GEOLOGY OF THE JACKSON MOUNTAINS,
NORTHWEST NEVADA

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ABSTRACT

The Jackson Mountains are located in the western Great Basin in Humboldt County, northwest Nevada. The range contains a Late Paleozoic to Mesozoic depositional sequence. This sequence records sedimentation, volcanism and deformation in a back-arc setting. The Mississippian to late Early Permian McGill Canyon Formation was deposited in basinal to slope to distal shelf environments, dominated by hemipelagic and turbiditic facies. In the Permian there was an volcanic arc andesite component, and a nearby contemporaneous carbonate platform shed olistostromes into the unit. The McGill Canyon was laid down in an area between the McCloud arc and the Havallah back-arc basin. The late Middle Triassic to middle Norian Bliss Canyon Formation was laid down in basinal to fore-reef to carbonate platform to lagoonal to terrigenous littoral environments. Both of these formations are offlap sequences deposited on an east-facing, back-arc margin. The Bliss Canyon represent the western margin of the Early Mesozoic marine basin of the western Great Basin. From the late Norian to the Bathonian, several stages of subearial volcanism and alluvial epiclastic sedimentation laid down the Happy Creek Formation, a thick arc andesite volcanic pile. The Happy Creek is part of the Early Mesozoic Cordilleran magmatic arc province. In the Bathonian, this volcanic pile was cut by a conjugate sinistral high-angle wrench fault system as volcanism waned. During the Callovian, sediments of the King Lear Formation filled in and then overlapped the wrench basins. These sediments were derived from the east, where a west-vergent thrust system was active. This phase of thrusting ceased by the

Oxfordian. Arc-related silicic volcanism and alluvial to fluvial sedimentation within the King Lear continued into the Aptian, when the thrusts were reactivated during a second phase. Both phases of thrusting verged both east and west. Stocks, dikes and sills of the Early Mesozoic Intrusive Suite are comagmatic with the volcanism in the Happy Creek and King Lear, and intrude the sedimentary units. This suite both plugs and is truncated by the wrench faults and the first phase of thrusting, but is cut by the second phase. The Jackson Mountains are part of the Black Rock terrane in northwest Nevada. Within this terrane, the rocks share a common tectonic history and stratigraphy distinct from the neighboring terranes, and are separated from them by Mesozoic thrust and strike-slip faults.

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CHAPTER ONE:
INTRODUCTION

The Jackson Mountains are a north-south elongate range lying within the Basin and Range province in the northwestern corner of Nevada, in Humboldt County. The surrounding desert basins are at 3900 feet elevation, and the range reaches elevations over 9000 feet. The basin to the west is the Black Rock Desert, while that to the east is appropriately named Desert Valley. Relief along the flanks of the Jackson Mountains is quite steep. The range is unpopulated and is used for cattle range by ranches scattered about the flanks; the cattle appear quite content with the arrangement and are reluctant to vacate the premises, especially when on a road that somebody needs to use. Road access independant of bovine obstructions is poor to worse. As 4WD is a necessity in most of the range, a vehicle was kindly provided by the Division of Geological and Planetary Sciences at Caltech. The main hazards of field work in the Jackson Mountains are rattlesnakes (who are actually gentlemen for snakes, as they warn you when they're feeling ornery), weather, terrain and loneliness. The local coyotes also like to give impromptu operatic concertos in the early morning hours, but have proven sensitive to loud and firmly expressed critical review.

The field area mapped extends from 41°7'30" to 41°30' and 118°15' to 118°37'30", and included the pre-Cenozoic bedrock, with a total area of about 500 km^2. Climatic conditions are semiarid (with heavy snow in the winter), and summer temperatures varied from very hot to below freezing,

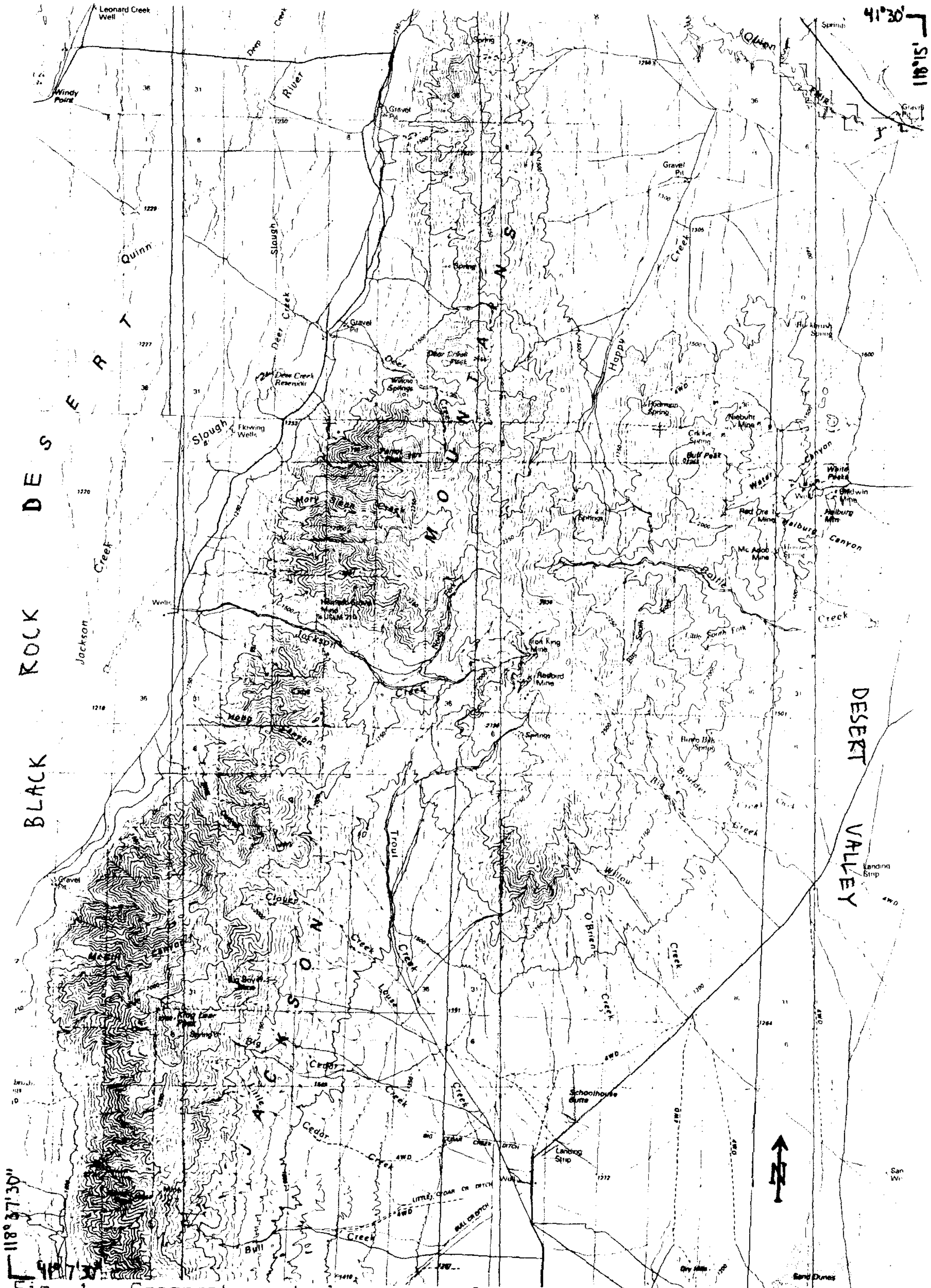


Fig. 1 Geography and placenames of the mapped field area in the Jackson Mountains, NW Nevada. Scale approximately 1:200,000.

with high wind a common problem. Blizzards in September are not unknown, though certainly unasked-for. Sage brush covers the lower slopes, juniper patches spot the higher elevations, and aspen groves shelter in the numerous stream valleys with year-round running water. Rock exposures are rather good where slope-wash is not too thick, although desert varnish can unkindly obscure lithologic features. The desert varnish, laterally heterogenous geology, shallow slope-wash cover and extensive jointing in the rocks of the range often make it difficult to recognize lithologies at a distance, particularly on air photos. Close-spaced traverses were undertaken over the entire field area (see fig. 1) to reliably characterize the geology. The rugged nature and remoteness of the terrain made it necessary (or at least much safer) to have a field assistant at all times. This strenuous chore was shared by my wife, Linda Maher, my father, Harmon Maher Sr., and by David Bissiri and Mike Kokinos. Several summer field mapping courses taught by Prof. Jason Saleeby and a paleomagnetic sampling field trip with Prof. Joseph Kirschvink (both from Caltech) also aided greatly in data collection.

During the field traverses detailed structural and stratigraphic observations were made, along with the collection of several hundred samples for later biostratigraphic, petrographic, and geochemical analysis. Separate trips were made for bulk samples for U/Pb zircon geochronology and to drill an additional several hundred samples for paleomagnetic analysis. There is nothing quite like packing 120 lb of diorite or a rock drill and gallons of water several miles to test your devotion to basic scientific research. Mapping was on 1:18,000 scale, on expanded U.S.G.S. topographic 7.5′ and 15′ base maps with contour intervals of 20 or 40 feet (mapping was done on the 7.5′ quadrangles contained within the King Lear, Donna Schee, and Bottle Creek 15′ quadrangles, plus the Hobo Canyon 7.5′ quadrangle). The summers of 1984 to 1986 and parts of the summers of 1983 and 1987 were spent mapping, for a total of about 7 memorable (and highly enjoyable) months. Several weeks of mapping were also done in the neighboring Pine Forest Range in 1987 with Jay Lieske, but this work will not be discussed in great detail (certainly not the kind of detail I plan to subject the innocent reader to in a few pages).

