneath the lowest conspicuously coarse-clastic rocks in the section, at least implicitly assuming an interdigitating, time-transgressive formational boundary. Other authors have despaired of separating the two formations and lumped them into an undifferentiated Chainman-Diamond Peak unit, particularly in areas of uncertain structure (e.g., Smith and Ketner, 1978). Viewing the contact as everywhere gradational, Harbaugh and Dickinson (1981) interpreted the entire "undifferentiated Chainman Shale and Diamond Peak Formation" interval as a retrogradational succession of basin-slope and submarine-fan deposits overlain by a progradational succession of delta-slope and delta-platform deposits. The age span of this submarine-fan to deltaic complex would be from Kinderhookian through Chesterian (e.g., Poole and Sandberg, 1991), more than 30 m.y., or nearly all of Mississippian time.

A new approach to the Chainman-Diamond Peak problem was made possible by recognition of a low-angle unconformity within the Diamond Mountains section at a level low in the Diamond Peak Formation (Trexler and Nichtman, 1990; Trexler and Cashman, 1991). This unconformity separates different facies systems over a wide area, and on this basis the rocks within the Diamond Peak/ Chainman section below the unconformity were designated the Diamond Range sequence and those above it the Newark Valley sequence. Palynomorph fossils from the Diamond Range sequence in belt A in the vicinity of Eureka, Nevada (Fig. 1), are of Osagean through middle Meramecian age (Trexler et al., 1995), equivalent to the age span of the Sadlick sequence in the distal part of the Antler foreland (belts D–E, Fig. 1; Fig. 3). Fossils (including foraminifers and goniatites) from near the base of the Newark Valley sequence have ages equivalent to Mamet foraminiferal zone 16 of earliest Chesterian age (Trexler and Nichtman, 1990; Alan Titus, personal communication, 1995). Thus the base of the Newark Valley sequence is permissibly correlative with that of the Maughan sequence, as recognized in the distal foreland (Fig. 3). The upper limit of the Newark Valley, as originally defined, was the base of the Green Springs sequence, named by Trexler and Nichtman (1990) for strata correlated with the Diamond Peak Formation and overlying strata in the lower Ely Limestone in the White Pine Range (loc. GS, Fig. 1). The Green Springs-Newark Valley sequence boundary was originally dated as being in or below foraminiferal zone 18 (mid Chesterian). Subsequently, strata of latest Chesterian, or even early Pennsylvanian age were included in the sequence (Trexler and Cashman, 1991; Trexler et al., 1995; Perry, 1995), extending it to the top of the Diamond Peak/ Chainman group and even into the Ely Limestone. Further study of the sequence stratigraphy of the White Pine Range (Crosbie, 1997) now places the upper boundary of the Newark Valley sequence within the late Chesterian (see the Road Log for Stop 10). The still younger Green Springs sequence of Trexler and Nichtman (1990) contains goniatites equivalent in age to Mamet foraminiferal zone 19 (Alan Titus, personal communication, 1997) near its base,

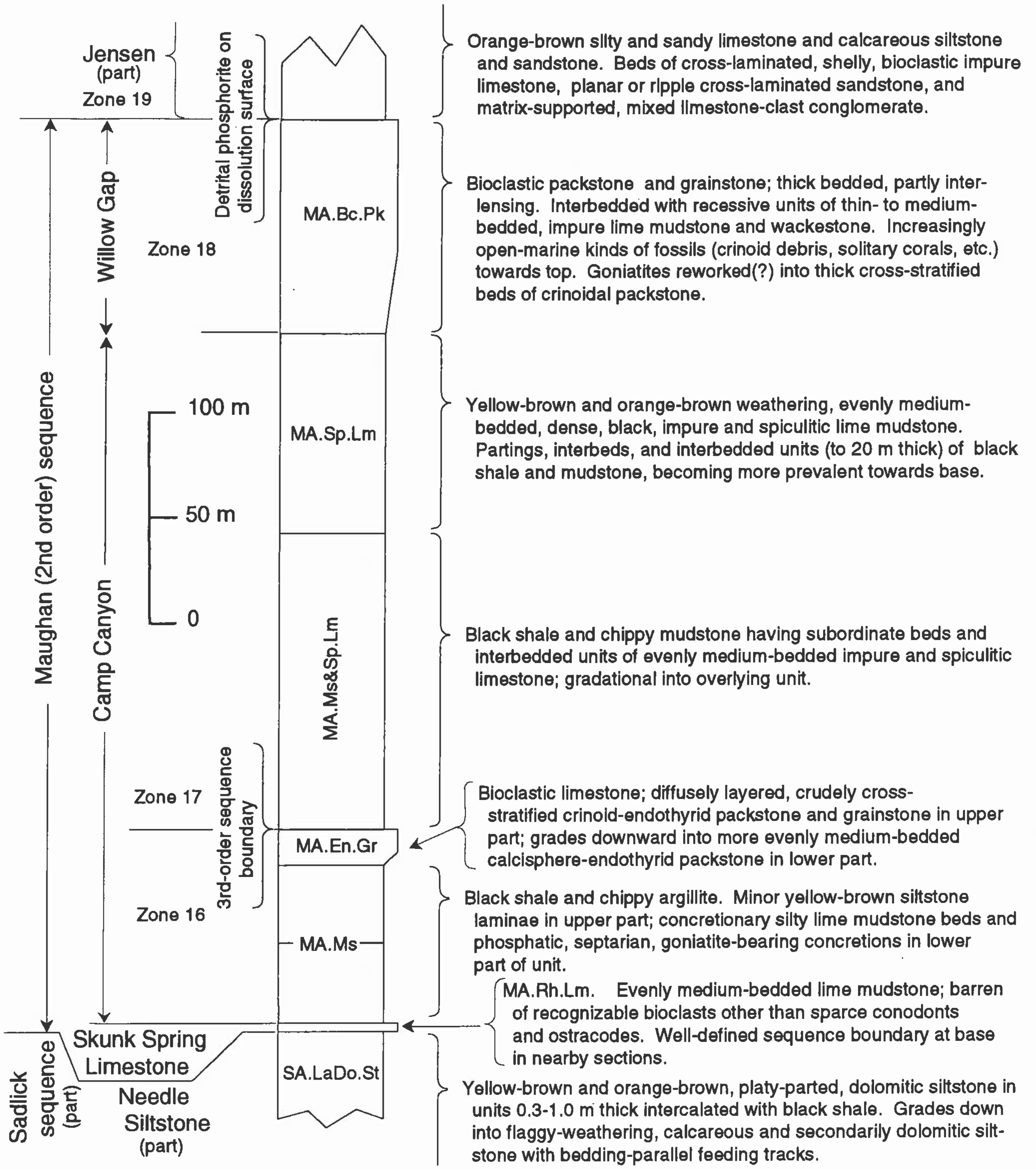


Figure 10. Stratigraphic column of the Maughan sequence in the Confusion Range (W1/2 sec. 4, T.18 S., R.16 W.). Lithosome designations are from Silberling et al. (1995). Maughan sequence: MA.Bc.Pk, Bioclastic packstone; MA.Sp.Lm, Spiculitic lime mudstone; MA.Ms&Sp.Lm, Mudstone and spiculitic lime mudstone; MA.En.Gr, Endothyrid grainstone; MA.Ms, Mudstone; MA.Rh.Lm, Rhythmically bedded lime mudstone. Sadlick sequence: SA.LaDo.St, Laminated dolomitic siltstone. Numbered zones indicate correlations based on goniatites by Alan Titus (personal communication, 1995) with the Mamet foraminiferal zones reported by Webster et al. (1984) from Granite Mountain (GM, Fig. 1). The Jensen, Willow Gap, Camp Canyon, Skunk Spring Limestone, and Needle Siltstone were all originally defined as members of the Chainman Shale.

which is about 400 m below the base of the Ely Limestone. A relatively thin additional sequence is provisionally recognized by Crosbie (1997) in the White Pine Range between the Newark Valley/Maughan and the Green Springs sequences. Alternatively, these strata could be interpreted to be the uppermost part of the Newark Valley/Maughan sequence highstand systems tract, making the Maughan a synonym of the Newark Valley sequence.

Recognition of the Diamond Range-Newark Valley sequence boundary within the Diamond Peak/Chainman complex creates a problem in lithostratigraphic terminology, because the pelitic rocks of the "Chainman Shale" in the proximal foreland are evidently wholly older than, and may not have been genetically continuous with, the typical Chainman Shale farther east (Fig. 3). Because "Chainman Shales" in both usages of the name are potential source rocks for fossil-fuels, this is not only a semantic problem but also has economic importance (Trexler et al., 1995). To avoid the ambiguous usage of "Chainman Shale," we recommend elevating this name to group rank and applying Dale Canyon Formation to all of the clastic rocks of the Chainman group that form the Diamond Range sequence in the proximal foreland. The Dale Canyon Formation was introduced in the vicinity of Eureka, Nevada, by Nolan et al. (1974) for "coarse grit, with some . . . fine conglomerates, and . . . black shale" lithologically similar to rocks of the overlying "Chainman Shale" and grading up into them. The upper limit of the Dale Canyon Formation is here extended up section to include the "Chainman Shale" of the Diamond Range sequence (Fig. 3). The name Dale Canyon has also been applied in the Piñon and Sulphur Spring Ranges (in the vicinity of DC, Fig. 1) by Carlisle and Nelson (1990) and Johnson and Visconti (1992) to genetically similar, coeval rocks.

The systematically folded type "section" of the Dale Canyon at Packer Basin (PB, Fig. 1) in the southern Diamond Mountains is conformably underlain by a relatively thin unit (about 50 m thick) that was originally referred by Nolan et al. (1974) to the "Joana Limestone," following the then common practice of assigning all limey rocks between the Devonian Pilot Shale and the Diamond Peak/Chainman clastics to the Joana. Rocks of this unit, recently re-named the Island Mountain Formation (Nichols and Silberling, 1995), consist of pelagic-hemipelagic spiculitic siliceous argillite, spiculitic and radiolarian lime mudstone, and distinctive matrix-supported gravity flows of bioclast-lithoclast grainstone to coarse sedimentary breccia. These strata form the basal part of the Diamond Range sequence and have been interpreted as foredeep lag deposits formed during the initial subsidence of the foredeep prior to the time when the encroaching RMA became the source of the more voluminous clastics of the Dale Canyon Formation (Nichols and Silberling, 1995). The Island Mountain Formation was defined to include the initial foredeep deposits assigned by different authors to the "Joana Limestone," "Camp Canyon equivalent," "Webb Formation," Tripon Pass Limestone, shale of Homestead Canyon, Davis Spring Formation, and Kinkead Spring Limestone from the northern Piñon Range (near FC, Fig. 1) south to the Reveille Range (RR, Fig. 1), all along belt A of the foreland. The true lower member of the Joana Limestone, that forms the Morris sequence in belts B and eastward in the foreland, can be traced into the eastern part of belt A in the southern Diamond Mountains and northern Pancake Range, where it is progressively cut out to the west and then missing beneath the Island Mountain Formation, as shown on Figure 11. Lithic clasts in gravity-flow sedimentary breccia and turbidite units in the Island Mountain Formation are derived not only from the Joana but also from the underlying Devonian Pilot Shale and Guilmette Limestone. Current directions from flute casts and cross beds associated with the gravity-flow units are generally southward, parallel to the foredeep trend (Goebel, 1991; Nichols and Silberling, 1995). The scarps of inferred flexural-extension faults along the distal side of the foredeep were proposed by Nichols and Silberling (1995) as a possible source for the coarse lithic debris deposited in the Island Mountain.

Conodonts obtained from gravity-flow units in the Island Mountain Formation are mostly late Kinderhookian forms of the same age as those from the unconformably underlying lower member of the Joana Limestone, from which these conodonts apparently were redeposited along with pre-Mississippian forms. The youngest conodonts in the Island Mountain are early Osagean, as are other kinds of fossils from localities south of Eureka (Nichols and Silberling, 1995). Reports of Kinderhookian fossils from the Dale Canyon or "Chainman Shale" are equivocal. Conodonts from the Dale Canyon are reported by Johnson and Visconti (1990) at two sections. Those from Dry Canyon (DC, Fig. 1) are actually from the Island Mountain Formation and listed as including early Osagean as well as late Kinderhookian forms, and those reported from Devils Gate (DG, Fig. 1) have never been documented. The supposedly Kinderhookian ammonites reported from the Piñon Range (near FC, Fig. 1) and the southern Diamond Mountains (near TC, Fig. 1) (Johnson and Visconti, 1990; Poole and Sandberg, 1991) are mostly undifferentiated, potentially long ranging forms tentatively identified as *Protocanites lyoni*. This species and a new subgenus of pericyclid ammonites from the Dale Canyon Formation near Papoose Canyon (PC, Fig. 1) in the Piñon Range were dated as late Kinderhookian by Gordon (1986), but as determined by Whitaker (1985) the stratigraphic occurrence is well above early Osagean conodonts collected by him from near the base of the formation.

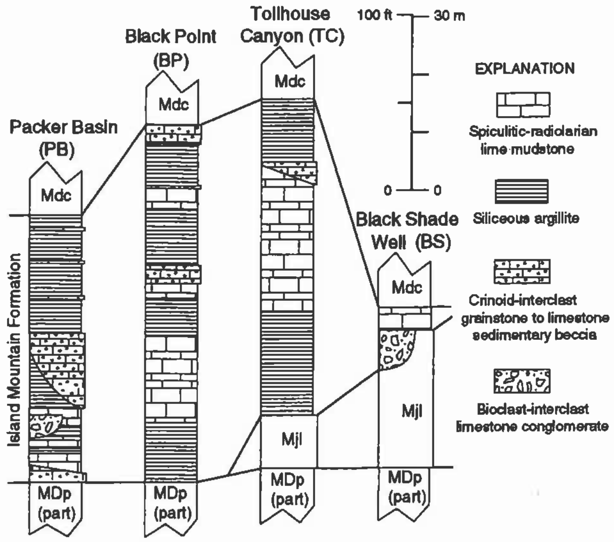


*Figure 11. Stratigraphic columns of the Island Mountain Formation from Packer Basin (**PB**), Black Point (**BP**), Tollhouse Canyon (**TC**), and Black Shade Well (**BS**) (see Fig. 1). Mdc, Dale Canyon Formation; Mjl, lower Joana Limestone; MDp, Pilot Shale. Modified from Nichols and Silberling (1995).*

The angular discordance and boundary separating the Diamond Range and Newark Valley sequences (Trexler and Nichtman, 1991) was interpreted by Giles and Dickinson (1995) as simply the toplap contact formed as braid-delta platform deposits prograded across a delta slope. The wide extent of this relationship, however, favors its original interpretation as an unconformity. It is almost continuously traceable from north of Walters Canyon (WA, Fig. 1) in the Diamond Mountains southward into the Pancake Range, a distance of about 75 km. Strata closely above the contact are everywhere roughly the same age within foraminiferal zone 16. Even more compelling is the strong angular unconformity beneath strata of this age in the Piñon Range, as at localities FC and PA (Fig. 1).