Previous reconnaissance mapping of the Jackson Mountains includes that of Willden (1963, 1964) for the Nevada Bureau of Geology, and Russell (1981, 1984) for a Ph.D. thesis under Prof. R. Speed at Northwestern University. These studies supplied a base for my own detailed field work, and originally defined the major rock units and indicated their ages and some of their significant relationships. My own contribution was detailed examination, description, and redefinition of the rock units and their relationships, and accurate characterization of the nature and history of the igneous activity and the depositional and deformational phases in the range.

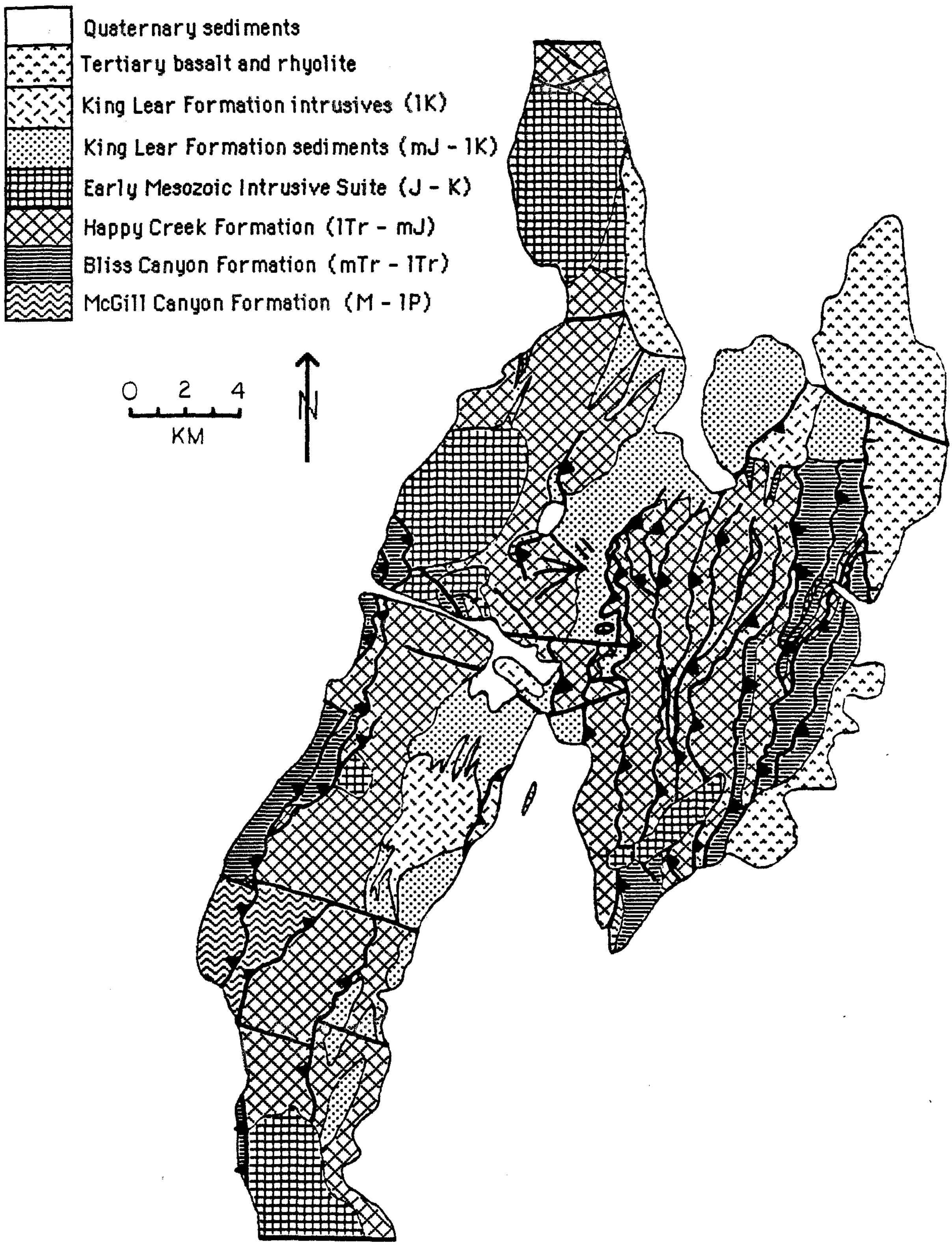
The geological signficance of the Jackson Mountains lies in its geographic position and the excellent preservation of its original geologic elements through its complex history. The range lies at the northern terminus of the Paleozoic and Mesozoic belts of the southern United States Cordillera, just before the extensive Tertiary and Quaternary volcanic province of the Snake River Plain, the Cascades and the Columbia Plateau. The volcanic cover extensively obscures the complex pre-Cenozoic geology in that area. The Jackson Mountains therefore provide a link between the southern province of the Cordillera and the scattered Paleozoic and Mesozoic exposures in eastern Oregon, as well as the resumption of the Cordilleran belt in northern Washington and British Columbia. The Cordillera changes dramatically along strike, and the geology in this area may help clarify some of those lateral relationships.

The Jackson Mountains also are positioned within the belt of accreted terranes, and have been proposed in the past to be a portion of an exotic terrane themselves with docking ages from the Early Triassic to Late Jurassic (e.g., Speed, 1977, 1988; Davis et. al., 1978; Hamilton, 1978; Coney et. al., 1980; Saleeby, 1983; Erskine and Moores, 1984). The geology in the range therefore serves as a test of the record of the accretionary and dispersionary terrane history along the Cordilleran margin from the Late Palezoic into the Mesozoic. An understanding of this area will help unravel the nature and timing of events along that margin.

A third aspect of the Jackson Mountains is that the range has never been deeply buried, extensively metamorphosed, undergone batholith intrusion, or been pervasively overprinted by deformational events, unlike the great majority of equivalent exposures in the southern United States Cordillera. It hence has a very well-preserved and complete record of many aspects of Cordilleran evolution. For this reason, the range serves to illuminate the nature of intra-arc evolution and the magmatic, structural and depositional processes that operate along plate margins in such complex and changeable tectonic environments.

Four Late Paleozoic and Mesozoic formations (see fig. 2) are present in the Jackson Mountains (the McGill Canyon and Happy Creek units are here proposed to have formation rank, and the Bliss Canyon Formation renamed from the Boulder Creek beds and also given formation rank). The Mississippian to Early Permian McGill Canyon Formation consists of basinal hemipelagic and turbidite sedimentary rocks overlain by andesite flows, related epiclastics, and shelfal limestone and terrigenous clastic lithologies. The Bliss Canyon Formation consists of Ladinian to middle Norian nonvolcanogenic rocks, with a carbonate bank and terrigenous littoral complex prograding eastward over a basinal province with hemipelagic argillite and terrigenous and calcarenite turbidites. The contact between the Bliss Canyon and McGill Canyon Formations may be a disconformity. The Bliss Canyon is conformably but abruptly overlain by a thick very shallow marine to subaerial andesitic volcanic pile, the late Norian to Bathonian Happy Creek Formation. The pile includes alluvial fan epiclastic rocks and volcanics ranging from basaltic andesite to dacite. Late in its history the volcanic pile was cut by a conjugate east-west high-angle oblique-slip fault system with fault-bounded basin formation. The basins were then filled and overlapped by the Callovian to Aptian King Lear Formation. This unit consists dominantly of molasse shed from two phases of thrusting, of late Middle to Late Jurassic and late Early Cretaceous age. Several silicic volcanic centers and numerous tuffs are contained in the King Lear stratigraphy. The thrusts occupy two distinct domains, west-vergent on the east flank of the range and east-directed on the west flank, with the high-angle fault domain in between. Both thrust domains were active in both

FIG.2 GEOLOGIC MAP OF THE JACKSON MOUNTAINS, NW NEVADA



orogenic episodes, and some of the high-angle faults were reactivated as tear surfaces. Stocks and sills of the Early Mesozoic Intrusive Suite are comagmatic with the volcanic facies in the Happy Creek and King Lear Formations and intrude the pre-Cenozoic rocks of the range. The pre-Cenozoic strata are unconformably overlapped by Tertiary basalt and rhyolite flows and Quaternary alluvial sediments. Basin and Range deformation is negligible, with minor eastward tilting of the range and reactivation of thrusts as moderate-angle normal faults.

The treatment of the geology of the Jackson Mountains will include chapters on: (i) McGill Canyon Formation; (ii) Bliss Canyon Formation; (iii) Happy Creek Formation; (iv) King Lear Formation; (v) Early Mesozoic Intrusive Suite; (vi) character and history of metamorphic episodes in the range; (vii) structural geology; (viii) a summary of the geology of the Jackson Mountains; (ix) tectonic interpretation of that geologic history; (x) regional correlation and tectonic evolution, from a viewpoint in the Jackson Mountains. Appendices and a reference section will follow, and a 1:50,000 scale geologic map and set of cross-sections accompanies the text.

Rock description and identification is based on the bedding thickness scale of Ingram (1954), Folk (1980) and Pettijohn et. al. (1972) for sandstone classification, Folk (1962) and Dunham (1962) for carbonates, Streickeisen (1973, 1979) for plutonic and volcanic lithologies, and the DNAG time scale has been used throughout.

CHAPTER TWO:

THE LATE PALEOZOIC OF THE JACKSON MOUNTAINS:

THE MCGILL CANYON FORMATION

The McGill Canyon Formation is upgraded in status here from the McGill Canyon Unit of Russell (1981, 1984). The strata of the unit were placed within the Happy Creek Volcanic Series when the area was originally discussed by Willden (1963, 1964). Most of the rocks assigned to this series by Willden have been separated in this work into the Happy Creek and Bliss Canyon Formations (which Russell called the Happy Creek Igneous Complex and the Boulder Creek Unit, respectively). The McGill Canyon Formation as defined here (and corresponding quite closely in area to the exposure mapped by Russell as the McGill Canyon Unit) covers approximately 17 km^2 in the southern Jackson Mountains, to the west and northwest of King Lear Peak in the area of McGill Canyon (locality 1 on fig. 3). Detailed field mapping has clarified the internal structure and hence the stratigraphic relationships within the exposure area since these earlier studies.

The stratigraphy is contained in two west-dipping thrust sheets, and is separated from strata of the Bliss Canyon Formation (upper sheet) and the Happy Creek Formation (lower slice) by a dextral tear fault. The tear itself is displaced slightly along the upper thrust. This upper thrust has been reactivated in Tertiary time with net normal displacement. The tear fault is a Middle Jurassic structure reactivated during Middle to Late Jurassic thrusting and offset during a second phase

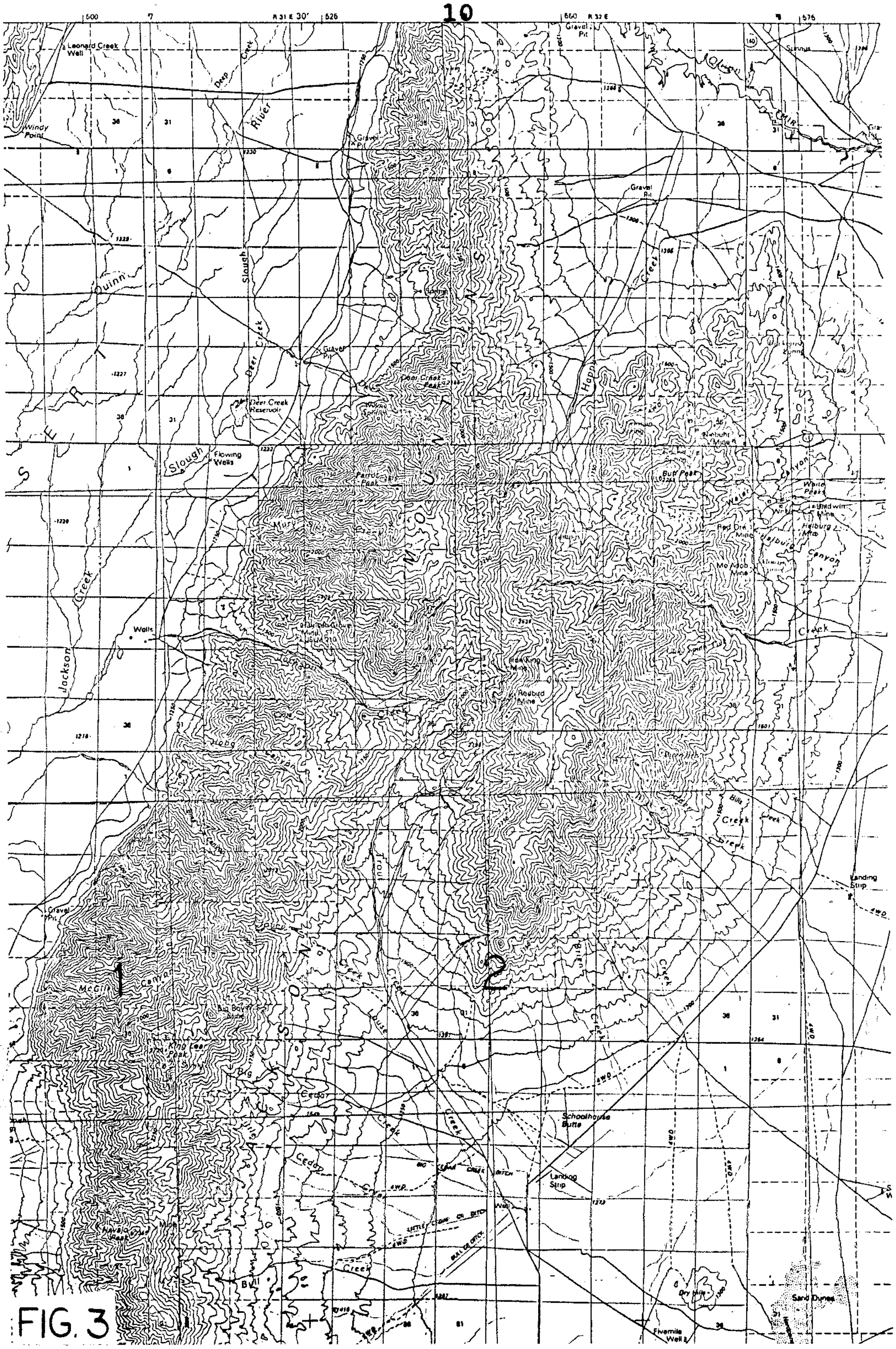


FIG. 3

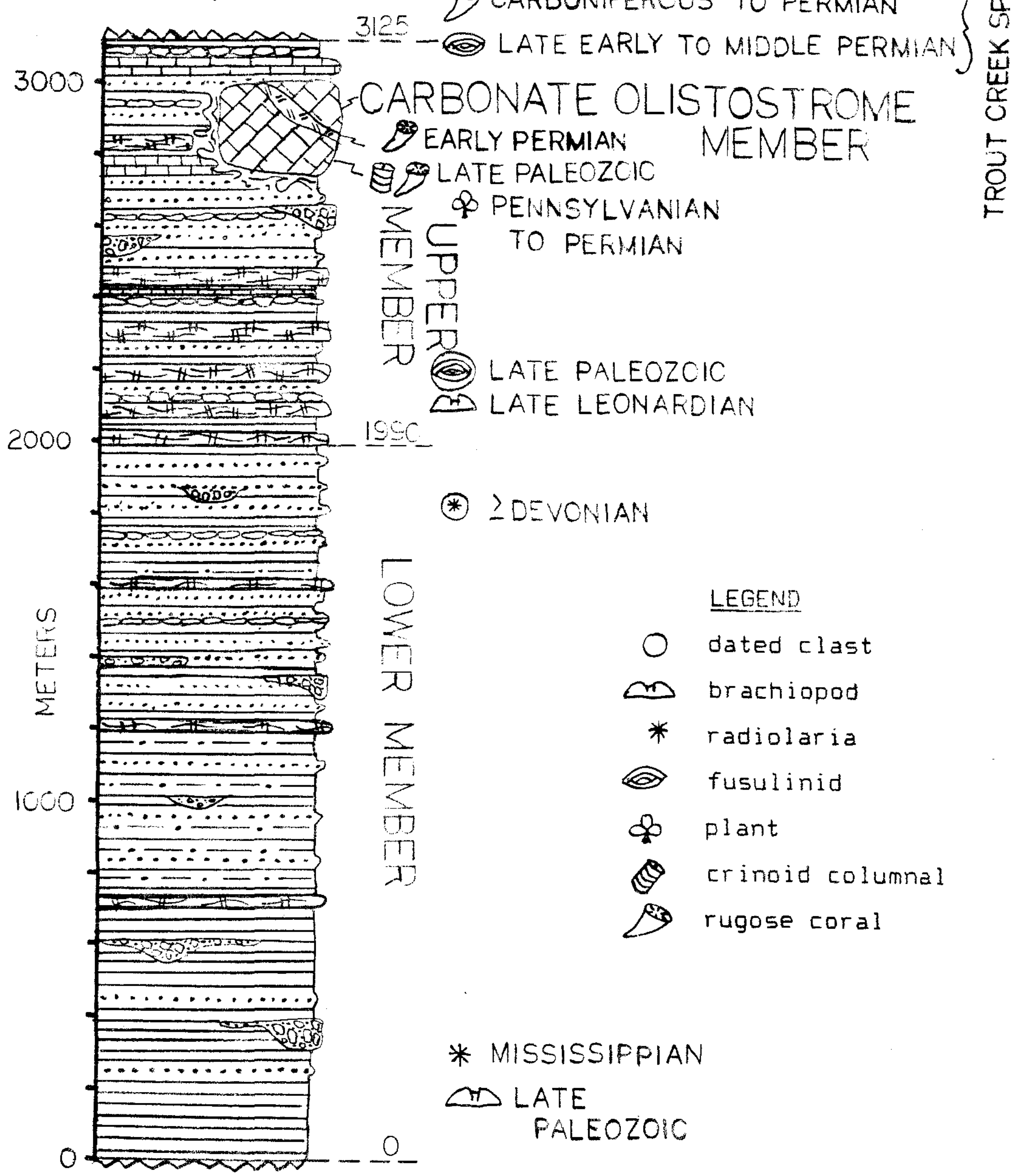
Map of the field area in the Jackson Mountains, NW Nevada, showing the locations of the two area of outcrop of the McGill Canyon Formation. 1 is the proposed type area around McGill Canyon, and 2 is the small area of exposure at the tip of Trout Creek Spur. The stratigraphic section is taken from area 1.

of middle Cretaceous thrusting. The result is that there are two fault-juxtaposed steeply east-dipping strucutral sections of the McGill Canyon Formation, both abruptly truncated on the north (see map and cross-sections). Both sections have been internally deformed on an outcrop scale, though coherent and conformable on a larger scale, and have been metamorphosed only to zeolite facies. There is an additional small sliver of inferred upper member McGill Canyon Formation in a stream gully at the very southern tip of the Trout Creek Spur in the eastern Jackson Mountains (locality 2 on fig. 3).

A composite stratigraphic sequence for the formation is reconstructed here, but due to the structural complications and patchy exposure, individual bed thicknesses and placements are only approximate. I formally propose here the McGill Canyon Formation as a lithostratigraphic unit (I.U.G.S. guidelines, 1976) with a stratigraphic type area in the area west and northwest of King Lear Peak, including McGill Canyon and the unnamed canyon between McGill and Alaska Canyons just to the north.

For purposes of mapping and discussion, the McGill Canyon Formation is here informally subdivided into three members (see the stratigraphic section and fence diagrams, fig. 4, 5): (1) the lower member - a sequence, at minimum 1990 meters thick, of green and grey argillite and bedded chert, lithic chertarenite, cherty/lithic conglomerate, and rare andesite flows; (2) the upper member - green and grey argillite and bedded chert, andesitic flows and related epiclastics, cherty/lithic arenite and conglomerate, and fossiliferous limestone in a sequence a minimum of 1100 meters thick; and (3) the olistostromal

FIG. 4



Stratigraphic column for the McGill Canyon Formation of the Jackson Mountains, NW Nevada. Thicknesses and placement of beds are approximate. Section is composite, and was taken from the proposed type area around McGill Canyon.

carbonate member - a suite of olistostromal blocks of fossiliferous limestone emplaced as sedimentary slides tens to hundreds of meters in size into and encased by the upper member.