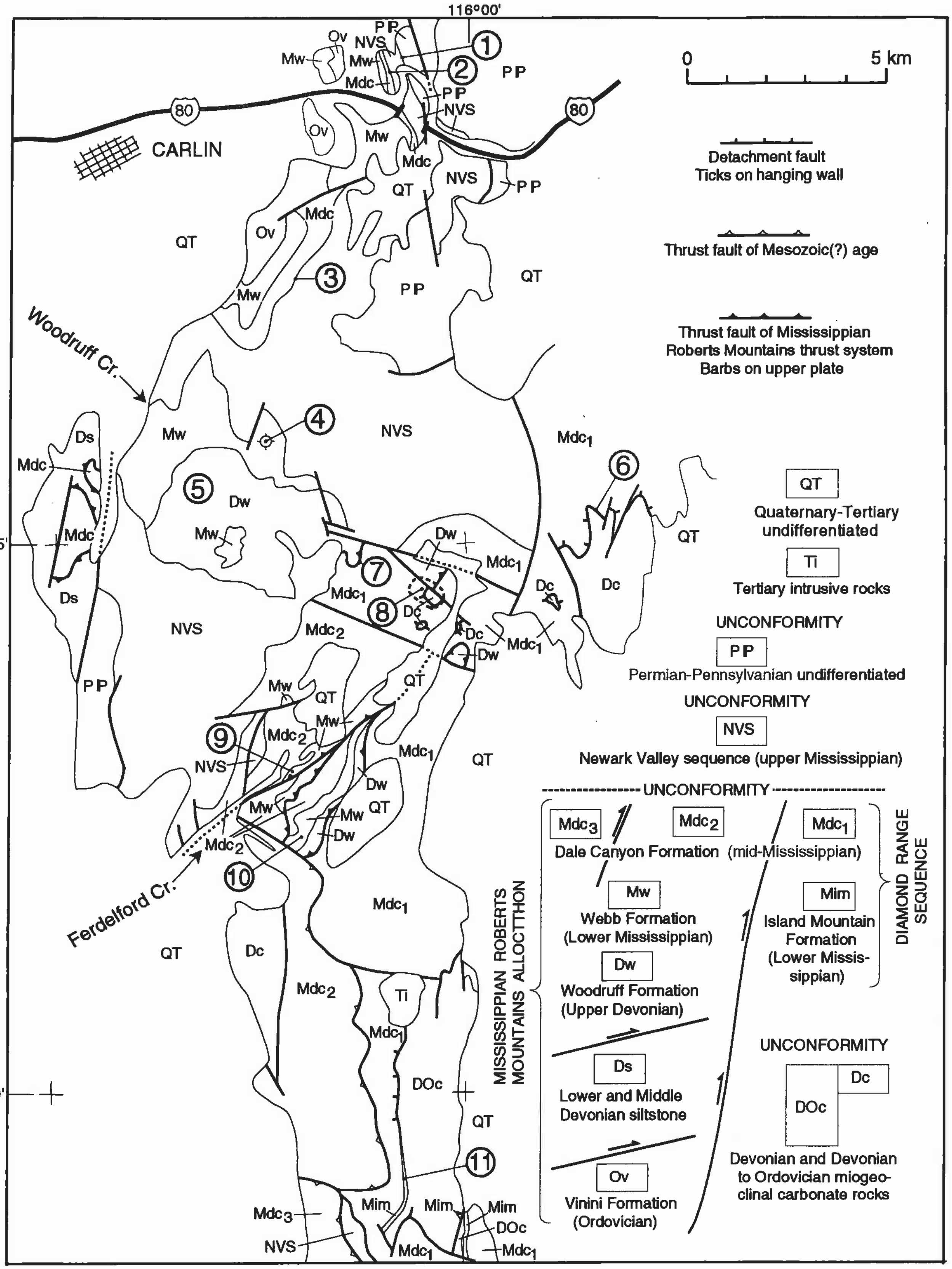
Piñon Range

Exposures in the Piñon Range (CC, FC, and PC, Fig. 1) provide the only known example in the eastern Great Basin of overlap by Mississippian strata of structures that can be related to Mississippian Roberts Mountains thrusting. They also show that thrusting was distributed among several imbricate slices involving mid-Mississippian foredeep strata, and they provide evidence that Roberts Mountains thrusting may have been polyphase, the final mid-Mississippian episode of thrusting having been preceded by one or more Late Devonian episods. Because of their importance to the understanding of the Antler foreland, relationships in this area are described here in some detail.

Present knowledge of the complex geology of the Piñon Range results from a number of different comprehensive studies, starting with that by Dott (1955). A major USGS investigation from the mid 1950's through the 1960's culminated in the regional geologic map of the range by Smith and Ketner (1978), which has served as the basis for all subsequent work in the Piñon Range. Field studies carried out at about the same time in the Sulphur Springs Range (at and south of DC, Fig. 1), into which the geology of the Piñon range merges to the south, was eventually published by Carlisle and Nelson (1990). Other important contributions include those by J.G. Johnson and his students (Johnson and Pendergast, 1981; Whitaker, 1985; and Johnson and Visconti, 1992), work by Newmont Exploration Limited in conjunction with development of the Rain mine and nearby properties (Thoreson, 1991; Putnam and Henriques, 1991), and new interpretation by the Mobil Oil Corporation (Carpenter et al., 1993;) leading to the drilling in 1990 of the 9050 ft deep Petan Trust well, which, though noncommercial, provides singular subsurface control. Of particular value is the characterization by Jansma (1988) of structural fabrics associated with Roberts Mountains thrusting.

Arguments about which, if any, Mississippian strata overlap Roberts Mountains thrusting in the Piñon Range continue to the present time. The first to address these relationships was Dott (1955), whose contribution to this argument is sometimes overlooked, because it was not the focus of his paper. Because usages of the name "Diamond Peak Formation" were so varied, Dott (1955) established the Tonka Formation as its probable local equivalent in Carlin Canyon (CC, Fig. 1), and he interpreted the Tonka to have been deposited across the Paleozoic thrust beneath the highly deformed, overthrust Ordovician rocks in the area. The Tonka was not closely dated within the Mississippian by Dott, but the lowest fossiliferous rocks within it are now known to belong to Mamet foraminiferal zone 16 (Trexler, 1990). Most of the Tonka therefore represents the Newark Valley sequence which is thus established as younger than Roberts Mountains thrusting.

The relationship of the Tonka to the Roberts Mountains allochthon in Carlin Canyon is complicated, however, by the discovery of an angular unconformity (loc. 2, Fig.12) within the conglomeratic lower part of the Tonka. This unconformity is beneath the level of the lowest known fossils in the Tonka. Conglomerate clasts in the upper Tonka comprise resistant, mineralogically stable, rock types, while the lower Tonka contains appreciable amounts of unstable lithic clasts such as limestone and arkosic sandstone (Trexler and Silberling, unpublished data). The lower Tonka apparently was deposited on the RMA and actually represents the Diamond Range rather than the Newark Valley sequence. On Figure 12 it is reassigned to the Dale Canyon


116°00'
0
5 km
CARLIN
80
Woodruff Cr.
Ferdelford Cr.
40°37.5'
40°30'
Detachment fault
Ticks on hanging wall
Thrust fault of Mesozoic(?) age
Thrust fault of Mississippian
Roberts Mountains thrust system
Barbs on upper plate
QT
Quaternary-Tertiary
undifferentiated
Ti
Tertiary intrusive rocks
UNCONFORMITY
PP
Permian-Pennsylvanian undifferentiated
UNCONFORMITY
NVS
Newark Valley sequence (upper Mississippian)
UNCONFORMITY
Mdc3
Mdc2
Mdc1
Dale Canyon Formation (mid-Mississippian)
Mw
Webb Formation
(Lower Mississippian)
Mim
Island Mountain
Formation
(Lower Missis-
sippian)
DIAMOND RANGE
SEQUENCE
Dw
Woodruff Formation
(Upper Devonian)
MISSISSIPPIAN ROBERTS
MOUNTAINS ALLOCTTHON
Ds
Lower and Middle
Devonian siltstone
UNCONFORMITY
Dc
DOc
Devonian and Devonian
to Ordovician miogeo-
clinal carbonate rocks
Ov
Vinini Formation
(Ordovician)

Formation. A short distance south of Carlin Canyon along the west side of the Piñon Range, Smith and Ketner (1978) show a similar relationship at locality 3 (Fig. 12). Here, their "Diamond Peak Formation" (the equivalent of most of the Tonka) lies depositionally on an unfaulted downward succession of "Chainman Shale," "Webb Formation," and the RMA. Unfortunately, exposures in this part of the range are too poor to preclude other possible interpretations.

Farther south in the Piñon Range in the vicinity of Ferdelford Canyon, Smith and Ketner (1968, 1978) assigned siliceous mudstone that conformably underlies the "Chainman" to the Webb Formation, and they interpreted the Webb to depositionally overlie both autochthonous Devonian carbonate rocks and allochthonous Late Devonian rocks of the Woodruff Formation carried on a pre-Webb thrust (i.e., part of the Roberts Mountains thrust system). From this they concluded that the Roberts Mountains thrust was pre-late Kinderhookian, the age of conodonts reported from the Webb (Smith and Ketner, 1968). Subsequently, Ketner and Smith (1982) recognized that the highly deformed, allochthonous "type" Webb (loc. 10, Fig. 12) is lithologically different than the "Webb" (i.e., the Island Mountain Formation) deposited on nearby autochthonous Devonian carbonate rocks (as at loc. 11, Fig. 12). The thrust that juxtaposes the type Webb and the Island Mountain Formation is therefore that beneath the Woodruff Formation underlying the type Webb (loc. 10, Fig. 12). This thrust, which overrides a considerable structural thickness of autochthonous Dale Canyon Formation, would thus be mid-Mississippian age or younger. Actually, the overthrust relationship of the RMA to "Chainman Shale" strata that would now be assigned to the Dale Canyon Formation in the Devils Gate area (DG, Fig. 1) had been reported many years before by Roberts et al. (1958), who attributed this observation to C.W. Merriam.

Johnson and Pendergast (1981) enlarged on the reinterpretation of the Webb Formation and concluded that the RMA everywhere from the Piñon Range southward to Devils Gate (DG, Fig. 1) was thrust over autochthonous "Chainman Shale" (i.e., Dale Canyon Formation), thought by them to be as old as Kinderhookian. Additionally, they recognized that some "Diamond Peak/Chainman" strata of early Osagean age in the Piñon Range (near loc. PC, fig. 1) rest depositionally on pre-Mississippian rocks of the RMA. These strata were assumed by them to post-date Roberts Mountains thrusting, and on this basis they assigned an early Osagean age to the Roberts Mountains thrust. Carpenter et al. (1993, 1994) generally followed this reasoning but on the basis of new age determinations assigned the age of RMA emplacement as being between the early Osagean and early Meramecian. In their reinterpretation of Smith and Ketner's map, they recognized that the Tonka Formation (of the Newark Valley sequence), which nearly surrounds the Webb and Woodruff Formations in Woodruff Canyon (in the vicinity of loc. 5, Fig. 12), is for the most part depositional, rather than faulted, on the Webb and Woodruff, as shown on Figure 12.

Figure 12. Generalized geologic map of the northern Piñon Range modified from Smith and Ketner (1978). Encircled numbers designate localities referred to in text.

Structural studies such as those by Jansma (1988) form the basis for relating Antler-age map-scale faults and Mississippian stratigraphy in the Piñon Range. Among the study areas where she characterized the structural style and fabrics associated with Roberts Mountain thrusting is the area in Ferdelford Canyon in the vicinity of localities 8 and 9 (Fig. 12) and that in Woodruff Creek (loc. 5, Fig. 12). Our structural data in these areas only serves to confirm her analysis of the strain history, some of which is shown on Figure 13.

We view the map pattern in Ferdelford Canyon, however, as a series of Roberts Mountains thrust imbricates involving partial successions of the Woodruff, Webb, and Dale Canyon Formations (Fig. 12), rather than as a window of Webb and Woodruff ("preflysch") beneath a decollement carrying the Dale Canyon ("flysch" or "Chainman Shale") (Jansma, 1988; Jansma and Speed, 1993). In Ferdelford Canyon mesoscopic structures include polyphase, coaxial folds in a gradient of generally increasing strain stratigraphically downward toward the imbricate thrust surfaces. Projected stereographically, poles to both bedding and axial planes form a distinct west-northwest to east-southeast girdle, with a concentration of points representing steep dips; fold axes and planar-intersection axes plunge shallowly north-northeast and south-southwest (Fig. 13). The sense of overturning in asymmetric folds is towards the east-southeast. This structural fabric and geometry also characterizes the Woodruff and Webb Formations in Woodruff Canyon where they are depositionally overlain with strong angular unconformity by the Tonka Formation of the Newark Valley sequence, which is only slightly arched over a northwest-trending (Mesozoic?) axis (Fig. 13). The poorly exposed thrust (loc. 7, Fig. 12), which was originally mapped by Smith and Ketner (1978) and appears to be the basal imbricate carrying the Webb and Woodruff over the autochthonous Dale Canyon of the Rain mine area (loc. 8, Fig. 12), apparently is also depositionally truncated by the Tonka, although the contact relationships are obscured by high-angle faults. This thrust is interpreted as an offset segment of the basal imbricate of the Roberts Mountain thrust system exposed in Ferdelford Canyon (near locs. 9 and 10, Fig. 12). In summary, the structurally lowest imbricates of the Roberts Mountains thrust system

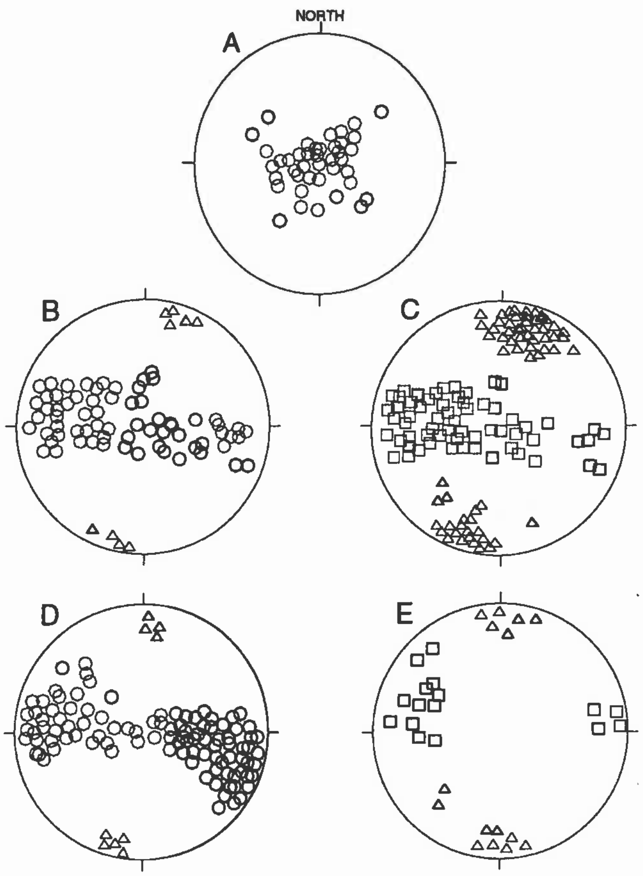


*Figure 13. Lower-hemisphere stereographic projections of structural-fabric data from the northern Piñon Range. (**A**) Poles to bedding (circles) of Newark Valley sequence ("Diamond Peak formation" of Smith and Ketner, 1978) near Woodruff Canyon. Bedding attitudes are taken from Smith and Ketner (1978) within one mile of the partly depositional and partly faulted contact between the gently arched Newark Valley sequence and the underlying more highly deformed Webb and Woodruff Formations. (**B and C**) Webb and Woodruff Formations, Woodruff Canyon (near loc. 5, Fig. 12); net data taken from Jansma (1988). B, Poles to bedding (circles) and fold axes (triangles) of upright to ESE-verging overturned folds. C, Poles to axial planes (squares) and fold axes (triangles) of ESE-verging folds. (**D and E**) Webb and Woodruff Formations, Ferdelford Canyon area (including localities 9 and 10, Fig. 12); net data taken from Jansma (1988). D, Poles to bedding (circles) and fold axes (triangles) of upright to ESE-verging overturned folds. E, Poles to axial planes (squares) and fold axes (triangles) of ESE-verging folds.*

These data show that the coaxial, polyphase folds of the Webb and Woodruff Formations associated with imbricates of the Roberts Mountains thrust system in Ferdelford Canyon have the same geometry of deformation that characterizes the Webb and Woodruff in the vicinity of Woodruff Canyon. Furthermore, they show that this deformation occurred prior to deposition of the unconformably overlying Newark Valley sequence, which locally is only little deformed.

and the structural fabrics associated with them are depositionally overlapped by mid-Late Mississippian strata, and the foredeep was thus actively deforming and subsiding during Osagean and Meramecian (mid-Mississippian) time.