In the stratigraphic discussion, the following topics will be discussed in order for each member: (1) stratigraphy, depositional features, petrography and sedimentology; (2) age data; and (3) the depositional environment and history. A discussion of the architecture and evolution of the formation as a whole will conclude the discussion. Regional tectonics and stratigraphic correlation will be discussed in chapter 10.

Lower Member

Stratigraphy The observed thickness of the lower member of the McGill Canyon Formation is approximately 1990 meters. No lower stratigraphic contact is exposed. The member consists primarily of olive green to dark grey, thinly bedded, finely laminated, and otherwise rather featureless argillite. Dark grey homogenous to thinly laminated chert beds a few cms thick are also interbedded with the argillite, although they are not common and are apparently restricted to the upper part of the member. Intercalated with the argillite and chert are thin, planar, and laterally continuous interbeds (10 to 80 cm thick) of dark greenish-grey, moderately sorted fine- to medium-grained sandstone. This arenite is cross-laminated to planar-laminated on a cm scale, normally graded and passes gradationally upwards to argillite. The beds have abrupt bases. This set of features is characteristic of the Bouma T_{cde} turbidite sequence. The bulk of the lower member belongs to facies C2 (organized sand-mud units) and E1 (disorganized muds) of Stow (1985).

Channel-fill lenses and thick laterally discontinuous beds up to several meters thick, containing chert-quartzite pebble to cobble conglomerates and breccia with a lithic wacke or lithic arenite matrix are present within the member (though greatly subordinate to the argillite and arenite encasing them). These coarse-grained facies exhibit scoured bases, clast- to matrix-support and bimodality with good sorting, and normal grading. Intraformational argillite rip-up intraclasts are a common feature, and the beds change and fine upwards from basal massive conglomerates, to planar laminated pebbly sandstone, to cross-laminated arenite, to finer-grained planar-laminated arenite, and then gradationally up to the argillite. Not all of the facies in this series are always present; some of the conglomeratic beds are composite (with truncation of the lower bed by the base of the upper), or alternatively the lower or upper portions may be missing. In the former case these beds are then transitional into the thin-bedded, graded arenites. Some of the conglomerates differ in nature and display none of the above features, but instead are diamictites that are more clast-deficient with very poorly sorting, and are massive and homogenous with mud-matrix support. The clasts in the conglomerates are rounded to well rounded and are composed of chert, quartzite, vein quartz and intrabasinal rip-ups. The chert component is made up of grey to green tuffaceous and/or radiolarian recrystallized chert, while the quartzite is a grey meta-quartz arenite and the vein quartz can be deformed and strained. The intrabasinal conglomerate lithologies include flaggy rip-ups of argillite, and limestone and andesite fragments - the former two lithologies identical to the bedded facies in the upper member and to the carbonate olistostrome member. The volcanic clasts are also very similar to the rare flows occurring in the lower member.

Petrographic examination shows that the sandstones contain detrital quartz, metamorphic rock fragments (phyllite to schist), and chert/quartzite grains in subequal amounts, with minor mica, feldspar and heavy minerals and with little clay matrix. The quartz is often deformed and polygonized with partial annealing and recrystallization. The metamorphic rock fragments and chert and quartzite grains are well-foliated with some white mica growth. The sandstones are moderately sorted and subangular lithic chertarenites (Pettijohn et al., 1973; Folk, 1980). Overly-close packing, suturing of the siliceous grains, and stylolitization in the arenites are characteristic, and locally several generations of slaty cleavage have developed in the argillitic facies of the McGill Canyon Formation in general. The medium-grade metamorphic episode evident in the sedimentary component is a sourceland feature predating their inclusion in the McGill Canyon Formation, as the deformational fabrics end at the grain boundaries, vary in the degree of development from grain to grain, and are of random orientations.

Rare plagioclase phyric green andesitic flows (with an unidentifiable mafic phase) are present at several levels in the section, and are thick and massive. Identical green dikes intrude the section (apparently cogenetic with the flows). The dikes and flows have undergone very low-grade (static?) metamorphism and recrystallization and have lost much of the original textures.

The lower member displays chaotic soft-sediment deformation with broken formation characteristics well-developed in some areas (particularly the more argillitic horizons); this is also associated (perhaps genetically) with the andesitic flows. There is no systematic nature evident in this deformation, and cleavage is only very locally developed in some of the most contorted strata.

Within the lower member, the conglomeratic and coarser-grained facies are restricted to an area between McGill Canyon and the tear fault to the north, and are most abundant just north of McGill Canyon. This lens of coarser-grained rocks lies in the upper part of the member, and thins and fines to the north and especially the south along strike. Even within the lens, the bulk of the member is still the thin-bedded argillite, chert and fine-grained arenite facies.

Both the intraformational clasts and the bedded facies within the member exhibit chlorite, white mica and quartz replacement and growth, and the feldspar (in the volcanic and clastic facies) has been extensively sericitized. The mafic mineral(s) in the andesites are completely replaced by chlorite. The limestone clasts in the member (there are no bedded carbonates in the section) have been recrystallized to a coarse-grained light grey spar, with depositional features preserved.

Age Some biostratigraphic data has been reported by Russell (1981) (the stratigraphic placement is mine): (1) A Late Paleozoic fauna, including the brachiopod Derbyia, from the lower part of the lower member, identified by B.R. Wardlaw, 1980. (2) Mississippian radiolaria in bedded siliceous argillite of the lower part of the lower member, identified by D.L. Jones and N.J. Silberling, 1980. This siliceous argillite was not reworked but was part of the section (D.L. Jones, pers. comm., 1988). (3) Devonian or older radiolaria in a detrital chert clast in the upper lower member, identified by D.L. Jones, 1980.

Depositional Environment The argillite was deposited in a basinal setting by hemipelagic processes (Stow and Piper, 1984); the chert similarly was laid down by distant volcanic eruptions and/or by radiolaria blooms. The lack of organic traces and remains, together with the color of the strata, implies a rather anoxic nature to the basin. Where lacking a turbidite component, these hemipelagic deposits were laid down in a deep basin plain to slope setting. The slope facies are distinguished by the presence of the tuffs, the paucity of distal fine-grained low-density turbidites, the presence of debris flows and interfingering with the upper fan coarse-grained turbidite facies (see below).

The conglomerates belong to the R_3 and S_1 through S_3 turbidite facies of Lowe (1982), while the finer-grained sandy portions correspond to his T_t turbidite facies (which includes the Bouma T_{cde} sequence). The R_3 facies is defined as normally-graded gravelly high-density suspension deposits. The S_1 to S_3 facies are sandier high-density traction to suspension deposits, characterized by normal grading, a substantial gravelly component, and a sequence of cross-bedded and channelled (S_1- traction sedimentation) to planar bedded (S_2 - traction carpet) to graded (S_3 - suspension) textures. This sequence passes up into the T_t facies, low-density sandy traction to suspension deposits, featuring the upper part of the traditional Bouma sequence: cross-laminated medium- to fine-grained arenite (traction deposits), becoming planar-laminated (suspension) and then argillitic (background hemipelagic sedimentation) upwards. The R_3 and S facies commonly have scoured bases, as is observed in the outcrop area; the intraformational rip-ups are also characteristic of more upcurrent turbidite facies.

The laterally continuous finer-grained arenite beds are distal low-density sandy turbidites. These facies and the hemipelagic argillite and chert (where interstratified with the arenites) were deposited in a lower turbidite fan setting upon the basin plain. These turbidites might have been generated by storm events in shallower environments (Einsele and Seilacher, 1982). The coarse-grained turbidite lenses and thick discontinuous beds are more diagnostic of channelled suprafan lobes within the mid-fan to an upper fan and slope setting (Walker, 1978). The soft-sediment slumps and the debris flows are more characteristic of the slope environment, while the organized turbiditic conglomerates are more typical of the distributary channels of the middle to upper fan setting (Shanmugam and Moiola, 1985). The coarser and more proximal facies, restricted to an area in the upper part of the member and the more northern part of the area, are inferred to be a submarine turbidite fan complex. This fan complex interdigitates with the lower fan facies (thin-bedded arenites and argillite), and they in turn with the basin plain deposits (argillite and chert). In its upper reaches, the fan is overlain by and interfingers with the slope facies. The more proximal portions of the fan contain all the coarse-grained turbidites, as well as the rare andesitic flows; the latter may have fed through the same feeder channel as the turbidites.

Upper Member

Stratigraphy The transition from the lower to the upper member is quite gradational and conformable, and was observed in both thrust slices. It is marked by the appearance of volumetrically important volcanic flows and related epiclastic facies in the section over an interval of tens of meters, by decreasing argillite and increasing arenite content, and by the presence of beds and slide blocks (the latter belonging to the olistostromal carbonate member) of crinoidal biosparrudite. The observed vertical extent of the upper member in the main outcrop area around McGill Canyon in the southwest Jackson Mountains is 1135 meters, and no upper stratigraphic contact is exposed; the upper contact is a thrust fault. In the Trout Creek Spur occurrence, it is possible that higher levels of the member are preserved, and that they are conformably to disconformably overlain by the base of the lower member of the Bliss Canyon Formation.

The andesitic volcanics are flows, fairly thick (with no pillow structures present) and olive green in color, with a fine-grained groundmass and hyalopilitic to pilotaxitic textures. Typically, large (several mm) plagioclase laths and pyroxene or amphibole phenocrysts are present, but the feldspar has been extensively replaced by fine-grained sericite, and the mafics entirely replaced by chlorite, calcite, Fe-oxides and epidote, though the original igneous textures are preserved. Quartz phenocrysts are present in some of the flows. The matrix to the flows, inferred to originally have been cryptocrystalline with microlites of plagioclase, Fe-oxides and mafic mineral, has been metamorphosed to a fine network of chlorite, sericite, calcite and quartz.