Because the lowest set of imbricates of the Mississippian Roberts Mountains thrust system overrides autochthonous Dale Canyon Formation, that depositionally overlies miogeoclinal Devonian carbonate rocks, the existence of a thick slab-like RMA in the subsurface across the Piñon range, as interpreted by Carpenter et al. (1993; 1994), is precluded. Inspection (at the Nevada Bureau of Mines and Geology) of cuttings from the Petan Trust well (loc. 4, fig. 12) indicates that this basal imbricate is intersected at a depth of about 870 feet in this well. Above this depth, starting at about 450 feet, cuttings of abundantly radiolarian siliceous argillite are indicative of the Webb or Woodruff. From well depths of about 870 to 2700 feet cuttings are mostly of black argillite, becoming progressively silty with depth. From about 2700 to the base of terrigenous-clastic rocks at 4500 feet, quartz-chert sandstone forms 20 to 80% of the cuttings, mixed with black argillite. The section penetrated below about 870 feet is thus interpreted to be the autochthonous Dale Canyon Formation. It has a structural thickness comparable to that of the Dale Canyon section from its partially detached contact with Devonian carbonate rocks in the Rain mine (loc. 8, Fig. 12) up to the level of the basal thrust imbricate (loc. 7, Fig. 12). Unspecified fossils of uppermost Devonian or lowermost Mississippian age are reported by Carpenter et al. (1993) from the stratigraphically highest carbonate rocks beneath the Mississippian clastic section in the Petan Trust well. These carbonate rocks are possibly those of the Island Mountain Formation, which characteristically contain reworked Famennian and Kinderhookian microfossils. Some amount of detachment between rocks of the Diamond Range sequence and Devonian carbonate rocks on a bedding subparallel attenuation fault is expected in the Petan Trust well, because such a fault forms this contact in outcrops farther east, as at locality 6 (Fig. 12).

Different imbricates of the Mississippian Roberts Mountains thrust system carry different facies of the Dale Canyon clastic rocks. The autochthonous (with respect to Roberts Mountain thrusting) Dale Canyon Formation (Mdc_1 on Fig. 12) is conspicuously sandy and conglomeratic in its lower part. In comparison, the allochthonous Dale Canyon (Mdc_2, Fig. 12) of Ferdelford Canyon, which rides on the structurally lowest Roberts Mountains thrust imbricates, is much more fine grained. In this facies of the Dale Canyon finely fractured argillite predominates, packets of sandstone are scarce, and conglomerate is confined to two lenticular channel-fill(?) units low in the section in Ferdelford Canyon. Still different facies of the Dale Canyon are represented in structurally higher Roberts Mountains thrust imbricates that carry the Ordovician Vinini Formation. Laterally persistent gravity-flow units of coarse conglomerate form an important part of the lower few hundred meters of the Dale Canyon (Mdc_3, Fig. 16 in the Road Log) that depositionally rests directly on the Vinini south of Papoose Canyon (PA, Fig. 1) in the Piñon Range. The Dale Canyon Formation at the west end of Carlin Canyon (loc. 2, Fig. 12; CC, Fig. 1), i.e., the part of the original Tonka Formation beneath the unconformity within the Tonka as originally described, may be part of this or still another Roberts Mountains thrust imbricate.

Thrust slices of the RMA that contain both rocks of the Ordovician Vinini Formation and of the Late Devonian Woodruff Formation also contain slivers of deep-water Silurian and Devonian rocks representative of the outer miogeoclinal ramp. Such slivers occur in the Piñon Range (just south of PA, Fig. 1) where they constitute the "Roberts Mountains Formation" of Smith and Ketner (1978) and in the Roberts Mountains (at GC, Fig. 1) where they are the allochthonous "Devonian Shale and Limestone" of Murphy et al. (1984). These examples of outer-ramp miogeoclinal rocks within the Mississippian RMA, although relatively small, are significant in being juxtaposed contrasting facies of their autochthonous, lower plate equivalents. For such relationships, an explanation involving multiple generations of Roberts Mountains thrusting was proposed by Murphy et al. (1984) for the Roberts Mountains and also by Carlisle and Nelson (1990) with respect to rocks of the RMA in the Sulphur Springs Range (near DC, Fig. 1). A somewhat different interpretation of relationships in the Roberts Mountains by Jansma and Speed (1995) restricts the RMA to the allochthonous Ordovician rocks and views not only the Devonian strata but also the Lower Mississippian strata as fault slices underthrust, duplexed, and underplated beneath the RMA. They imply that all of these particular slices were assembled during Early Mississippian underthrusting, but they also accept the possibility that the RMA in general may have been emplaced during two or more episodes of thrusting.

The concept of successive episodes of Roberts Mountains thrusting during the Late Devonian and mid-Mississippian makes possible the hypothesis that the Mississippian RMA is multigenerational and formed of Mississippian foredeep and wedge top deposits, that are synorogenic with Mississippian Roberts Mountains thrusting, as well as older rocks that had already been juxtaposed during older Roberts Mountains thrusting. As a possible significant example of this, the "Roberts Mountains thrust" that emplaced basinal Ordovician rocks of the Vinini Formation over Late Devonian and older, outer-ramp, miogeoclinal rocks in the gold mines of the main Carlin trend immediately northwest of Carlin may be a Late Devonian thrust. Rocks forming both the upper and lower plates of this "Roberts Mountains thrust" then may be part of the Mississippian RMA. The lower-plate Devonian and Silurian miogeoclinal strata of the Carlin trend, in the lower plate of this possibly Late Devonian thrust, are much more outer ramp in character (e.g., as described by Ekburg et al. 1991) than are their platform equivalents less than 20 km distant in the Piñon Range. Transport of these outer-ramp miogeoclinal rocks as part of the Mississippian RMA may be responsible for this apparent juxtaposition of facies. Upper-plate rocks of the Vinini Formation in the Carlin trend may extend southeastward to include exposures of the Vinini near the west entrance to Carlin Canyon (Fig. 12). Here they are apparently depositionally overlain by Mississippian strata of the Webb and Dale Canyon Formations, and as described above, this facies of the Dale Canyon is quite different from that ("Mdc_2," Fig. 12) which characterizes structurally lower imbricates of the Mississippian Roberts Mountains thrust system a short distance to the south in the Piñon Range. The possible Late Devonian Roberts Mountains thrust, its upper and lower plates, and Mississippian strata deposited on its upper plate thus all may have been thrust farther eastward during the mid-Mississippian episode of Roberts Mountains thrusting. Only the frontal imbricates of this Mississippian Roberts Mountains thrust system are represented in the Piñon and Sulphur Spring Ranges and the Roberts Mountains.

The hypothesis of a multigenerational Mississippian RMA requires that, farther west than the Piñon Range and Roberts Mountains, the sole of the Mississippian Roberts Mountains thrust cut upward through its miogeoclinal footwall from a deeper crustal level along which the thrust propagated. This hypothesis for the Mississippian RMA is significant for modeling the paleotectonics of the Mississippian Antler foreland, because if true, the tectonic load during this phase of Antler orogenesis would have included not only the foredeep fill and the classic upper plate rocks, such as the Vinini Formation, but also substantial parts of the miogeocline itself.

PALEOTECTONIC INTERPRETATION

In a broad sense, all of the Mississippian strata in eastern Nevada and western Utah belong to the Antler foreland and are genetically related to Antler orogenesis. However, the paleotectonic controls on sedimentation, as reflected by the stratigraphic record, changed dramatically and episodically through Mississippian time.

The Mississippian foreland shows a pronounced continentward overstepping of the foregoing Devonian Antler foreland, as represented by the Pilot Shale of late Frasnian (mid Late Devonian) to earliest Kinderhookian age. Although commonly characterized as being mostly a deep basinal, submarine-fan deposit, the Pilot interfingers with lagoonal carbonates such as the upper Devils Gate Limestone (Nichols and Silberling, 1995) and West Range Limestone (Silberling et al., 1995), and some of its fine-grained clastic parts resemble the distal-foreland siltstone lithosome of the Mississippian Sadlick sequence, suggesting relatively shallow, restricted, possibly hyposaline sedimentation. Several third-order sequences can be recognized within the Pilot (Macke et al., 1994; Giles and Dickinson, 1995); but altogether they indicate deposition in the distal, restricted, possibly backbulge (Goebel, 1991), part of the Late Devonian to earliest Mississippian Antler foreland. A backbulge setting for at least some parts of the Pilot Shale accords with the possibility of one or more episodes of Late Devonian Roberts Mountains thrusting, as discussed above.

The Pilot Shale is depositionally overlain by the late Kinderhookian Morris sequence in belts A, B, and C (Fig. 1). The more open-marine carbonate rocks of the Morris sequence represent a marked change in the paleotectonic regime, and a pronounced eastward shift of equivalent foreland facies tracts (Giles and Dickinson, 1995).

During deposition of the Morris sequence the continental margin was apparently passive and Antler orogenesis inactive. Shallow-marine carbonate rocks of the Morris sequence (Fig. 3B) blanket most of the foreland region and exhibit only relatively minor lateral change. They extend as far west as the southern Diamond Mountains (at TC, Fig. 1) and northern Pancake Range (at BS, fig. 1), but elsewhere in belt A they are erosionally truncated beneath the base of the Diamond Range sequence (Fig. 11). To the north of Utah, rocks of Morris-sequence age merge into the Mississippian carbonate succession of Idaho and Wyoming (e.g., Poole and Sandberg, 1991). Farther west, in Nevada, conodont-based age determinations indicate that deep-marine clastic rocks deposited in proximal parts of the foredeep and occurring in part within imbricates of the RMA are correlative with the Morris sequence, assuming that the conodonts in question are not reworked into younger strata. At present, the paleogeographic connection between these foredeep rocks and the shelfal rocks of the Morris sequence is obscure.

The Sadlick sequence and age-equivalent Diamond Range sequence document return to a paleotectonically active foreland during Osagean and much of Meramecian time. In the proximal foreland, pronounced subsidence, deposition of gravity-flow terrigenous-clastics derived from the Antler orogenic belt, and synorogenic involvement of these deposits in active thrusting and possibly flexural-extension normal faulting are all recorded in the Diamond Range sequence. To the east in the distal foreland, the essentially correlative Sadlick sequence represents subsidence of modest magnitude followed by development of topographic restriction, as indicated by the strata representing the Delle phosphatic event. This, and the lack of stratigraphic connection between these two correlative sequences, suggest that an emergent area of low relief separated them. The mid-Mississippian stratigraphic record is thus consistent with the foredeep, peripheral bulge, and backbulge paleogeography of a flexural-loading system (Fig. 2).

Some aspects of the mid-Mississippian stratigraphy that seemingly conflict with a flexural-loading paleotectonic regime during Sadlick sequence time can be explained by the inferred active encroachment of the RMA during this time and the consequent continentward migration of flexural-loading paleogeographic elements. At the latitude of Eureka, Nevada, exposures of the deep-subtidal Harris Canyon member of the Joana Limestone of the Sadlick sequence in the White Pine Range (as at locs. AS and HC, Fig. 1) are geographically close to the permissibly correlative debris-flow limestones of the Island Mountain Formation (the "Tripon Pass" limestones of some authors) at the base of the Diamond Range sequence in the southern Diamond Mountains (locs. TC and PB, Fig. 1; Figs. 3A and 3B). Direct down-slope transport of material from the Harris Canyon into the debris-flow limestones of the Island Mountain might seem reasonable, as first postulated by Poole (1974). This could even be the case despite some telescoping of their original lateral separation on Mesozoic thrust faults of the Central Nevada thrust belt and despite foredeep-parallel, southward-directed, gravity-flow sediment transport (Nichols and Silberling, 1995) during Island Mountain deposition. However, the eastward pinch-out of the Island Mountain Formation, and the absence of any sections where the Island Mountain is seen to have been deposited over the Harris Canyon, argues for the existence of an emergent belt that separated the Island Mountain and the Harris Canyon. Originally the crest of this emergent belt or forebulge may have been considerably farther west than the present extent of the Harris Canyon. The absence of a shoreward western facies of the Harris Canyon that would have been originally deposited on the distal side of this forebulge can be explained by postulating that this shore facies necessarily would have been bowed up and eroded away as the forebulge progressively migrated east during Sadlick-sequence time.

Another possible paleotectonic relationship between the Island Mountain Formation and the Harris Canyon member of the Joana is suggested by the general stratigraphic model for peripheral foredeeps recently proposed by Sinclair (1997). In this model, three lithologically distinct units make up an "underfilled [foreland basin] trinity." The units of the trinity are viewed as lateral facies that form a distinctive retrogradational stratigraphic succession as the foredeep and peripheral bulge migrate progressively cratonward. They are formed of (1) ramp carbonate rocks deposited on the cratonic margin of the basin (the lower unit of the trinity), (2) hemipelagic mud rocks deposited offshore from the cratonic margin (the middle unit), and (3) deep-water turbiditic siliciclastic rocks deposited towards the orogenic margin (the upper unit). If the Harris Canyon represents the lower unit of this trinity, it would have been deposited on the distal ramp of the foredeep, and its eastern extent would have been limited by the forebulge which had migrated eastward to a position about where it is located by Giles and Dickinson (1995) for this period of time. The fine-grained rocks of the Island Mountain would correspond to the middle unit of the trinity. However, because the Island Mountain is nowhere seen to overlie the Harris Canyon, the direction of forebulge migration would have to have reversed itself dramatically during Island Mountain and Harris Canyon time. The forebulge would have to have jumped westward to a position between the distal Island Mountain and the Harris Canyon to explain the lack of stratigraphic continuity between these two rock units. Either of these model-driven suppositions is possible as an explanation of the apparent separation of the Island Mountain and Harris Canyon. Further study is needed for a definitive solution to their relationship.