Green coarse-grained volcanic epiclastic arenites are also present in the upper member, and appear to be associated spatially with the volcanic flow facies. The arenites are dominantly arkosic (with subangular detrital plagioclase grains and subordinate volcanic rock fragments), coarse-grained but not conglomeratic, only moderately well-sorted, and have very little clay matrix. What matrix there is has been altered to chlorite. The plagioclase grains have largely been sericitized, but have some evidence of pre-depositional cataclastic deformation. The volcanic rock fragments have been replaced and altered in the same fashion as the surrounding volcanic flows in the section.

Other sandstone lithologies are also present in the section, with abundant chert/quartzite detrital grains and metamorphic rock fragments. These arenites are thin- to thick-bedded, are graded or massive and planar-laminated to cross-laminated, and are laterally continuous. They locally also contain intrabasinal crinoid and other fossil fragments, though otherwise they are very similar to the arenite facies in the lower member. Several distinct sources (volcanic and cherty/lithic) are seen in the sandstone petrography, with relatively little mixing in any one bed. A tree stem (Calamites, S. Ash, written communication, 1986) was found near the top of the upper member in a chertarenite matrix.

The green and grey argillite matrix of the upper member is also very similar to that of the lower member. The exceptions are extensive bioturbation in the upper part of the upper member, and the presence of some black intervals containing reduction spots. Grey thin-bedded and laminated chert is abundantly interstratified with the argillite in the upper member. In some cases this chert was silicified as an eogenetic (very early diagentic) hardground, with scouring of a partially lithified bed taking place during the deposition of the overlying bed. The chert is tuffaceous, with laminae of very fine-grained angular igneous mineral fragments in a cryptocrystalline silica matrix, and commonly with recrystallized radiolaria as well. Conglomerates and breccias similar to those in the lower member (but here dominated by volcanic clasts, with subordinate intrabasinal limestone and argillite rip-ups, and uncommon granitic cobbles) are also present. The matrix is an arkosic wacke or volcanic arkose. The conglomerates are usually present in large channel-fill deposits and thick (10 meters or more) laterally discontinuous beds and lenses within the predominantly argillite, chert and thin-bedded fine-grained arenite sequence. As in the lower member, the extrabasinal clasts in the conglomerates are well-rounded and ellipsoidal.

Also present within the upper part of the member are abundant thin- to thick-bedded (1 to 10 meters or more) beds of light grey biosparrudite (Folk, 1962), containing crinoid columnals, bryozoans, brachiopod spines, sponge spicules and coated carbonate grains in a spar cement. One distinctive feature observed is silica replacement and cementation in the voids. The detrital component in these calcarenites (a more general term for detritally-worked bioclastic sediments) is clean, moderately well- to well-sorted, abraded and broken with large-scale trough cross-bedding. The carbonate beds do not exhibit scouring, grading, or mixed provenance. The calcarenites thicken and are more common upwards in the member. A second carbonate facies (much rarer) has angular limestone intraclasts and bioclastic fragments floating in a wacke or arenite matrix.

The overall trend in the upper member is of coarsening-upwards, and increasing dominance by the volcanic epiclastic component, with the chert/quartzite lithic component diminishing in influence. Andesite flows are common in the lower two-thirds of the member, but not the top third, while calcarenites become more common upwards throughout the section. The conglomeratic channel-fill lenses and beds become larger and less numerically abundant upwards, and are restricted mostly (in both fault slices) to the area just north of McGill Canyon. The abundance of graded beds decreases upwards in the section, the arenite/argillite ratio increases, and bioturbation and lenticular bedding become more common.

The small outcrop located near the tip of Trout Creek Spur and placed in the upper member lies at the base of an east-dipping thrust fault (structurally overlying the lower Happy Creek Formation and related intrusives). The member here consists of grey argillite, green volcanic arenite, conglomerate, and grey to brown recrystallized limestone beds. The conglomerate contains chert, quartzite and limestone clasts and intraformational argillite rip-ups, and is very similar to the conglomeratic facies seen in both members of the McGill Canyon Formation in locality 1. The strata are somewhat deformed and metamorphosed. The nature of the contact with the overlying Bliss Canyon Formation (the lower member) is not exposed, but is inferred to be conformable to disconformable. The strata are distinctly more like the upper member of the McGill Canyon Formation in lithology and age (see next paragraph) than any other unit in the range (or even the area), and are therefore correlated to the uppermost part of the upper member.

Age The Calamites tree stem from the upper part of the upper member is Pennsylvanian to Permian (S. Ash, written communication, 1986). The megacrinoids are Late Paleozoic. Several biostratigraphic assignments are reported by Russell (1981) (once again, the stratigraphic placements are mine): (1) Late Paleozoic fusulinids of uncertain paleogeographic affinity in a limestone clast in a polymict breccia in the lower upper member, identified by N.J. Silberling (1979). (2) Late Leonardian (late Early Permian) Spiriferella drashi, a brachiopod in volcanic epiclastics of the lower part of the upper member, identified by B.R. Wardlaw (1979).

In addition, Willden (1964) also mentions a collection of corals and fusulinids from the Trout Creek Spur area. The corals are clisiophyllid forms, of Carboniferous to Permian age, identified by H. Duncan, 1956. The fusulinids were identified by L.G. Henbest in 1957 as species of Parafusulina, Pseudofusulina, and Schwagerina, possibly of late early to middle Permian age (and of the McCloud faunal province). The disconformably(?) superjacent lower member of the Bliss Canyon Formation in this area has Middle Triassic fossils several hundred meters higher in the section.

Depositional Environment

The presence of abundant sedimentary slide blocks within the member indicates a slope to shelf setting; the blocks could not travel any great distances out onto the basin plain. The bedded limestones are interpreted here as tempestites (Einsele and Seilacher, 1982). These beds are storm-related beds and formed during unusually energetic events, with carbonate biogenic debris (from either offshore bars or a platform or from carbonate-dominated littoral environments) remobilized and laid down further out on the shelf. The carbonate grains in these tempestites have a very shallow normal marine origin. Skeletal material breakage, sorting, cross-bedding and abrasion indicate high-energy sedimentary processing. The fauna are also normal shallow marine. These beds do not have the characteristics of turbidites; their lack of graded bedding or exotic clasts (though these are not necessarily definitive), the bed thickness, and the extent of trough-style cross-bedding within the beds leads to their interpretation as submarine mega-ripples and dunes migrating under the impetus of occasional and very high-energy storm events. A single thicker bed might be composite, representing the accumulative effect of several storms. These deposits, though below normal wave-base, are restricted to shelf settings above storm wave-base. Carbonate diamictite and polymictite debris flows are present (though uncommon) in the section, and are debris flows caused by mass instability.

The conglomeratic beds and lenses are interpreted to be high-density and gravelly to sandy turbidite deposits, similar in origin to those of the lower member. Upwards in the upper member, however, the trend is to thicker but more laterally-restricted and less common conglomeratic deposits, to greater restriction to an area just north of McGill Canyon, and to the disappearance of the more distal and laterally continuous, low-density sandy turbidites. This implies an environment in the upper fan and then in the feeder channel, within the slope and shelf settings respectively (Walker, 1978). Outside of the fan complex, the upper member is dominated by slope and shelf facies. The sandy low-density turbidites disappear up-section, and the sandstone beds become thicker, more abundant and shallower in origin (displaying ripple cross-lamination and bioturbation). These sandstones are probably also storm deposits, below normal wave base and situated in the distal shelf location, but not representing such extraordinary high-energy events as the carbonates. Argillite and chert still make up an important contribution to the section, but are less abundant. Much of the argillite is silty. The argillite is also siltier, indicating closer proximity to the source.

The andesitic beds in the lower part of the member are interpreted as submarine, fairly viscous volcanic flows. The epiclastic facies, though distinct in provenance, share the turbiditic features and interpretation of the chert/quartzite lithic arenites. A cessation of proximal volcanism is indicated in the upper part of the member by the disappearance of the flow facies. The tuffaceous cherts continue in the upper part, so distal volcanism was still taking place.

Olistostromal Carbonate Member

Only one occurrence of this member is large enough to be mapped separately from the upper member, but a number of other, smaller slide blocks of the same lithology exist. These blocks are all encased in localized broken formation within the upper member. The large block in particular has highly contorted and deformed argillite and chert strata of the upper member (though with no exotic, extraformational lithologies involved - an indication of a non-tectonic origin). The deformation takes place around the bottom and sides, but not the top (implying emplacement of the blocks during sedimentation of the upper member, and later overlapping). The blocks range in size from 10 meters or less to the largest one, 300 by 400 meters (in map view). Internal bedding can be at a high angle to that in the surrounding matrix. All of these features serve to distinguish them from the bedded carbonates of the upper member.

Stratigraphy The suite of carbonate slide blocks are identical in composition and texture to the limestone beds within the upper member, and hence are correlated to them as lateral equivalents. Lithologically the blocks are biosparrudites (Folk, 1962) - large skeletal fragments and whole organisms in a coarse sparry cement (or less commonly a carbonate mud matrix - biomicrites). Texturally they are fossiliferous grainstones (Dunham, 1962), and exhibit clast-support, no matrix, large rounded biochems, good sorting, and have broken and abraded grains. The biological component includes crinoids, gastropods, solitary rugose corals and bryozoa, and many of the grains are coated (some are actually oncolitic). The largest block also contains plagioclase + hornblende andesite porphyry dikes, which cut across bedding and also are very similar to those in the upper member (including the type and extent of replacement and alteration). As the dikes are truncated by the edges of the block, they are interpreted as having been intruded in situ before the blocks were emplaced in their present stratigraphic position.

The carbonate beds at the tip of Trout Creek Spur have intertidal features (the flat-pebble conglomerate), and as they contain mafic dikes not seen in the rest of these beds, they may actually be blocks of the olistostromal carbonate member. The alternative is that they are the logical culmination of the shoaling trend in the formation, and are in place.

Age A solitary rugose coral from the largest block has been identified by C. Stevens (pers. comm., 1986) as Lytvolasma sp. - of Early Permian age and Uralian affinity. Russell (1981) also reports Late Paleozoic crinoid columnals and solitary rugose corals in this block (identified by N.J. Silberling, 1979).

Depositional Environment The slide blocks are diagnostically shallow marine facies. The presence of coated grains and oncolites (a feature caused by photosynthetic algae - Bathurst, 1971; Scoffin, 1987) is indicative of a very shallow, moderate-energy warm environment with a very low clastic input. The nature of the fossil component, and the detrital characteristics of the biochem grains also support such a setting. I infer a very shallow but subtidal setting on a carbonate marine platform or barrier reef for these facies (reef here does not imply an organic boundstone, but only a shallow marine topographic mound formed by the accumulation of biogenically-produced carbonate debris). Intertidal to supratidal indicators are missing (flat-pebble conglomerates, stromatolites, etc...), except at the Trout Creek Spur locality.

The thickness implies a relatively long-lived carbonate province. The carbonate beds within the upper member are composed of bioclastic sediment derived from this province, so the geographic separation of the two members (which are inferred to be lateral equivalents) could not be too great. The presence of andesitic volcanic facies within both members is another tie; both members were situated within the fringes of active volcanism.

The sliding and emplacement of the olistostromal member took place during sedimentation of the upper member, and after lithification of the olistostromal member. This latter event is one, however, that occurs very rapidly in carbonate settings such as the one proposed (Bathurst, 1971, Scoffin, 1987). If andesitic volcanism shut off in the two areas at the same time, the olistostromal member was originally deposited synchronously with the lower part of the upper member, and was probably resedimented very soon after; this is compatible with fossil data.

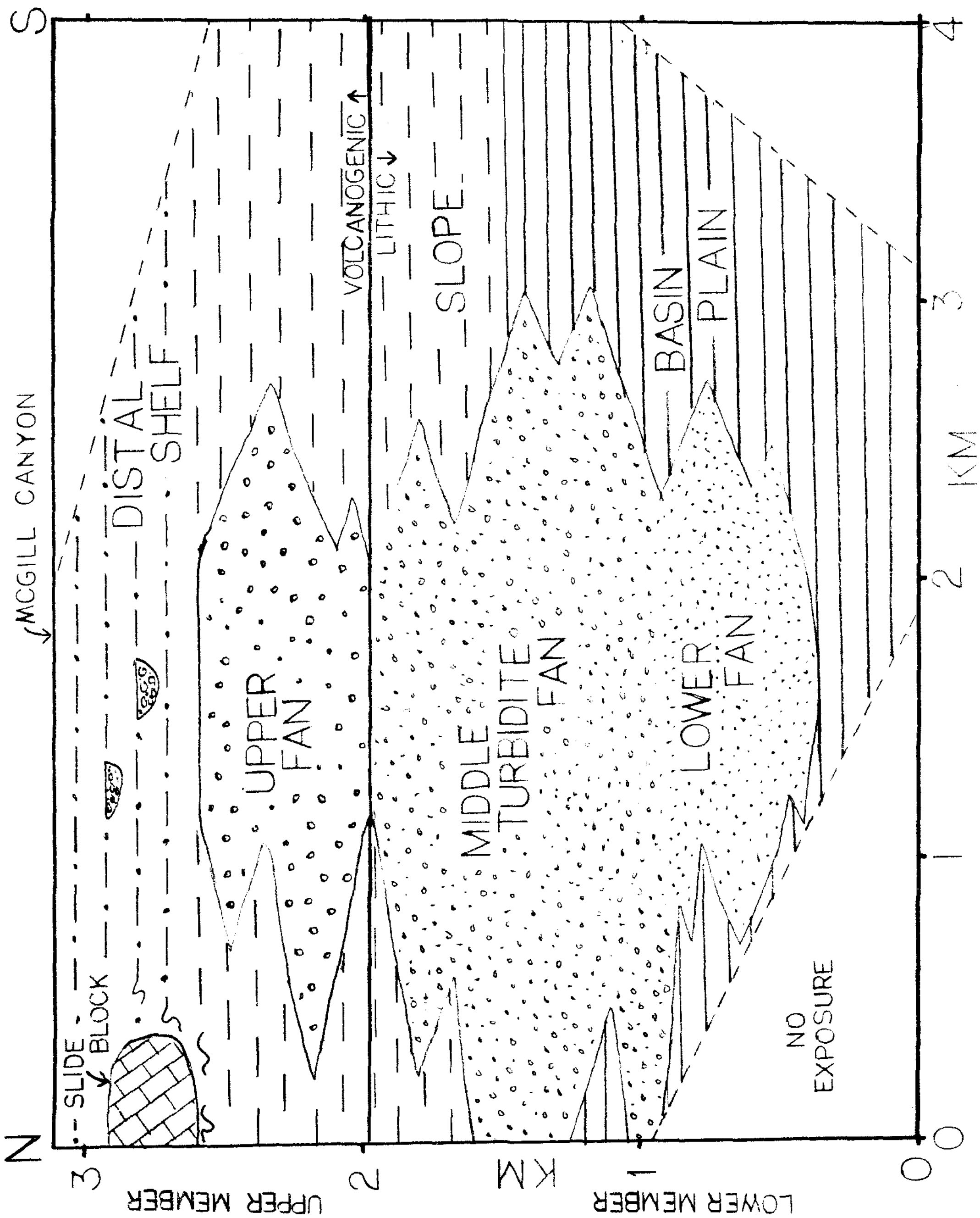
Stratigraphic Framework for the McGill Canyon Formation

The biostratigraphic data indicates that the lower member extends as far back as the Mississippian in age. The rest of the lower member is Mississippian to Lower Permian, extending perhaps into the early Leonardian. The lower part of the upper member is of at least Late Leonardian age, and the member extends up into the late early to middle Permian at the top. No stratigraphic evidence of a hiatus in sedimentation was observed. The carbonate slide blocks are also of Early Permian age, and probably correlate to the lower part of the upper member.

The lower member displays predominantly a single sourceland in

FIG. 5

Fence diagram showing facies for the McGill Canyon Formation of the Jackson Mountains, from the proposed type area around McGill Canyon, west of King Lear Peak.



its petrography. The sourceland was composed of a chert- and quartzite-bearing sedimentary basin assemblage, deformed and metamorphosed to medium grade (textures are phyllitic to schistose). This source area included rocks as old as Devonian or older, as evidenced from dated clasts. Some deeper unroofing in this source area is indicated in the upper member, as granitic cobbles also appear. The provenance in the upper member clearly indicates multiple source areas: a metasedimentary one (the same as already mentioned for the lower member, and also seen in the Bliss Canyon Formation), an intrabasinal shallow marine carbonate source, and an active andesitic volcanic complex. These latter two sources were much less evident in the lower member (though clasts of these affinities are present in the conglomeratic turbidites, they are not a significant component in the sandstone petrography). The source for the cherty/lithic debris and for some of the volcanic epiclastics was subaerial as well, as the extrabasinal clasts are well-rounded and display signs of fluvial or littoral working, and the deposits contain tree fossils. The paleogeography of neccessity must have been fairly complex, in order for these various geologic provinces to have been juxtaposed as evidenced in the petrography.

A prograding large submarine turbidite fan complex and feeder channels can thus be delineated within the lower and upper members, which interfingers with the basin plain to slope to shelf facies. The overall character of the McGill Canyon Formation is regressive, shoaling from basin plain and lower turbidite fan to middle fan to upper fan and slope to turbidite feeder channel and distal shelf environments. The shoaling trend continues through the younger Bliss Canyon, Happy Creek and King

Lear Formations. More proximal carbonate platform or barrier reef environments are also represented in the olistostromal carbonate member, before their down-slope voyage. The upwards thickening and increasing occurrence of the carbonate beds in the upper member also support shoaling. The overall scenario is one of sedimentation rates greater than subsidence (Einsele and Seilacher, 1982).

The fan was apparently quite long-lived, persisting from the Mississippian to the Early Permian, despite its small preserved size-only about 3 km across in outcrop. This implies a fairly stable setting. Though it is impossible to be sure without knowledge of the paleobathymetry, the fan appears to have been of the radial (sandy) type (Walker, 1978; Stow, 1985). Such fans are characterized by a sandier nature, smaller size, and true fan shape. They also have feeder channels, slumps, slides, and debrites associated with the slope and upper fan areas. In these fans the coarse-grained turbidites are restricted to the suprafan lobes and channels, while fine-grained turbidites are more widely interstratified with the hemipelagic facies. All of these features (except the radial morphology) are observed in the submarine fan complex in the McGill Canyon Formation.

The basin itself was restricted and anoxic, as signs of bioturbation and organic activity do not appear until local regression and offlap established the shallower facies within the upper member. The basin plain sediments are olive green to black and reduced. Most of the fossil material even in the upper member is transported (as in the case of the calcarenites), except for the radiolaria in the cherts, which are pelagic.

The sedimentation rate for the entire formation is about .03 to .05 m/1000 yrs (calculated as extending over about 80 m.y., from the Mississippian to the middle Permian). These rates are quite typical of arc basins, but are high for cratonic basins (epeiric seas) and passive margins and are low for foreland basins, trenches or successor basins (Schwab, 1976). The basin was normally- to over-supplied, judging by the sedimentation rates, the fan progradation, and overall shaoling and offlap (Brown and Fisher, 1977). Over-supplied basins are normally small, partly confined and located in tectonically active areas.