Following deposition of the Diamond Range sequence and laterally equivalent Sadlick sequence, other paleotectonic controls influenced deposition of the more or less correlative Newark Valley and Maughan sequences during the latest Meramecian and most of the Chesterian. A flexural-loading model does not seem applicable at the latitude of northern Nevada and Utah during this time span. Not only is structural evidence lacking for renewed thrusting of the RMA during this time, but depositional patterns are quite different from those of the mid-Mississippian. The preexisting foredeep became the site of partly coarse-clastic fluviatile and deltaic deposition (the Antler "Molasse" of Poole and Sandberg, 1991) sourced generally from the west in the Antler orogenic belt. Uplifted foredeep fill of the Dale Canyon Formation was recycled into the Newark Valley sequence, as indicated by the more resistant composition of clasts in Newark Valley as compared with the Diamond Range sequence (Perry, 1995; Trexler, unpublished data). Farther east, Newark Valley/Maughan sequence deposits grade into predominantly paralic pelitic rocks and then, as in the eastern part of belt C (Fig. 1) in the Confusion Range (CR, Fig. 1), into restricted-marine calcareous and spiculitic fine-grained siliciclastic rocks. Still farther east, these rocks are laterally supplanted by open-marine, generally shallow-water carbonate rocks that accumulated to substantial thicknesses and prograded westward.

This part of Late Mississippian history in the more proximal foreland has been attributed to isostatic recovery following emplacement of the RMA (Johnson and Pendergast, 1981). In the more distal foreland (belts D and E, Fig. 1), however, the exaggerated thickness of the Maughan sequence requires some further explanation. One possibility is the initiation of late Paleozoic tectonism associated with the Ancestral Rocky Mountains and convergence at the southern margin of the continent, leading to an increase in subsidence in north-central Utah that culminated with development of the Oquirrh basin (Kluth and Coney, 1981; Geslin, 1993).

ROAD LOG

This field trip starts from and returns to Salt Lake City, Utah, and extends over a period of three days. The route and location of stops are shown on Figure 14. Because of the constraints imposed by time, access, and logistics, sampling the Mississippian record of the Antler foreland necessarily will be somewhat disjunct, inevitably jumping both stratigraphically and paleogeographically from one place to another. Figure 15 shows what features of the foreland can be observed at different field-trip stops on two diagrammatic cross sections, one roughly paralleling U.S. I-80 and the other U.S. Hwy. 50.

Day 1.

From Salt Lake City, Utah, head west on U.S. I-80.

At about Exit 84, view to north is along the axis of the Stansbury Island anticline. The conspicuous light-colored unit on east and west limbs, on either flank of the island, is shallow-marine sandstone. This unit occupies the same position within the Upper Devonian section as the Stansbury Formation but is quite different from this formation at its type area in the Stansbury Mountains immediately to the southwest. Quartzites of the type Stansbury were influenced during deposition by differently directed currents and are intercalated with large thicknesses of dolostone-clast conglomerate derived from subjacent Paleozoic strata (Trexler, 1992). The mechanism of basement uplift here at that time is enigmatic.

At about Utah Mile Post (MP) 80, view to south is along the east-dipping Mississippian and Devonian section along the east flank of the Stansbury Mountains. The unit being quarried out along strike is the Upper Mississippian Great Blue Limestone, which is refined into lime in the processing plant at Flux.

Roadcuts along I-80 near MP 77 are Cambrian rocks describing minor anticlines and synclines in the axial region of a major antiform. This has led to the speculation that the Stansbury Island anticline is displaced right laterally from the antiform at the north end of the Stansbury Mountains along a fault paralleling I-80, despite the different expressions of the Stansbury Formation in the two ranges (Cashman, 1992).

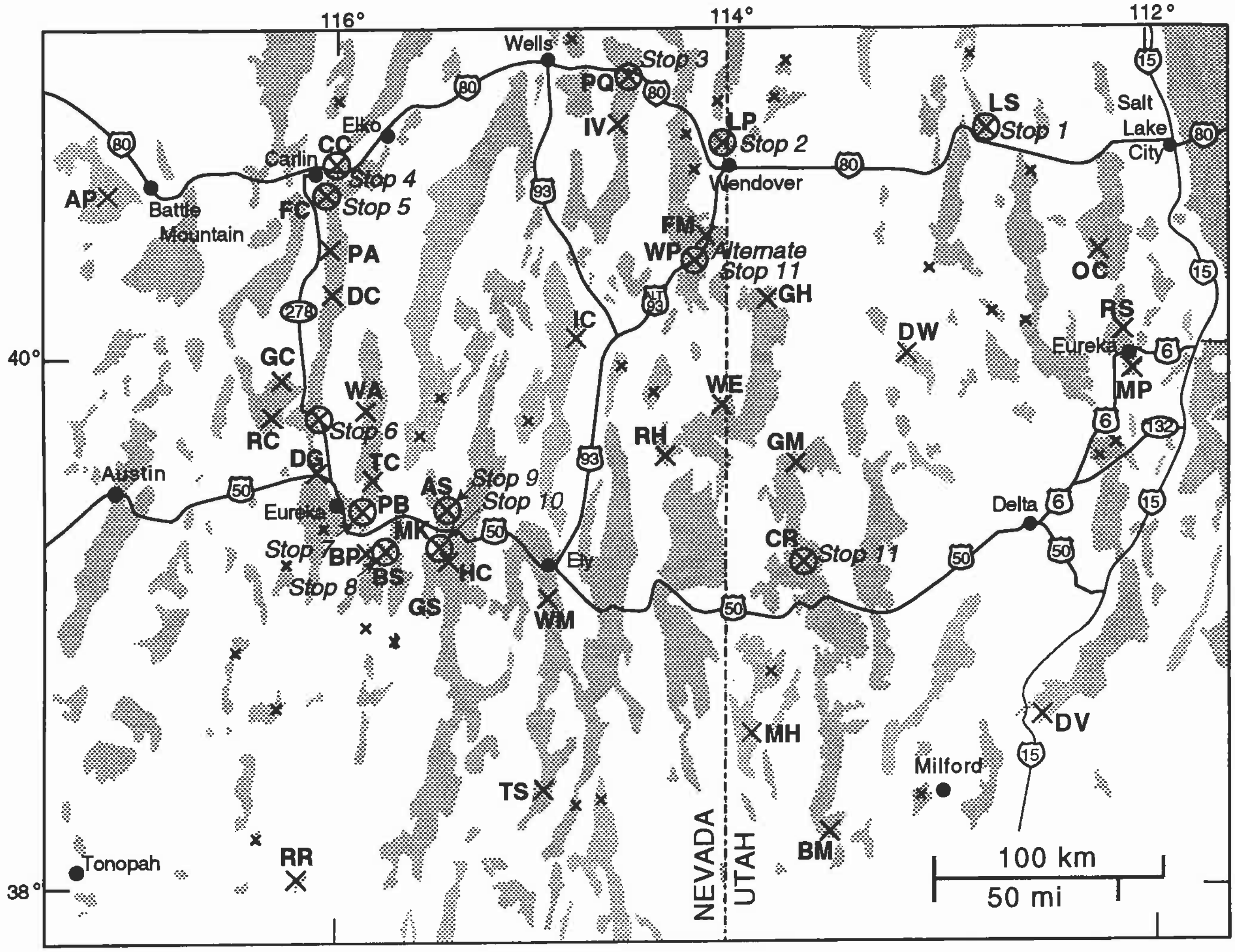


Figure 14. Index map of parts of eastern Nevada and western Utah showing the location of the field-trip stops. Symbols of locality names as for Figure 1. Shaded areas are exposures of pre-Tertiary strata and(or) plutonic rocks from Stewart and Carlson (1978) and Hintze (1980).

Exit 70 (Delle Exit). Turn off.

0.00 miles at stop sign. Turn right (north), then immediately turn left (west) onto indifferently paved frontage road.

At 3.1 miles, cross railroad tracks and then take the right fork onto the gravel road heading north. This is the access road to the radio towers atop the south end of the Lakeside Mountains.

Beginning at about 4.0 miles, is a spectacular view ahead of the east-dipping Mississippian-Devonian section forming the south end of the Lakeside Mountains below the radio towers. Follow this road around the flat and then northward uphill to the crest of the spur, where the road turns west and follows the ridge top uphill towards the radio towers.

8.5 miles. Park on the bulldozed flat.

Stop 1. [SE1/4, SW1/4, sec. 6, T.1 N., R.8 W.; Delle, Utah, 7.5′ quadrangle] This is the type section of the Delle Phosphatic Member (originally defined as the basal member of the Woodman Formation by Sandberg and Gutschick, 1994), which forms the upper part of the Sadlick sequence. As explained in the text, Woodman and Needle Siltstone of the Chainman Shale are synonymous as originally defined.

The objectives at this stop are to examine (1) the lithologically peculiar strata representing the Delle phosphatic event and (2) the immediately underlying carbonate rocks of the Gardison Limestone characteristic of the lower Sadlick sequence (see Fig. 7). The full thickness (of about 100 m) of the Gardison and the subjacent upper Kinderhookian Fitchville Formation, of the Morris sequence (about 35 m

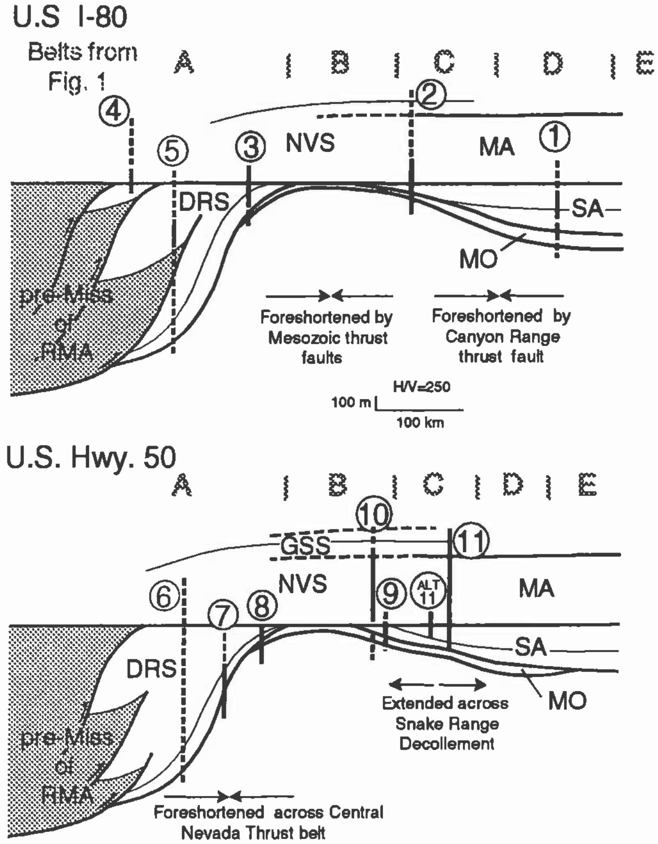


Figure 15. Conceptual cross sections of Mississippian Antler foreland rocks generally following the routes of U.S. I-80 and U.S. Hwy. 50 (see fig. 14) and showing the stratigraphic span of the rocks that can be traversed (solid vertical lines) and observed in the distance (dashed vertical lines) at each of the field-trip stops (encircled numbers). GSS, Green Springs sequence; NVS, Newark Valley sequence; MA, Maughan sequence; DRS, Diamond Range sequence; SA, Sadlick sequence; MO, Morris sequence; RMA, Roberts Mountains allochthon. Cross sections are palinspastically corrected for the estimated horizontal deformations indicated by the opposing and diverging arrows.

thick), can be seen in the distance to the east beneath the Delle, which is offset from the type Delle by a major, east-side up normal fault in the canyon which the radio-tower road follows.

Salient features that can be observed at this locality are:

—The abrupt change from carbonate sedimentation of the Gardison Formation to the phosphatic, cherty, and partly clastic strata characteristic of the Delle phosphatic event.

—Evidence for dissolution of limestone, especially at the Gardison-Delle contact and higher in the section where pinch-and-swell structures resulted from dissolution of original limestone beds.

—Thin, stratigraphically isolated beds of black chert, evidently replacing sparsely radiolarian limestones.

—Peloidal and oolitic phosphorite layers, mainly as reworked, detrital aggregates, but in part as diagenetic replacements.

—Ubiquitous secondary dolomite as cement or matrix of silty to fine-grained sandy clastic rocks; second to quartz, dolomite is volumetrically the most important constituent of these rocks.

—Subordinate amount of black shale in Delle section.

—Suggestion that several red layers in lower part of Delle section may have resulted from oxidation of original iron sulfides concentrated in the basal parts of parasequences.

—Appearance of dense, internally featureless, yet distinctive "ostracode lime mudstone" units, particularly that at the top of the type Delle section. These are interpreted as having been deposited from hyposaline water at modest depths (Jewell et al., 1996).

Return to I-80 at Exit 70. Continue west on I-80.

From about MP 64 to 40, pre-Tertiary rocks in the low hills at the north end of the Cedar Mountains and south end of the Grassy Mountains, through which the highway traverses, are all of the Permian-Pennsylvanian Oquirrh Formation, the great thickness of which is estimated in kilometers. Bedding is mostly steep to overturned. About 20 miles north of I-80, looking north from about MP 40, is the south end of the Newfoundland Mountains in which the Oquirrh is not represented. Instead Permian strata resembling those of northeastern Nevada rest directly on Devonian rocks. The contrasting upper Paleozoic section in the Newfoundland Mountains thus appears to be structurally juxtaposed with that in the Grassy Mountains. On Figure 1 the major structure responsible for this is suggested to be the northern continuation of the Canyon Range thrust system, the sub-Cenozoic break-out of which would cross beneath I-80 about here on the east side of the Great Salt Lake Desert.

[The time zone changes from Mountain to Pacific at the Utah-Nevada state line]

Exit 410 (West Wendover exit). Turn off.

0.00 miles at stop sign. Turn right (north) onto paved frontage road.

0.2 miles. Turn right (north) onto unpaved track leading up A-1 Canyon into the Leppy Hills.

1.6 miles. Fork in road; stay right. From about 3.00 miles up canyon to the saddle, exposures along the road are of the Famennian (upper Upper Devonian Pilot Shale, "shale" being a misnomer for dolomitic siltstones, some of which lithologically resemble those of the Mississippian Woodman and Needle Siltstone units, as at Stop 1).

3.55 miles. Park here at saddle.

Stop 2. [1/2 mile NE of Leppy Peak; Leppy Peak, Nev.-Utah 7.5′ quadrangle; see Schneyer (1990)] The object at this stop is to observe the depositional contact between dolomitic siltstone of the Sadlick sequence and the underlying upper Kinderhookian limestone that represents the Morris sequence. The siltstone, assigned by Nichols and Silberling (1993) to the "Needle Siltstone," is essentially the same unit as the Delle Phosphatic Member in the Lakeside Mountains, and the subjacent limestone is the lower Joana Limestone, correlative with the Fitchville Formation farther east. Strata equivalent to the upper part of the Joana, such as the Gardison Limestone at Stop 1, are missing in this section at the Needle-Joana disconformity. Thus, the upper part of the Sadlick sequence (the Needle) appears to have overstepped its lower part in a westward direction, suggesting a topographic barrier to the west.

Onset of the Delle phosphatic event at the base of the Needle Siltstone is marked by a thin rind of detrital phosphate on the surface of the solution-compacted upper part of the lower Joana Limestone. This is best observed on the hill immediately west of the saddle, just uphill from the power-line poles. The internal stratigraphy of the Joana Limestone and its contact with the underlying Pilot Shale, are best exposed on the next spur to the south about 1/4 kilometer away.