CHAPTER THREE:

MIDDLE TO UPPER TRIASSIC STRATIGRAPHY OF THE JACKSON MOUNTAINS:

THE BLISS CANYON FORMATION

The Bliss Canyon Formation as defined here comprises the nonvolcanogenic Middle to Upper Triassic sedimentary rocks of the Jackson Mountains. The unit includes the nonvolcanogenic portions of the various sub-units of the Boulder Creek Beds and units T_Rx, T_Ry, W, and R of Russell (1981; 1984), and the nonvolcanogenic Triassic portions of the undivided Permian-Triassic and the unnamed Triassic units of Willden (1963). The volcanogenic portions are placed in either the Happy Creek or McGill Canyon Formations. The Boulder Creek Beds were originally defined by Russell (1981) in the Boulder Creek area of the southeast Trout Creek Spur of the Jackson Mountains, but due to unrecognized structural complications Russell included portions of the younger Happy Creek and older McGill Canyon Formations. This and the fact that the best exposures of this unit in the Jackson Mountains are in Bliss and Alaska Canyons on the west side of the range have prompted the redefinition of this unit and a proposed change in name to the Bliss Canyon Formation. I formally propose here the Bliss Canyon Formation as a lithostratigraphic unit with a stratotype in the area south of Alaska Canyon (I.U.G.S. guidelines, 1976); the unit was <u>not</u> named the Alaska Canyon Formation because of the potential for geographic confusion!

Outcrop of the Bliss Canyon Formation occurs in two strips on the eastern and on the western flanks of the range with a total area of about

50 km^2 (fig. 6). The unit is involved in two thrust systems (east-vergent on the west and west-vergent on the east). All sequences of the Bliss Canyon Formation have been faulted and folded by these orogenic episodes, and in some cases by an earlier episode of east-west high-angle faulting and by later Tertiary Basin-Range normal faulting. No complete, undeformed section exists. Contact metamorphism is seen near the latest Triassic to Early Cretaceous Happy Creek and King Lear hypabyssal volcanic complexes and the related Early Mesozoic intrusive bodies, and regional very low-grade metamorphism exists throughout both belts of exposure. The result of these complications is that reconstruction of a complete and representative section of the Bliss Canyon Formation for a given area was not always possible. Throughout the following discussion an attempt has been made to remove the effects of subsequent deformation so as to discuss the original stratigraphic relationships, lithologies and depositional facies within the Bliss Canyon Formation. Thicknesses and placement of facies within the discussed intervals are rather approximate, as all sections described here are structurally restored, and lateral variation within the unit is significant.

For purposes of mapping and discussion the Bliss Canyon Formation has been divided into three informal time-transgressive members, with a maximum restored thickness on the order of one km (fig. 14): (1) a lower, dominantly argillite sequence with interbedded cherty and quartzose arenite, calcarenite, and chert-pebble conglomerate; (2) a middle massive limestone interval; and (3) an upper member of quartzose and cherty conglomerate, arenite, and argillite, with a minor volcanogenic component appearing at the very top, just before the

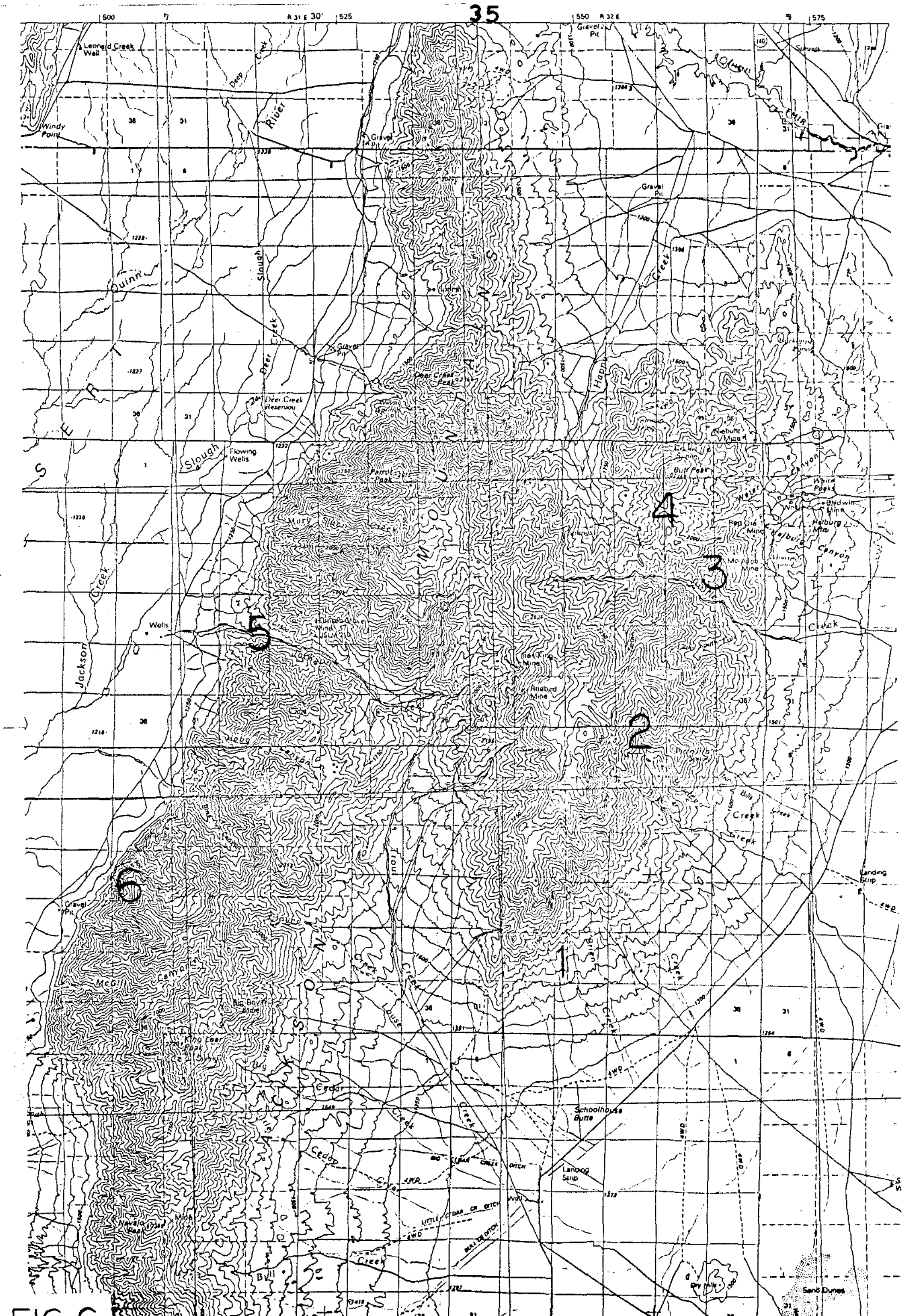


FIG.6

Map of the Jackson Mountains with section outcrop areas: (1) southern Trout Creek Spur; (2) Boulder Creek area; (3) region north of Bottle Creek; (4) area south of Buff Peak; (5) the mouth of Jackson Creek; (6) Alaska and Bliss Canyons (the stratotype).

volcanic flows, flows breccias and sandstones of the overlying Happy Creek Formation. All three members are seen in the western outcrop belt, while the eastern outcrop belt consists entirely of the first member. In all areas there is a gradational and conformable transition into the Happy Creek Formation. The lower depositional contact of the Bliss Canyon Formation is not seen in the Jackson Mountains, with the possible exception of an area at the very tip of Trout Creek Spur. Here, field relations indicate that grey argillite of the lower member of the Bliss Canyon may disconformably overlie the upper part of the upper member of the McGill Canyon Formation. Everywhere else there is structural truncation at the base of the Bliss Canyon Formation.

The Bliss Canyon Formation will be treated by geographic location. The unit has been divided up for convenience into the several outcrop areas. From south to north on the east side of the range, the areas to be discussed are: (1) the southern tip of Trout Creek Spur; (2) Boulder Creek to south Bottle Creek, from the southwest Trout Creek Spur; (3) the north Bottle Creek area; and (4) the area just south of Buff Peak. From north to south on the west side of the range, there is: (5) the region about the mouth of Jackson Creek; and (6) the area around Alaska and Bliss Canyons.

The discussion of the Bliss Canyon Formation will involve: (1) description of field and petrographic observations of lithology, sedimentary structures and relationships, provenance, and paleocurrent data, if any, for each geographic locality; (2) age and biogeographic constraints for each locality; (3) interpretation of the important parameters of the sedimentary facies of the interval for that area.

Finally, inferred regional depositional environments and paleogeographic conditions for the Bliss Canyon Formation as a whole will be discussed. Regional comparison, correlation and tectonic evolution will be treated later (chapter 10).

(1) Trout Creek Spur

This is the only area in the Jacksons where the depositional contact between the Bliss Canyon Formation and the underlying McGill Canyon Formation may be preserved. The gradational contact with the overlying Happy Creek Formation is also seen in this area; both contacts were mapped within a single thrust sheet. See fig. 7 and 8 for the stratigraphy.