The Needle Siltstone is only about 20 m thick; it is apparently overlain by quartz-chert sandstone uphill to the west, but the nature of the contact is uncertain.

Return to I-80 at Exit 410. Continue west on I-80.

Nevada MP390. Silver Zone Pass over the Toano Range. Granitic rocks along highway are part of the Silver Zone pluton of Late Jurassic age.

The general northerly trend of pre-Tertiary rock units and structural boundaries is deflected to more easterly trends in northeasternmost Nevada (Fig. 1). This anomaly, informally termed the "Wendover deflection," has been ascribed variously to a change in original depositional trends, southeast-vergent thrust faulting, pre-Tertiary strike-slip faulting, and/or accommodation faulting between different domains of Tertiary extension.

Exit 376 (Pequop Exit). Turn off.

0.00 miles at stop sign. Turn left (southwest). At 0.25 miles turn left (south) again, and then at 0.4 miles turn right (northwest) onto old highway. Pelitic rocks of Dale Canyon Formation of the Diamond Range sequence crop out on either side of valley. Thick chert-pebble conglomerate bed in road cut at 1.15 miles; broken and sheared quartz-chert sandstone in black, finely fractured argillite at 1.65 miles.

2.15 miles. Contact of Dale Canyon with Tripon Pass Limestone. Park near here.

Stop 3. [Central E1/2, SW1/4, sec. 19, T.37 N., R.66 E.; Pequop Summit, Nev. 7.5′ quadrangle] This is the easternmost exposure of Mississippian Antler foredeep rocks at this latitude. The Tripon Pass Limestone here is mostly radiolarian, spiculitic, hemipelagic/pelagic lime mudstone interpreted to have been deposited during the initial subsidence of the foredeep, prior to the time when the Antler orogen became the source of the more voluminous clastics of the overlying Dale Canyon Formation. It occupies the same position in the foreland as does the Island Mountain Formation south of Carlin. Near Stop 3, it is faulted against the underlying Guilmette Formation. However, about 3.5 miles south along strike it unconformably overlies 15–20 m of crinoidal limestone of the lower Joana Formation disconformably overlying the Guilmette. The structural thickness of the deformed Tripon Pass is more than 100 m where both its base and top are exposed, according to the geologic map by Camilleri (1994).

Return to I-80 at Exit 376. Continue west on I-80.

The type locality of the Tripon Pass Limestone is about 15 miles northeast of Wells, where it is at least as deformed by Mesozoic structures as it is at Pequop Summit and is more varied lithologically.

From Wells to Elko the mid-Tertiary Ruby-East Humboldt metamorphic core complex occupies the high ranges to the southeast of the highway. Tertiary extension in this region is of palinspastic importance for reconstruction of the Antler foreland. Mississippian exposures in the northern Piñon Range near Carlin may originally have been part of the sedimentary cover above the crystalline rocks now in the footwall of the Ruby core complex (Howard, 1992).

Stop 4. [E1/2, W1/2, sec. 21, T.33 N., R.53 E.; Carlin East, Nevada 7.5′ quadrangle] This is at west end of Carlin Canyon on the old highway that follows the big bend of the Humboldt River to the north of the tunnels on I-80 (Fig. 12). Access depends on the state of road work in the tunnels. Either turn right off of westbound I-80 onto old highway just before crossing the Humboldt River and entering the tunnels, or, if entrance to the old highway is blocked, continue west towards Carlin, turn off at Exit 282 (East Carlin Exit), cross over I-80 on the overpass to the frontage road on the south side, turn left (east) and follow the frontage road, which becomes the old highway and passes back under I-80 to follow the big bend of the river.

Carlin Canyon is well known for the photogenic angular unconformity (loc. 1, Fig. 12) seen in many geology textbooks. At this unconformity Lower Pennsylvanian Strathearn Formation limestone overlies more steeply dipping, Upper Mississippian Tonka Formation conglomerate and grainstone, a relationship that was taken as evidence for an early Pennsylvanian deformation termed the "Humboldt orogeny" (Ketner, 1977). This deformation is regarded as an evolutionary phase of the late Antler orogeny by Trexler et al (1991). Jansma and Speed (1990) suggested that this contact is a Mesozoic fault, but careful inspection shows evidence of a paleosol on the Tonka surface and incorporation of Tonka clasts into the basal Strathearn strata. Only minor slip occurred on the unconformity surface during Mesozoic folding.

A second angular unconformity occurs in Carlin Canyon within conglomerates of the Tonka Formation (loc. 2, fig. 12). This unconformity can easily be seen looking north across the river from the hills south of U.S. I-80, and it can be mapped along the ridge west of the canyon. Conglomerates beneath the unconformity (now included in the Dale Canyon Formation) dip more steeply than those above it, and a significant proportion of their clasts are of non-resistant rock types. The Dale Canyon apparently lies conformably on the Webb Formation. Though covered, its contact with the Webb can be traced along the western slope of the ridge west of the canyon. The Webb, in turn, is apparently depositional on the Ordovician Vinini Formation in nearby exposures (Fig. 12).

The multiple angular unconformities here represent progressive deformation of these strata in mid-Late Mississippian and latest Mississippian to early Pennsylvanian time. The pattern is strikingly similar to well documented examples of progressively rotated strata in the footwall of thrust-belt structures in the Ebro Basin of the eastern Pyrenees Mountains (Anadón et al, 1986).

Return to I-80 via Exit 282. Back track (east) on I-80 to Elko, Nevada. End of Day 1.

Day 2.

Head west on I-80 to Exit 279 (Central Carlin and Eureka exit). Turn off I-80.

0.00 miles at stop sign. Turn left (south).

0.2 miles. At stop sign turn right (west) onto Chestnut Street (=Nevada Hwy. 221).

1.7 miles. Intersection with Nevada Hwy. 278. Turn left (south).

12.0 miles. Turn left (east) off of Hwy. 278 opposite the "Brown Ranch" onto the gravel road into Ferdelford Canyon and the Piñon Range.

At about 14.0 miles. View ahead on the north side of Ferdelford Canyon of thick, massive conglomerate of the Newark Valley sequence unconformably overlying black argillite of the Dale Canyon Formation (Diamond Range sequence) which forms the smooth, dark slopes below the conglomerate. Above the conglomerate, and down-faulted to the west against it, are peritidal cycles of shallow marine clastic and impure carbonate rocks containing the rich shelly fauna of the "Ferdelford fossil beds." The late Osagean-early Meramecian age interpretation reported by Smith and Ketner (1975) for the Ferdelford beds dates back to the 1960's, and it conflicts with the early Meramecian palynological age determination by the Mobil Oil Company (Carpenter et al., 1993, 1994) for strata that are low in the Dale Canyon Formation and far below the level of the Ferdelford fossil beds.

15.6 miles. Improved (?) gravel road turns south into tributary canyon. Continue straight up main canyon on dirt track.

15.75 miles. Tributary on north side of main canyon follows a north-northwest-trending high-angle normal fault that drops the Newark Valley-Diamond Range sequence boundary down to the east.

16.35 miles. Park here (or take the side road headed south if your vehicle can negotiate the steep arroyo crossing). Walk (or drive) up the track that starts up the main (or south) fork of Ferdelford Canyon.

Stop 5. [Central NE1/4 sec. 13, T.31 N., R.52 E.; Ravens Nest, Nevada 7.5′ quadrangle] From the end of this track (at 16.65 miles) hike up the small spur immediately to the northeast. Outcrops of the Webb Formation atop this spur (loc. 9, Fig. 12) show the mesoscopic, polyphase folding, whose fabric and geometry (Fig. 13) has been associated with the Mississippian Roberts Mountains thrust (RMT) by Jansma (1988). The Webb here is riding on an imbricate of the RMT which follows this fork of Ferdelford Canyon (and which was reoccupied by the younger normal fault of opposite displacement which Ferdelford Canyon follows farther downstream). The type Webb Formation (at loc. 10, Fig. 12) is in the lower of the two other imbricates of the RMA on the opposite side of the canyon.

The contact between the Webb and overlying argillites of the Dale Canyon Formation appears to be depositionally gradational, but was interpreted here as a low-angle younger-on-older Antler-age fault by Jansma and Speed (1993).

From this vantage point, the crest of the ridge of Devonian and older autochthonous (with respect to the RMT) miogeoclinal rocks northeast of locality 11 (Fig. 12) can be seen on the skyline several kilometers to the south-southeast. To the northwest, the gently dipping, little-deformed strata of the Newark Valley sequence rim the Ferdelford basin, unconformably overlying the more deformed rocks of the Diamond Range sequence.

Backtrack down Ferdelford Canyon to Nevada Hwy. 278.

Reset mileage to 0.00. Turn left (south) onto Hwy. 278.

1.5 and 1.65 miles. Tanks and pump(s) of Tomera Ranch oil field on east side of highway. Production has been from Tertiary volcanic and sedimentary rocks, but oil shows are reported from deeper drilling into the "Chainman" and "Devils Gate Formations" (Hansen and Ransome, 1990).

At about 9.0 miles, the prominent smooth, rounded hill (in the NE1/4, sec. 13, T.29 N., R.52 E.; Papoose Canyon, Nevada 7.5′ quadrangle) about 3 miles east of the highway in the near distance is capped by the Ordovician Vinini Formation in the upper plate of the RMA. Exposures in the vicinity of this hill document some of the critical relationships between the RMA, Dale Canyon Formation, and overlapping Newark Valley sequence. Time constraints prevent visiting this area, but some of these relationships are illustrated on Figure 16.

Within the RMA in this area, the Vinini is locally thrust over a sliver of Silurian outer-ramp miogeoclinal rocks (south of Fig. 16), which is in turn thrust over the Upper Devonian Woodruff Formation. The sole thrust of this imbricate stack carries the Woodruff over autochthonous (with respect to the RMT) Dale Canyon Formation. Depositionally overlying the Vinini Formation is an allochthonous, distinctly different facies of the Dale Canyon Formation. All of these units and the unconformably overlying Newark Valley sequence are folded over a major northwest-trending and plunging major anticline (the "Willow Creek anticline" of Carpenter et al., 1993), which is one of a series of kilometric-scale Permian or younger (probably Mesozoic) folds involving the Newark Valley sequence and younger Paleozoic rocks. On the nose and northwest flank of this anticline at locality PA (Figs. 1 and 16) the angular discordance between the gently dipping Newark Valley Sequence and the steeply overturned Diamond Range sequence rocks of the allochthonous Dale Canyon Formation is about 110°. Thus, prior to being folded along with the Newark Valley sequence in Mesozoic time, the Dale Canyon that is part of the Roberts Mountains allochthon south of Papoose Canyon was steeply dipping and striking a little east of north, like the orientation of bedding and axial planes associated with the Roberts Mountains thrust in Ferdelford Canyon and elsewhere.

12.0 miles. Tanks of the Willow Creek oil field to the east of the highway.

About 21–22 miles. From 1982–1989, the Blackburn oil field east of the highway produced nearly 2 million barrels of oil from several wells in the lower part of the Tertiary section and the underlying "Chainman Shale" and Devonian miogeoclinal carbonate rocks (Johannesen and Cole, 1990).

From about 30 miles, ahead and to the west, the typical Roberts Mountains thrust crops out in broken fashion along the east flank of the Roberts Mountains. The relatively treeless, smooth slopes low on the east flank of the Roberts Mountains are underlain by the type Vinini Formation. Roberts Creek Mountain, the high point in the range, is part of a large window of Devonian and older Paleozoic miogeoclinal rocks below the thrust. Also below the thrust are Webb-like chert and siliceous argillite and discontinuous remnants of the Dale Canyon formation.

51.0 miles. Garden Pass.

About 54.0 miles. Crossing the highway ahead, the conspicuously well-bedded east-dipping strata are of the Permian Garden Valley Formation which rests depositionally on the Vinini Formation of the Roberts Mountains allochthon.

56.0 miles. Turn off at road intersection on left (northeast) side of highway. Park here.

Stop 6. [At intersection of Hwy. 278 and Diamond Valley road in N part, sec. 22, T.22 N., R.52 E] View of the Diamond Mountains to the east across Diamond Valley. The Newark Valley-Diamond Range sequence boundary, as originally described by Trexler and Nichtman (1990), is within the Diamond Peak Formation about half way up the bare west slope of the Diamond Range. The Island Mountain Formation at the base of the Diamond Range sequence is just above the tree line; its contact with the underlying Devonian Pilot Shale, however, is complicated by strike-parallel faults.

Continue south on Hwy. 278.

At about 69 miles looking west, Anchor Peak is formed of rocks of the RMA thrust over Dale Canyon Formation on the Roberts Mountains thrust. To the east, the high point of the Diamond Mountains is Diamond Peak, held up by Pennsylvanian Ely Limestone. Much of the slope below is Newark Valley sequence strata overlying the Diamond Range sequence.

73.0 miles. Intersection with U.S. Hwy. 50. Turn left (east) and follow Hwy. 50 through Eureka and over Pinto Summit to intersection with Nevada Hwy. 892 (about 15 miles beyond Eureka).

The conspicuous open-pit mine about 15 miles southeast of here on the west flank of the White Pine Range is the Mount Hamilton mine, developed in a precious-metal, poly-metallic skarn system in Cambrian sedimentary rocks intruded by Cretaceous plutonic rocks.

Several generally north-trending Mesozoic faults of the Eureka or Central Nevada thrust belt have been mapped or inferred between Eureka and the White Pine Range. Individual thrust juxtapositions are significant for unraveling the Antler foreland stratigraphy, but the total shortening probably is not great.

Turn left (north) onto Hwy. 892 at its intersection with Hwy. 50.

0.00 miles at intersection.

At 2.4 miles, turn left (west) onto gravel road, pass by buildings and small leach pad, and continue west on dirt track into Packer Basin on the east flank of the southern Diamond Mountains. Enter mouth of prominent canyon at about 4.5 miles. This canyon follows a major Mesozoic(?) left-slip tear fault.

4.85 miles. Park here.

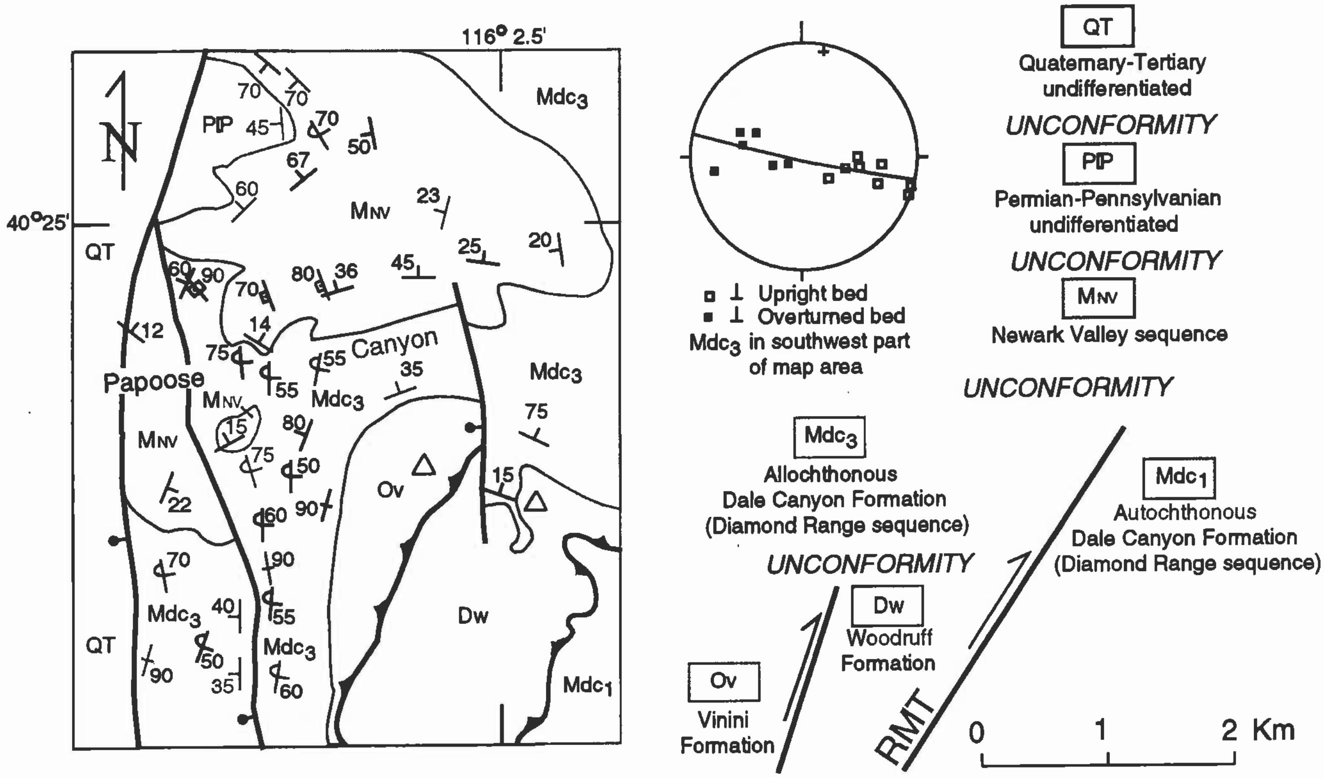


Figure 16. Geologic map of the area near Papoose Canyon (PA, Figs. 1 and 14) on the west flank of the Piñon Range. Lower-hemisphere stereographic projections of bedding attitudes in the allochthonous Dale Canyon Formation unconformably underlying the gently dipping Newark Valley in the southwest part of the map area show a geometry like that associated with Roberts Mountains thrusting farther north in the Piñon Range (Fig. 13).

Stop 7. [SW1/4, SE1/4, sec. 23, T.18 N., R.54 E.; Silverado Mountain, Nevada 7.5′ quadrangle; see Nolan et al. (1974)] The objective here is to examine the type section of the Island Mountain Formation (Fig. 11), which forms the low north-trending ridge north of the parking place. The lower Osagean Island Mountain is the basal part of the Diamond Range sequence beneath the Dale Canyon Formation. Its unconformable contact with the underlying Upper Devonian Pilot Shale is well exposed in the tributary canyon from the north, just up canyon from the parking place.

The upper Kinderhookian Joana Limestone of the Morris sequence is eroded away beneath the Island Mountain at this locality, but a 10 m thick remnant of the Joana separates the Island Mountain from the Pilot about 7 miles north of here at locality TC (Figs. 1 and 14).

Backtrack to intersection of Hwy. 892 and Hwy. 50. Turn left (east) onto Hwy. 50

At 0.5 miles east of intersection turn right (south) onto improved gravel road. This begins the route to Stop 8, and should not be attempted if the roads are excessively wet.

0.00 miles at intersection. Follow gravel road southward to fence corner at 2.65 miles and veer left onto the southeastward continuation of this road, leading into a broad north-draining valley in the Northern Pancake Range.

5.75 miles. Intersection; stay right (south).

6.6 miles. Black Shade Well and intersection with road to east. Continue south past well.

7.4 miles. Turn left (east) onto lesser traveled track.

7.8 miles. Mouth of small canyon cut through the Joana Limestone (of the Morris sequence). Park here.

Stop 8. [SE1/4, SE1/4, SW1/4 sec. 35, T.17 N., R.55 E.; Black Point, Nevada 7.5′ quadrangle; see Nolan et al. (1974)] The basal part of the Island Mountain Formation here consists of as much as 10 m of coarse limestone conglomerate, containing clasts of the lower member of the Joana Limestone (Fig. 11), of which only a thickness of about 16 m remains above its contact with the Pilot Shale. This conglomerate evidently fills a ravinement channel cut into an erosional surface on the Joana. The conglomerate is abruptly overlain by several meters of silicified, hemipelagic, spiculitic and radiolarian mudstone, which is in turn overlain by dark argillite of the Dale Canyon Formation. About 0.6 km farther north, about 22 m of Joana is overlain by several meters of radiolarian, spiculitic lime mudstone of the Island Mountain. These relationships are described by Nolan (1974), who was greatly puzzled by the peculiar lateral changes in thickness and character of his "Joana," which included the Island Mountain.

Backtrack to the intersection with Hwy. 50. Turn right and continue east on Hwy 50, crossing Pancake Summit and on to Little Antelope Summit on the crest of the northern White Pine Range. At 0.55 miles east of Little Antelope Summit turn left (north) off of highway onto gravel road.

0.00 miles at turnoff. Follow gravel road to north.

0.65 miles. At intersection, turn right (east) and follow this road past the well site of Tenneco Oil Co. Illipah Federal 1, a dry hole spudded in the Chainman Shale. Turn around at 2.85 miles, backtrack about 0.25 mile and park headed south along road.

Stop 9. [NE1/4, NE1/4, sec. 21, T.18 N., R.58 E.; Antelope Mountain, Nevada 7.5′ quadrangle] Hike up to the crest of the ridge to the east of the road where the Harris Canyon member of the Joana Limestone (Fig. 17) is well exposed. It represents the lower part of the Sadlick sequence and is permissibly correlative with the Island Mountain at the base of the Diamond Range sequence to the west.

Limestones of the Harris Canyon are characteristically evenly bedded; graded beds and lime mudstones represent a mixture of probable storm and hemipelagic deposits. At this exposure, soft-sediment deformation structures and rhythmic graded beds are common. The lower contact with the lower Joana (of the Morris sequence) is covered, but in nearby exposures it is sharp but gradational. The contact with the overlying Chainman Shale is a bedding-subparallel normal fault, possibly the same folded Mesozoic attenuation fault that affects the Joana-Chainman contact for at least 20 miles farther south in the White Pine Range. At this location, the contact is marked by the red jasper just above the road. Conodonts from the Harris Canyon are of both late Kinderhookian and Osagean ages.

Return to Hwy. 50. Near the intersection with Hwy. 50, the jasper zone following along the Joana-Chainman fault zone is conspicuous in the distance.

Turn left (east) onto Hwy. 50 and travel to Ely, Nevada, a distance of about 41 miles. End of Day 2. Overnight in Ely.

Day 3.

Head west out of Ely on Hwy. 50, and travel back to the Hamilton road turn off, a distance of about 36 miles. Turn left (east) onto Hamilton road

0.00 miles. Intersection of Hamilton road and Hwy. 50.

0.1 miles. Intersection with road south to Illipah Valley. Continue west.

4.9 miles. On crest of White Pine Range. Turn right (west) onto road leading down hill.

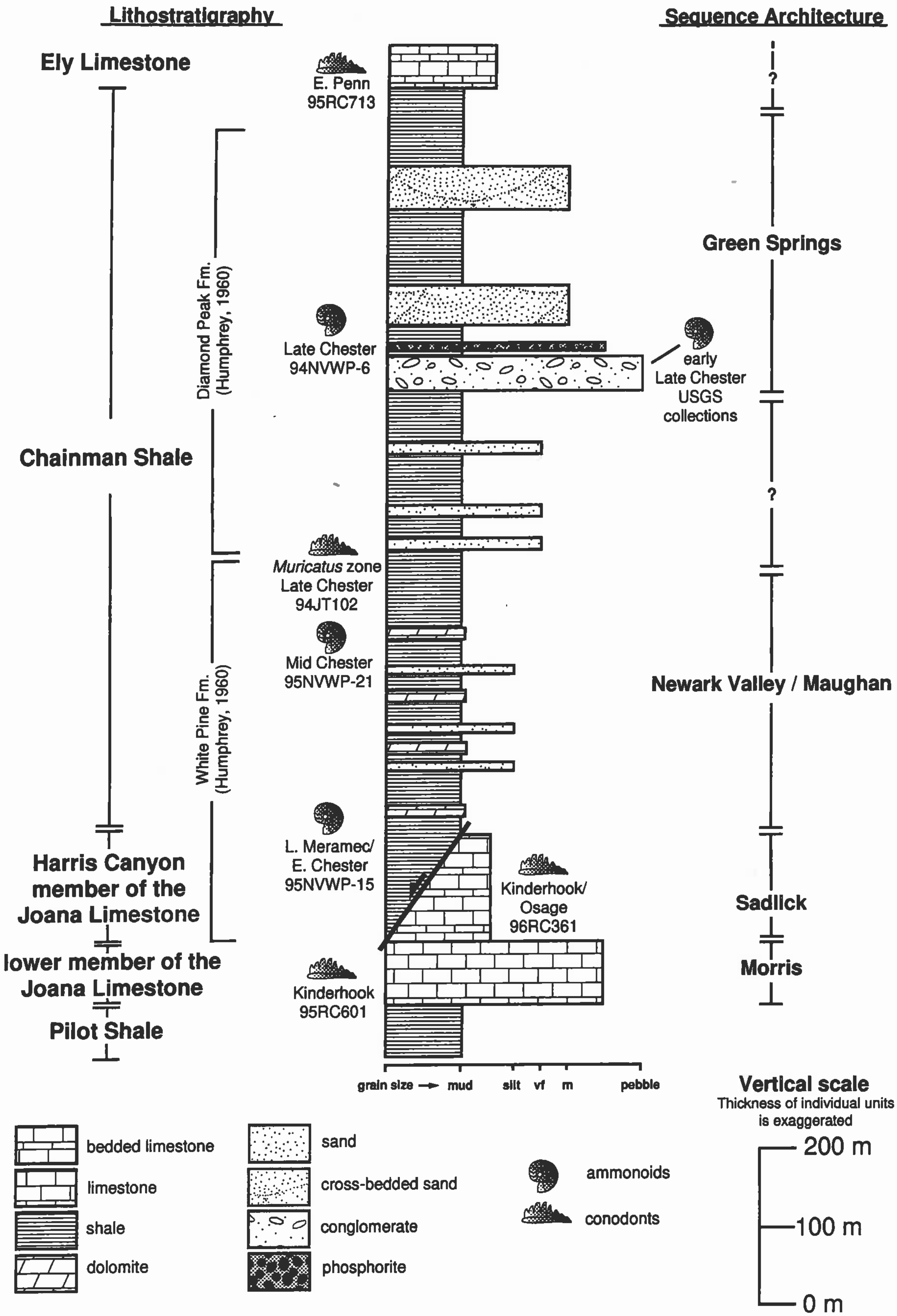
6.2 turn around and park.

Stop 10. [Center E1/2, sec. 30, T.17 N., R.58 E.; Hamilton, Nevada 7.5′ quadrangle] This is the Mokomoke Mountains section of the Mississippian (Crosbie, 1997). It corresponds to the left side of the generalized stratigraphic column of Figure 17, where the Harris Canyon member of the Joana Limestone, representing the Sadlick sequence, is cut out by a low-angle normal fault.

The thick-bedded crinoidal grainstone forming the low ridge south of the parking place is within the lower Joana Limestone, which represents the Morris sequence. Conodonts from the highest part of this limestone are Kinderhookian in age. The upper part of the Joana (the Harris Canyon member) and all expression of the Sadlick sequence are missing here due to attenuation faulting as well as probable uplift and erosion prior to deposition of the Newark Valley sequence.

Directly above the Joana is black fissile shale interbedded with thin, laterally extensive siltstones, lime mudstones, and wackestones. This recessive section was deposited in a distal prodelta to basinal setting having sediment sources

Figure 17. Stratigraphic column of the Mississippian strata in the White Pine Range (from Crosbie, 1997). Age determinations based on ammonoids (goniatites) are by Alan Titus (personal communications, 1995–1997) and those on conodonts by Mira T. Kurka (personal communication, 1997). According to Titus, goniatite collection 94NVWP-6 and the USGS collections correspond to Mamet foraminiferal zone (MFZ) 19; 95NVWP-21 to MFZ 18; and 95NVWP-15 to MFZ 16i. The inexplicably young "late Chester" age of the conodont collection designated 94JT102 and assigned to the muricatus *zone (equivalent to uppermost Mamet foraminiferal zone 19) in this section (Trexler et al., 1995) is contradicted by both the goniatite biostratigraphy and the interpretation of another conodont collection, as follows: (1) goniatites from near the level of collection 94JT102 in the Mokomoke Mountains section are age equivalent to lower Mamet zone 17, and (2) conodonts from only 62–68 m below the Ely were compared by D.L. Dunn (Sadlick, 1995) to those from the "Grove Church beds" which belong to the* unicornis *zone, the conodont zone next older from the* muricatus *zone.*


Lithostratigraphy
Sequence Architecture
Ely Limestone
E. Penn
95RC713
Diamond Peak Fm. (Humphrey, 1960)
Green Springs
Late Chester
94NVWP-6
early
Late Chester
USGS
collections
Chainman Shale
?
Muricatus zone
Late Chester
94JT102
Mid Chester
95NVWP-21
White Pine Fm. (Humphrey, 1960)
Newark Valley / Maughan
Harris Canyon
member of the
Joana Limestone
L. Meramec/
E. Chester
95NVWP-15
Kinderhook/
Osage
96RC361
Sadlick
lower member of the
Joana Limestone
Kinderhook
95RC601
Morris
Pilot Shale
grain size → mud silt vf m pebble
Vertical scale
Thickness of individual units
is exaggerated
200 m
100 m
0 m
bedded limestone
limestone
shale
dolomite
sand
cross-bedded sand
conglomerate
phosphorite
ammonoids
conodonts

from the RMA to the west. It is similar in character and age to other sections of the Newark Valley sequence to the west as well as to the Maughan sequence to the east. Beginning at the conspicuous change in slope and continuing up to the lowest outcrops of conglomerate is a lithologically distinctive unit representing a relative decrease in sea level and an influx of mature quartz sand of cratonic provenance. This unit could be either the upper highstand system tract of the Newark Valley/Maughan sequence or a separate younger sequence.

The Green Spring sequence of Trexler and Nichtman (1990) is recognized as beginning at the base of the conspicuous conglomerate unit and continuing upslope to about the level of the Hamilton road. The basal conglomerate represents lowstand braid- and delta-plain deposits, which are separated from overlying shale by a flooding, or transgressive surface. A phosphate-pebble lag a few meters above the conglomerate may signal the maximum flooding surface of the transgressive system tract. Up section, several units of nearshore quartz arenite, intercalated in the shaly succession, outline parasequences within the highstand systems tract.

The highest parasequence of the Green Springs sequence, or possibly a distinct sequence above the Green Springs, begins at about the Hamilton road below the first quartz-sandy grainstone bed and continues into the overlying lower Ely Limestone. The repetitive cliffs of Ely Limestone manifest the cyclothemic pattern well established in Pennsylvanian strata elsewhere.

Backtrack to Hwy. 50 and head east toward Ely.

In the canyon just east of the Hamilton road intersection, the conspicuously cyclic carbonate rocks along the highway are the Pennsylvanian Ely Limestone.

After crossing Robinson Summit in the Egan Range, headed towards Ely, note the steeply east dipping, overturned to the west Ely Limestone on Radar ridge in the distance south of the highway.

From Ely there are alternative routes back to Salt lake City. If the weather is nice, and late arrival in Salt lake City is no problem, take Hwy. 50 east from Ely to Stop 11 in the Confusion Range. Otherwise, head north from Ely on U.S. Highways 93 and Alternate 93 towards Wendover, visiting alternate stop 11 at White Horse Pass in the southern Goshute Mountains (see Fig. 14).

For stop 11, head east from Ely on Hwy. 50.

Leaving Ely, the geology to about the Nevada-Utah state line is dominated by the effects of large-magnitude Tertiary extension related to the Northern and the Southern Snake Range decollements (respectively the NSRD and SSRD; Gans and Miller, 1983; McGrew, 1993). The NSRD manifests a greater amount of east-west extension than the SSRD, and the accommodation zone between them projects more or less along the route of Hwy. 50. The structural style of these extensional systems is down-to-the-west tilting in the hanging walls of multiple east-dipping normal faults. About 250% extension on the NSRD from the Egan Range across the Snake Range is interpreted by Gans and Miller (1983). Palinspastically the pre-Tertiary rocks of the Confusion Range thus were originally much closer to those near Ely, although the degree to which this compensates for the effect of possible Mesozoic contractional structures is an open question.

On the west flank of the Schell Creek Range and near Connors Pass most of the exposures near the highway are of conspicuously disrupted Pennsylvanian and Permian strata. Low-angle normal faults separate these rocks from lightly metamorphosed Cambrian strata on the east flank of the range.

Looking east from the east flank of the Schell Creek Range across to the Southern Snake Range, the high point is Wheeler Peak (elev. 13,063 feet), the second highest peak in Nevada. Wheeler Peak is in the footwall of the SSRD, which crosses the range crest above the prominent cliffs of Cambrian limestone about five miles south of Wheeler Peak.

Following Hwy. 50 up the west slope of the Snake Range towards Sacramento Pass, the steeply east-dipping Paleozoic strata north of the highway, on the skyline of a spur off the Northern Snake Range, are overturned, having been multiply tilted down to the west past the vertical in the hanging wall of the NSRD.

About 10 miles east of Sacramento Pass, on the north side of the highway, the gently east-dipping trace of the NSRD is about half way up on the south-facing slopes of the Northern Snake Range. Brittle-deformed west-dipping Paleozoic strata of the hanging wall can be seen to butt into the footwall, which is formed of mylonites whose shear layering parallels the NSRD (Miller et al., 1987).

[At the Nevada-Utah border reenter the Mountain time zone and lose an hour.]

At Utah Milepost 16 on U.S. Hwy. 50, turn left (north) onto improved gravel road to Little Valley.

0.00 miles at intersection.

3.7 miles. Intersection; stay right, following road-sign direction to Conger Spring.

At 4.1 miles, take the left (north) fork of the road towards Conger Spring.

At 6.5 miles, pass side road to the right (east).

8.45 miles. Four-way intersection. Turn right (east) onto road to Camp Canyon and Skunk Spring. After about two miles this road enters the long, NNE-trending, strike valley between the Joana Limestone cuesta to the east and the Ely Limestone to the west. Little Mile and a Half Canyon forms the second of the drainage gaps through the Joana Limestone.

14.5 miles. North side of the drainage forming the third gap through the Joana. Park where convenient near here.

Stop 11. [SE1/4, sec. 17, T.18 S., R.16 W.; Conger Mountain, Utah 7.5′ quadrangle] Walk down section across the poorly exposed Needle Siltstone into the gorge through the upper Joana Limestone to the major dry fall, which is formed by the massive lime mudstone unit (MO.Dc.Lm) at the top of the lower Joana Limestone and the Morris sequence. This is the base of the columnar section of the Sadlick sequence shown on Figure 5.

Carbonate rocks of the lower part of the Sadlick sequence (the upper Joana Limestone) form a retrogradational parasequence set; they reflect protracted, though cyclic, drowning of the carbonate platform prior to onset of the Delle phosphatic event, onset of which corresponds to the base of the Needle Siltstone. The units or lithosomes labeled SA.Rh.Wk ("Sadlick sequence rhythmically bedded wackestone," Silberling et al., 1995) are interpreted as relatively deep subtidal, but above storm wave base deposits. They form the lower parts of four parasequences, the lowest of which grades up through crinoidal grainstones to oolitic grainstone at its top. The upper parts of the weakly differentiated second and the more obvious third parasequence are crinoidal grainstone and packstone. The fourth parasequence, at the top of the Joana, begins with lithosome SA.Rh.Wk which is directly overlain by deposits of the Delle phosphatic event. The upper part of each parasequence thus represents deposition in successively deeper-water, and following onset of the Delle event, subsidence below storm wave base was accompanied by restriction of the distal Antler foreland.

To examine the section of the Maughan sequence, continue north for about 2.5 miles on the Camp Canyon-Skunk Springs road and park on the limestone knoll around which the road passes to the west (in the central W1/2, sec. 4, T.18 S., R.16 W.; Cowboy Pass, Utah 7.5′ quadrangle). This limestone is lithosome MA.En.Gr in the lower part of the Maughan sequence (see Fig. 10).

Backtrack to Hwy 50 and continue east to Delta, Utah, U.S. Hwy. 6, Utah Hwy. 132 and U.S. I-15 north to Salt Lake City. Driving time from the Confusion Range turnoff is about 3-1/2 hours. End of day 3 and field trip by way of Stop 11.

If the alternate return to Salt Lake City from Ely is chosen, head north from Ely on U.S. Hwy. 93 and Alternate 93 to the vicinity of White Horse Pass in the Goshute Range.

About a mile or so <u>west</u> of the pass, cut over onto the old highway surface that parallels Hwy Alt. 93 on its northwest side. At 0.75 miles west of the pass turn north onto dirt track leading into a broad valley. Follow this track for about 0.8 miles, where it will probably be impassable. Park here.

Alternate Stop 11. [SW1/4 NE1/4 sec. 25, T.29 N. R.68 E.; White Horse Pass, Nev. 7.5′ quadrangle] By ranging around this valley to the west and north within 1/2 mile of the parking place, something can be seen of the Needle Siltstone (the upper part of the Sadlick sequence) and the sequence boundary at its top with the limestone of White Horse Pass (the lower part of the Maughan sequence). However, the main purpose of this stop is to illustrate that understanding the pre-Tertiary stratigraphy here, as in most of the Great Basin, requires an understanding of the structure, which may not be obvious.

The Mississippian rocks at White Horse Pass are in the footwall of an extensive low-angle Tertiary fault system. As in most of east-central Nevada the older Tertiary volcanic rocks in the hanging walls of these systems, though tilted, are roughly concordant with the depositionally underlying pre-Tertiary strata, which tend to be the youngest in the region (in this case of Permian and Triassic age). The meaning of this, as pointed out in the classic paper on "denudation" faults in the Sevier hinterland by Armstrong (1972), is that major Mesozoic folds and thrust ramps are not seen to affect the youngest pre-Tertiary strata in this region. Nevertheless, sufficient contractional and extensional strain of Mesozoic age is recorded in the Mississippian and older rocks of the White Horse Pass area to badly mislead the unwary and confuse the interpretation of the Antler foreland.

Work in progress and unpublished mapping by J.E. Welsh and by R.W. Allmendinger document a complicated progression of Mesozoic structural events. The earliest structures (D1) are map-scale folds associated with the contacts of the Jurassic pluton south of White Horse Peak. Widespread development of foliation in the limestone of White Horse Pass may also have resulted from D1. As a younger phase of D1, attenuation (bedding subparallel, younger on older) faults then faulted the Needle Siltstone, and locally the Joana Limestone beneath the Needle, down onto the folded Devonian carbonate rocks and faulted higher parts of the Chainman Shale onto the limestone of White Horse Pass. Following this, two generations of contractional structures (D2 and D3) involve the attenuation faults in folds and minor thrust faults. The lower contact of the Needle in the immediate vicinity of White Horse Pass is actually the lower of the two Mesozoic attenuation faults, although this would not be suspected without mapping and carrying out structural studies over a fairly large area. Calling upon an original depositional high to explain the absence of the Joana Limestone and Pilot Shale at White Horse Pass (Sadlick, 1965; Silberling et al., 1995) unnecessarily complicates the paleogeography of the Antler foreland.

Backtrack to Alt. Hwy. 93, continue on to Wendover, and return east on I-80 to Salt Lake City. Driving time from

White Horse Pass to Salt Lake City is about 2-1/2 hours. End of Day 3 and field trip by way of alternate Stop 11.

REFERENCES CITED

Anadón, P., Cabrera, L., Columbo, F., and Marzo, M.R., 1986, Syntectonic intraformational unconformities in alluvial fan deposits, eastern Ebro Basin margins (NE Spain), *in* Allen, P.A., and Homewood, P., eds., Foreland Basins: International Association of Sedimentologists Special Publication 8, London, Blackwell, p.259–271.

Armstrong, R.L., 1972, Low-angle (denudation) faults, hinterland of the Sevier orogenic belt, eastern Nevada and western Utah: Geological Society of America Bulletin, v. 83, p. 1729–1754.

Burchfiel, B.C., and Davis, G.A., 1972, Structural framework and evolution of the southern part of the Cordilleran orogen, western United States: American Journal of Science, v. 272, p. 97–118.

Baker, P.A., and Kastner, M., 1981, Constraints on the origin of sedimentary dolomite: Science, v. 213, p.214–216.

Camilleri, P.A., 1994, Mesozoic and Cenozoic tectonics and metamorphic evolution of the Wood Hills and Pequop Mountains, Elko County, Nevada [Ph.D. dissert.]:Laramie, University of Wyoming, 196 p.

Carlisle, D., and Nelson, C.A., 1990, Geologic map of the Mineral Hill quadrangle, Nevada: Nevada Bureau of Mines and Geology, Map 97, scale 1:48,000.

Carpenter, J.A., Carpenter, D.G., and Dobbs, S.W., 1993, Structural analysis of the Pine Valley area, Nevada, *in* Gillespie, C.W., ed., Structural and stratigraphic relationships of Devonian reservoir rocks, east central Nevada: Reno, Nevada Petroleum Society, 1994 Field Conference Guidebook, p. 9–49.

Carpenter, J.A., Carpenter, D.G., and Dobbs, S.W., 1994, Antler orogeny—Paleostructural analysis and constraints on plate tectonic models with a global analogue in southeast Asia, *in* Dobbs, S.W., and Taylor, W.J., eds., Structural and stratigraphic investigations and petroleum potential of Nevada, with special emphasis south of the Railroad Valley producing trend: Reno, Nevada Petroleum Society, 1994 Field Conference Guidebook, p. 187–240.

Cashman, P.H., 1992, Structural geology of the northeastern Stansbury Mountains, *in* Wilson, J.R., ed., Field guide to geologic excursions in Utah and adjacent areas of Nevada, Idaho, and Wyoming: Utah Geological Survey, Miscellaneous Publication 92-3, p. 171–178.

Crosbie, R.A., 1997, Sequence Architecture of Mississippian Strata in the White Pine Mountains, White Pine County, Nevada [Master's thesis]: Reno, University of Nevada, 200 p.

DeCelles, P.G., and Giles, K.A., 1996, Foreland basin systems: Basin Research, v. 8, p. 105–123.

Dott, R.H., Jr., 1955, Pennsylvanian stratigraphy of Elko and northern Diamond Ranges, northeastern Nevada: American Association of Petroleum Geologists Bulletin, v. 39, p. 2211–2305.

Ekburg, C.K., Rota, J.C., and Arkell, B.W., 1991, Geology and mineral deposits of the Maggie Creek subdistrict, Carlin trend, Eureka County, Nevada, *in* Raines, G.L., Lisle, R.E., Schafer, R.W., and Wilkinson, W.H., eds., Geology and ore deposits of the Great Basin: Reno, Geological Society of Nevada, p. 625–633.

Gans, P.B., and Miller, E.L., 1983, Style of mid-Tertiary extension in east-central Nevada, *in* Gurgel, K.D., ed., Geologic excursions in the overthrust belt and metamorphic core complexes of the intermountain region, Guidebook Part 1—GSA Rocky Mountain and Cordilleran Sections Meeting, Salt Lake City, Utah—May 2–4, 1983: Utah Geological and Mineral Survey Special Studies 59, p. 107–160.

Geslin, J.K., 1993, Evolution of the Pennsylvanian-Permian Oquirrh-Wood River Basin, southern Idaho [Ph.D. Dissertation]: Los Angeles, University of California at Los Angeles, 272 p.

Giles, K.A., and Dickinson, W.R., 1995, The interplay of eustasy and lithospheric flexure in forming stratigraphic sequences in foreland settings—an example from the Antler foreland, Nevada and Utah, *in* Dorobek, S.L., and Ross, G.M., eds., Stratigraphic evolution of foreland basins: Tulsa, SEPM (Society for Sedimentary Geology), Special Publication 52, p. 187–211.

Goebel, K.A., 1991, Paleogeographic setting of Late Devonian to Early Mississippian transition from passive to collisional margin, Antler foreland, eastern Nevada and western Utah, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the western United States—II: Pacific Section SEPM, v. 67, p. 401–418.

Gordon, M., Jr., 1986, Late Kinderhookian (Early Mississippian) ammonoids of the western United States: Paleontological Society Memoir 19, 36 p.

Gutschick, R.C., Sandberg, C.A., and Sando, W.J., 1980, Mississippian shelf margin and carbonate platform from Montana to Nevada, *in* Fouch, T.D., and Magathan, E.R., eds., Paleozoic paleogeography of the west-central United States: Rocky Mountain Section, S.E.P.M., Rocky Mountain Paleogeography Symposium 1, p. 111–128.

Harbaugh, D.W., and Dickinson, W.R., 1981, Depositional facies of Mississippian clastics, Antler foreland basin, central Diamond Mountains, Nevada: Journal of Sedimentary Petrology, v. 51, p. 1223–1234.

Hansen, J.B., and Ransom, K.L., 1990, Tomera Ranch Field, Pine Valley, Eureka County, Nevada, *in* Flanigan, D.M.H., Garside, L.J., and Hansen, M., eds., Oil fields and geology of the Pine Valley, Eureka County area, Nevada: Reno, Nevada Petroleum Society, p. 35.

Hintze, L.F., 1980, Geologic map of Utah: Utah Geological and Mineral Survey, scale 1:500,000.

Hintze, L.F., 1988, Geologic history of Utah: Brigham Young University Geology Studies Special Publication 7, 202 p.

Howard, K.A., 1992, Ruby Mountains metamorphic core complex—Deep crustal exposures exhumed from beneath the Pinon Range?, *in* Trexler, J.H., Jr., Flanigan, T.E., Flanigan, D.M.H., Hansen, M., and Garside, L., eds., Structural geology and petroleum potential of southwest Elko County, Nevada: Reno, Nevada Petroleum Society, 1992 Annual Field Trip Guidebook, p. 57–60.

Humphrey, F.L., 1960, Geology of the White Pine mining district, White Pine County, Nevada: Nevada Bureau of Mines Bulletin 57, 119 p.

Jansma, P.E., 1988, The tectonic interaction between an oceanic allochthon and its foreland basin during continental overthrusting; the Antler orogeny, Nevada [Ph.D. Dissertation]: Evanston, Illinois, Northwestern University, 316 p.

Jansma, P.E., and Speed, R.C., 1990, Omissional faulting during Mesozoic regional contraction at Carlin Canyon, Nevada: Geological Society of America Bulletin, v. 102, p. 417–427.

Jansma, P.E., and Speed, R.C., 1993, Deformation, dewatering, and decollement development in the Antler foreland basin during the Antler orogeny: Geology, v. 21, p. 1035–1038.

Jansma, P.E., and Speed, R.C., 1995, Kinematics of underthrusting in the Paleozoic Antler foreland basin: Journal of Geology, v. 103, p. 559–575.

Jewell, P.W., Silberling, N.J., and Nichols, K.M., 1996, Restriction and hyposalinity in the mid-Mississippian record of the distal Antler foreland, eastern Great Basin: Geological Society of America Abstracts with Programs, v. 28, No. 7, p. A-368.

Johannesen, D.C., and Cole, M.R., 1990, Blackburn oil field, Eureka County, Nevada, in Flanigan, D.M.H., Garside, L.J., and Hansen, M., eds., Oil fields and geology of the Pine Valley, Eureka County area, Nevada: Reno, Nevada Petroleum Society, p. 39–53.

Johnson, J.G., and Pendergast, A., 1981, Timing and mode of emplacement of the Roberts Mountains allochthon, Antler orogeny: Geological Society of America Bulletin, v. 92, p. 648–658.

Johnson, J.G., and Visconti, R., 1992, Roberts Mountains thrust relationships in a critical area, northern Sulphur Spring Range, Nevada: Geological Society of America Bulletin, v. 104, p. 1208–1220.

Keith, M.L., and Weber, J.N., 1964, Carbon and oxygen isotopic compositions of selected limestones and fossils: Geochimica et Cosmochemica Acta, v. 28, p. 1787–1816.

Ketner, K.B., 1977, Late Paleozoic orogeny and sedimentation, southern California, Nevada, Idaho, and Montana, *in* Stewart, J.H., Stevens, C.H., and Fritsche, A.E., eds., Paleozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 1, p. 363–369.

Ketner, K.B., and Smith, J.F., Jr., 1982, Mid-Paleozoic age of the Roberts thrust unsettled by new data from northern Nevada: Geology, v. 10, p. 298–303.

Ketner, K.B., Murchey, B.L., Stamm, R.G., and Wardlaw, B.R., 1993, Paleozoic and Mesozoic rocks of Mount Ichabod and Dorsey Canyon, Elko County, Nevada—Evidence for post-Early Triassic emplacement of the Roberts Mountains and Golconda allochthons: U.S. Geo- logical Survey Bulletin 1988-D, 12 p.

Kirk, E., 1933, The Eureka quartzite of the Great Basin region: American Journal of Science, v. 28, p. 27–43.

Kluth, C.J., and Coney, P.J., 1981, Plate tectonics of the Ancestral Rocky Mountains: Geology, v. 9, p. 10–15.

Link, P.K., Warren, I., Preacher, J.M., and Skipp, B., 1996, Stratigraphic analysis and interpretation of the Mississippian Copper Basin Group, McGowan Creek Formation, and White Knob Limestone, south-central Idaho, *in* Longman, M.W., and Sonnenfeld, M.D., eds., Paleozoic systems of the Rocky Mountain region: Rocky Mountain Section, SEPM (Society for Sedimentary Geology), p. 117–144.

Macke, D.L., Silberling, N.J., and Nichols, K.M., 1994, Ramp-margin wedge deposits in the Upper-Devonian Lower Mississippian Pilot Shale, eastern Nevada and western Utah: Geological Society of America Abstracts with Programs, v. 26, no. 2, p. 68–69.

McGrew, A.J., 1993, The origin and evolution of the southern Snake Range decollement, east central Nevada: Tectonics, v. 12, p. 21–34.

Merriam, C.W., and Anderson, C.A., 1942, Reconnaissance survey of the Roberts Mountains, Nevada: Geological Society of America Bulletin, v. 53, p. 1675–1728.

Miller, E.L., Gans, P.B., and Lee, Jeffrey, 1987, The Snake Range decollement, eastern Nevada: Geological Society of America Centennial Field Guide—Cordilleran Section, p. 77–82.

Moores, E., 1970, Ultramafics and orogeny, with models of the U.S. Cordillera and Tethys: Nature, v. 228, p. 837–842.

Murphy, M.A., Power, J.D., and Johnson, J.G., 1984, Evidence for Late Devonian movement within the Roberts Mountains allochthon, Roberts Mountains, Nevada: Geology, v. 12, p. 20–23.

Nichols, K.M., and Silberling, N.J., 1990, Delle Phosphatic Member—An anomalous phosphatic interval in the Mississippian (Osagean-Meramecian) shelf sequence of central Utah: Geology, v. 18, p. 46–49.

Nichols, K.M., and Silberling, N.J., 1993, Upper Devonian to Mississippian strata of the Antler foreland in the Leppy Hills, easternmost northern Nevada: U.S. Geological Survey Bulletin 1988-G, p. G1–G13.

Nichols, K.M., and Silberling, N.J., 1995, Initial foredeep deposits of the mid-Mississippian episode of Antler flexural-loading tectonics in the vicinity of Eureka, Nevada, *in* Hansen, M.W., and Trexler, J.H., Jr., eds., Mississippian source rocks in the Antler basin and associated structural and stratigraphic traps: Reno, Nevada, Nevada Petroleum Society, 1995 Fieldtrip Guidebook, p. 87–96.

Nilsen, T.H., and Stewart, J.H., 1980, The Antler orogeny—Mid-Paleozoic tectonism in western North America: Geology, v. 8, p. 298–302.

Nolan, T.B., 1974, Stratigraphic evidence on the age of the Roberts Mountains thrust, Eureka and White Pine Counties, Nevada: U.S. Geological Survey Journal of Research, v. 2, no. 4, p. 411–416.

Nolan, T.B., Merriam, C.W., and Blake, M.C., Jr., 1974, Geologic map of the Pinto Summit quadrangle, Eureka and White Pine Counties, Nevada: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-793, scale 1:31,680.

Perry, A.J., 1995, Depositional setting of the Upper Mississippian to Lower Pennsylvanian Newark Valley sequence, Diamond Range, Nevada, *in* Hansen, M.W., and Trexler, J.H., Jr., eds., Mississippian source rocks in the Antler basin and associated structural and stratigraphic traps: Reno, Nevada Petroleum Society, 1995 Fieldtrip Guidebook, p. 97–113.

Poole, F.G., 1974, Flysch deposits of the Antler foreland basin, western United States, *in* Dickinson, W.R., ed., Tectonics and sedimentation: Society of Economic Paleontologists and Mineralogists, Special Publication 22, p. 58–82.

Poole, F.G., and Claypool, G.E., 1984, Petroleum source-rock potential and crude-oil correlation in the Great Basin, *in* Woodward, J., Meissner, F.F., and Clayton, J.L., eds., Hydrocarbon source rocks of the greater Rocky Mountain region: Denver, Colorado, Rocky Mountain Association of Geologists, p. 179–207.

Poole, F.G., and Sandberg, C.A., 1991, Mississippian paleogeography and conodont biostratigraphy of the western United States, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the western United States-II: Pacific Section SEPM, v. 67, p. 107–136.

Putnam, B.R., III, and Henriques, E.Q.B., 1991, Geology and mineralization at the South Bullion deposit, Pinon Range, Elko County, Nevada—implications for western United States Cenozoic tectonics, *in* Raines, G.L., Lisle, R.E., Schafer, R.W., and Wilkinson, W.H., eds., Geology and ore deposits of the Great Basin—Symposium proceedings: Reno, Nevada, Geological Society of Nevada, p. 713–728.

Roberts, R.J., 1949, Structure and stratigraphy of the Antler Peak quadrangle, north-central Nevada [Abstract]: Geological Society of America Bulletin, v. 60, p. 1917.

Roberts, R.J., Hotz, P.E., Gilluly, J., and Ferguson, H.G., 1958, Paleozoic rocks of north-central Nevada: American Association of Petroleum Geologists Bulletin, v. 42, p. 2813–2857.

Ross, C.A., and Ross, J.R.P., 1987, Late Paleozoic sea levels and depositional sequences: Cushman Foundation for Foraminiferal Research, Special Publication 24, p. 137–149.

Sadlick, Walter, 1960, Some preliminary aspects of Chainman stratigraphy, *in* Boettcher, J.W., and Sloan, W.W., Jr., eds., Guidebook to the geology of east central Nevada:

Salt Lake City, Intermountain Association of Petroleum Geologists, 11th Annual Field Conference Guidebook, p. 81–90.

Sadlick, Walter, 1965, Biostratigraphy of the Chainman Formation (Carboniferous), eastern Nevada and western Utah [Ph.D. Dissertation]: Salt Lake City, University of Utah, 228 p.

Sadlick, Walter, 1995, Stratigraphic subdivision of the Chainman Formation, *in* Hansen, M.W., and Trexler, J.H., Jr., eds., Mississippian source rocks in the Antler basin of Nevada and associated structural traps: Reno, Nevada, Nevada Petroleum Society, Inc., 1995 Fieldtrip Guidebook, p. 3–43.

Sandberg, C.A., and Gutschick, R.C., 1984, Distribution, microfauna, and source-rock potential of Mississippian Delle Phosphatic Member of Woodman Formation and equivalents, Utah and adjacent states, *in* Woodward, J., Meissner, F.F., and Clayton, J.L., eds., Hydrocarbon source rocks of the Greater Rocky Mountain region: Denver, Colorado, Rocky Mountain Association of Geologists, p. 135–178.

Schneyer, J.D., 1990, Geologic map of the Leppy Peak quadrangle and adjacent area, Elko County, Nevada, and Toole County, Utah: U.S. Geological Survey Miscellaneous Investigations Map I-1938, scale 1:24,000.

Silberling, N.J., and Nichols, K.M., 1991, Petrology and regional significance of the Mississippian Delle Phosphatic Member, Lakeside Mountains, northwestern Utah, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States—II: 425–438, Pacific Section SEPM, v. 67.

Silberling, N.J., and Nichols, K.M., 1992, Depositional cycles of the Upper Devonian-Lower Mississippian limestone succession in the southern Lakeside Mountains, *in* Wilson, J.R., ed., Field guide to geologic excursions in Utah and adjacent areas of Nevada, Idaho, and Wyoming: Utah Geological Survey, Miscellaneous Publication 92-3, p. 131–145.

Silberling, N.J., Nichols, K.M., Macke, D.L., and Trappe, J., 1995, Upper Devonian-Mississippian stratigraphic sequences in the distal Antler foreland of western Utah and adjoining Nevada: U.S. Geological Survey Bulletin 1988-H, p. H1–H33.

Sinclair, H.D., 1997, Tectonostratigraphic model for underfilled peripheral foreland basins—an Alpine perspective: Geological Society of America Bulletin, v. 109, p. 324–346.

Smith, J.F., Jr., and Ketner, K.B., 1968, Devonian and Mississippian rocks and the date of the Roberts Mountains thrust in the Carlin-Pinon Range area, Nevada: U.S. Geological Survey Bulletin 1251-I, 18 p.

Smith, J.F., Jr., and Ketner, K.B., 1975, Stratigraphy of Paleozoic rocks in the Carlin-Pinon Range area, Nevada: U.S. Geological Survey Professional Paper 867-A, 87 p.

Smith, J.F., Jr., and Ketner, K.B., 1978, Geologic map of the Carlin-Pinon Range area, Elko and Eureka Counties, Nevada: U.S. Geological Survey Miscellaneous Investigations Map I-1028, scale 1:62,500.

Speed, R.C., and Sleep, N.H., 1982, Antler orogeny and foreland basin—a model: Geological Society of America Bulletin, v. 93, p. 815–828.

Stewart, J.H., and Carlson, J.E., 1978, Geologic map of Nevada: U.S. Geological Survey, scale 1:500,000.

Thoreson, R.F., 1991, Geology and gold deposits of the Rain subdistrict, Elko County, Nevada, *in* Raines, G.L., Lisle, R.E., Schafer, R.W., and Wilkinson, W.H., eds., Geology and ore deposits of the Great Basin—Symposium proceedings: Reno, Nevada, Geological Society of Nevada, p. 635–642.

Trexler, J.H., Jr., 1990, Paleozoic stratigraphy and tectonic history in the Carlin-Pinon Range area, a review, *in* Flanigan, D.M.H., Garside, L.J., and Hansen, M., eds., Oil fields and geology of the Pine Valley, Eureka County area, Nevada: Reno, Nevada Petroleum Society, p. 5–11.

Trexler, J.H., Jr., 1992, The Devonian-Mississippian Stansbury Formation at Stansbury Island and in the northeastern Stansbury Mountains, Utah, *in* Wilson, J.R., ed., Field guide to geologic excursions in Utah and adjacent areas of Nevada, Idaho, and Wyoming: Utah Geological Survey, Miscellaneous Publication 92-3, p. 161–169.

Trexler, J.H., Jr., and Cashman, P.H., 1991, Mississippian stratigraphy and tectonics of east-central Nevada: post-Antler orogenesis, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States—II: Pacific Section SEPM, v. 67, p. 331–342.

Trexler, J.H., Jr., Cole, J.C., and Cashman, P.H., 1996, Middle Devonian-Mississippian stratigraphy on and near the Nevada Test Site—Implications for hydrocarbon potential: American Association of Petroleum Geologists Bulletin, v. 80, p. 1736–1762.

Trexler, J.H., Jr., and Nitchman, S.P., 1990, Sequence stratigraphy and evolution of the Antler foreland basin, east-central Nevada: Geology, v. 18, p. 422–425.

Trexler, J.H., Jr., Snyder, W.S., Cashman, P.H., Gallegos, D.M., and Spinosa, C., 1991, Mississippian through Permian orogenesis in eastern Nevada: post-Antler, pre-Sonoma tectonics of the western Cordillera, *in* Cooper, J.D., and Stevens, C.H., eds., Paleozoic paleogeography of the Western United States—II: Pacific Section SEPM, v. 67, p. 317–329.

Trexler, J.H., Jr., Snyder, W., Schwarz, D., Kurka, M.T., and Crosbie, R., 1995, An overview of the Mississippian Chainman Shale, *in* Hansen, M.W., and Trexler, J.H., Jr., eds., Mississippian source rocks in the Antler basin of Nevada and associated structural traps: Reno, Nevada Petroleum Society, 1995 Fieldtrip Guidebook, p. 45–60.

Webster, G.D., Gordon, M., Jr., Langenheim, R.L., Jr., and Henry, T.W., 1984, Road logs for the Mississippian-Pennsylvanian boundary in the eastern Great Basin, *in* Lintz, Joseph, ed., Western geological excursions: Reno, Mackay School of Mines, v. 1, p. 1–86.

Whitaker, J.H., 1985, Geology of the Willow Creek area, Elko County, Nevada [Master's thesis]: Corvallis, Oregon State University, 116 p.