Stratigraphy The Bliss Canyon Formation in this area consists of (approximately) 850 m of section; the exact thickness cannot be known more accurately without better knowledge of the internal structure of the interval. The lithology is predominantly thin-bedded to laminated grey argillite. A bed of cross-stratified dolomitic limestone (a lower intraclastic breccia grading up into a grainstone) was noted from the middle portion. At the base of the interval there is grey argillite and chert, inferred to be disconformably overlying strata placed within the McGill Canyon Formation. The carbonate beds in these strata were earlier considered exotic blocks within the Boulder Creek beds (Russell, 1981; 1984), and have Late Permian fossils (Willden, 1964). This stratigraphic assignment seems unlikely to me as the overall lithologic assemblage is much more characteristic of the upper member of the McGill Canyon

FIG. 7

Composite stratigraphic sections for the Bliss Canyon Formation. See text for detailed treatment. Thicknesses and bed placement are approximate. For the outcrop area positions, see map. Stratigraphic units include: (i) lower member of Bliss Canyon Formation; (ii) middle member; (iii) upper member; (iv) the superjacent Happy Creek Formation; and (v) the subjacent (inferred) McGill Canyon Formation.

lithologic symbols

- argillite
- chert
- siltstone
- thin-bedded sandstone
- cross-bedded sandstone
- conglomerate
- volcanic breccia
- carbonate breccia
- thin-bedded limestone
- thick-bedded limestone
- augite-phyric flow
- plag-phyric flow
- volcanic greenstone

biostratigraphic data

- ammonite
- pelagic bivalve
- coral
- conodont
- radiolaria
- belemnite

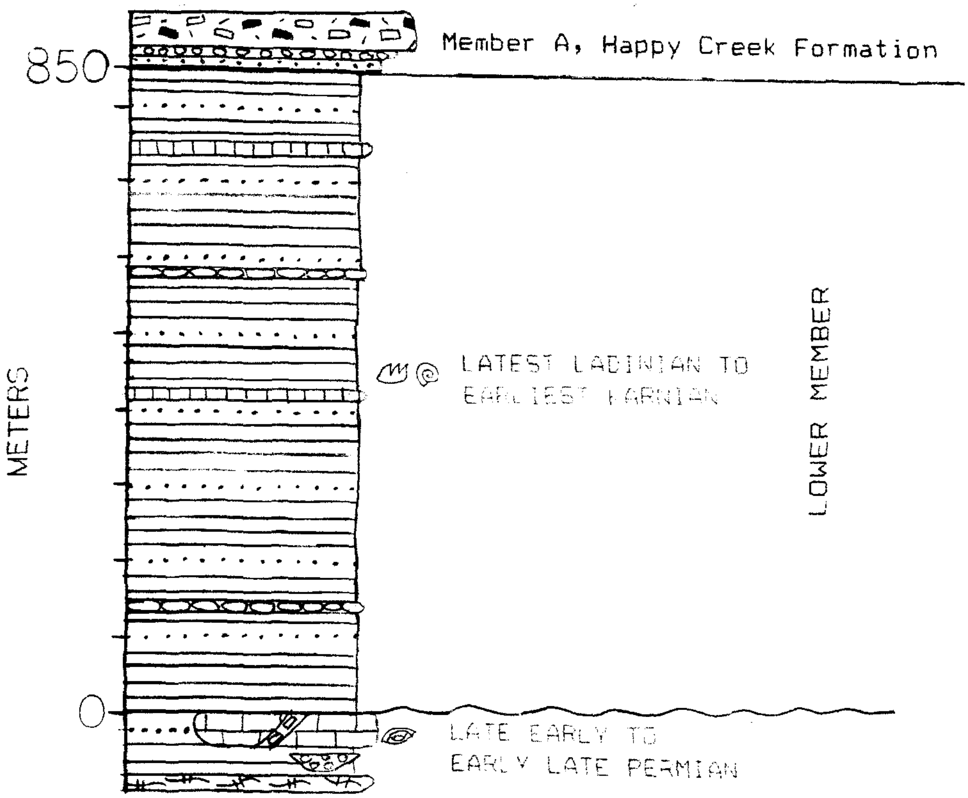
contacts

- conformable
- disconformable
- fault-bounded

FIG. 8

(1) Trout Creek Spur

Stratigraphy of the Bliss Canyon Formation, Jackson Mountains



Upper Member, McGill Canyon Formation

Formation (see earlier discussion) or of the uppermost Bliss Canyon Formation than of the base of the Bliss Canyon. As such large limestone blocks are not found elsewhere in the Bliss Canyon Formation, and also due to the stratigraphic and structural position, the McGill Canyon Formation is a much more likely candidate for this part of the sequence.

At the upper contact, the grey argillite of the Bliss Canyon grades into green volcanic arenite, then coarsens into volcanogenic conglomerate and breccia, and then is overlain by green andesite flows. The contact with the overlying Happy Creek Formation is put between the argillite and that green arenite and is remarkably consistent and mappable over the entire range.

Russell (1981) reported quartz- and chert-bearing fine-grained arenites as thin interbeds within the argillite, the latter also described by him as a calcareous chlorite-illite mudstone; feldspar was cited as only a minor constituent of the coarser lithologies. These sandstones are chert-arenite or sub-chert-arenite (Folk, 1980). Thin-bedded calcarenites and chert, soft-sediment deformational features (flame structures and load casts), and sometimes convolute cross-laminated to planar-laminated arenites grading up into argillite (Bouma T_{CDE} and T_{DE} sequences) were also described. I have assigned this entire interval to the lower member.

<u>Age</u> Russell (1981) reported latest Ladinian or earliest Karnian conodonts (identified by B.R. Wardlaw, 1980), and a latest Ladinian or earliest Karnian ammonite (identified by N.J. Silberling, 1980), both collected by Silberling from limestone-bearing highly deformed thin-

bedded siliceous mudstone. These samples came from the middle of this interval.

Depositional Environment The Trout Creek Spur section documents dominantly hemipelagic suspension-sedimentation (the thin-bedded grey argillite and chert), with low-energy base-absent fine-grained turbidite events (the thin-bedded, fine-grained grey lithic arenite beds with Bouma T_{CDE} and T_{DE} sequences) punctuating the record. These latter beds have a quartzose and cherty provenance, which indicates (though no point counts have been performed) that the sourceland was likely the cratonic or recycled orogen provinces of Dickinson and Suczek (1979) and Dickinson (1984), depending on the F to L_t ratio. Deposition was in a lower submarine fan setting (Walker, 1978). The calcarenite facies is too coarse and cross-stratified and has too abrupt a top to represent a classical proximal Bouma sequence (e.g., the C facies, with T_{A-E}), and seems to belong to the B2 facies (organized massive sandstone) instead (Walker and Mutti, 1973; Stow, 1984), more specifically to the S_1 sedimentary division of Lowe (1982). This would imply deposition by traction-sedimentation from a high-density sandy turbidite for the calcarenites, rather than from traction- passing upwards to suspension-sedimentation within a low-density muddy turbidite as would be appropriate for the $T_{(C)DE}$ arenite beds (Lowe, 1982; Stow and Piper, 1984). An additional conclusion that can be drawn from the modes of deposition as well as the lithology and provenance is that the source for the calcarenites was distinct from that for the lithic arenites, and probably closer as well (such high-density flows do not travel as far, and the S_1 sedimentary division is more proximal in terms of down-current evolution than the T divisions - Lowe, 1982). The dark grey color, preserved fine lamination and general lack of bioturbation indicate the environment may have had restricted circulation and anoxic bottom conditions.

(2) Boulder Creek

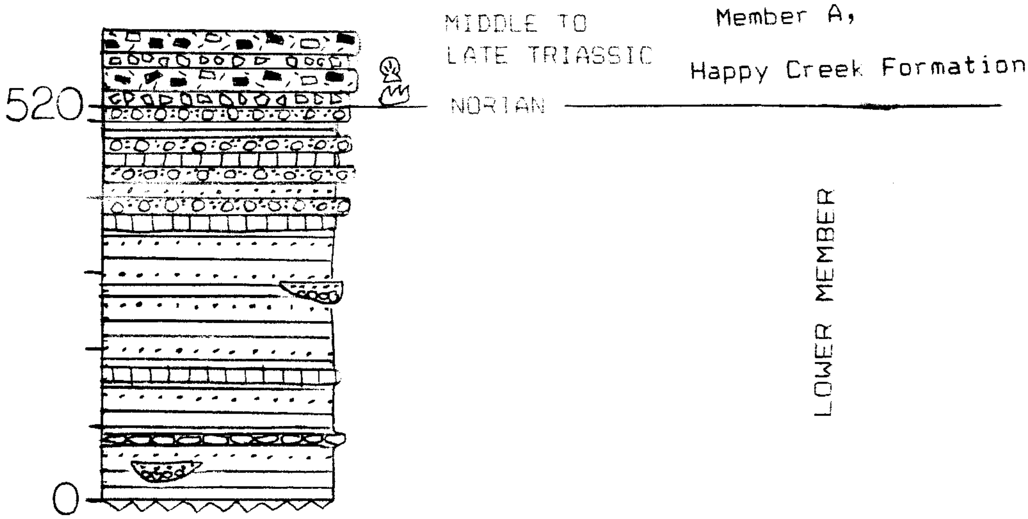
The eastern flank of the range contains several thrust slices of Bliss Canyon to Happy Creek Formation lithologies. South of Bottle Creek and in Boulder Creek to the vicinity of Burro Bills Spring there is about 520 m of section of Bliss Canyon below the basal Happy Creek, depending on correlation from one thrust sheet to another. No base was seen. See fig. 9 for the stratigraphy.

Stratigraphy The lower part of the interval consists mostly of grey thin-bedded to laminated argillite with thinly intercalated planar-bedded grey fine- to very fine-grained sandstone. The proportion of coarser clastics vs. argillite appears to increase somewhat upsection, and some of the sandstones begin to weather brown. In addition, there are rare small channel-fill lenses and thin to thick beds of green chert-pebble conglomerate; these definitely become more abundant up-section. In one area near the headwaters of Boulder Creek this is the dominant lithology at the top of the interval, with interbeds of argillite, sandstone and calcarenite. In other areas the thin-bedded argillite and fine-grained arenite facies predominate at the top. The conglomerates are characterized by large argillite and sandstone rip-up slabs (of the same

FIG.9

(2) Boulder Creek

Stratigraphy of the Bliss Canyon Formation, Jackson Mountains



lithology as the bedded strata), rounded chert cobbles and pebbles, and angular micritic to sparry limestone blocks, all in matrix-support in a sandy matrix. The clasts are commonly imbricated and normally graded; no reverse grading was seen. Typically they possess abrupt and even scoured bases and fine upwards to grey very fine-grained arenite and argillite. They are laterally continuous on a tens-of-meters scale. At the very top of the Bliss Canyon section in this area, angular clasts of plagioclase phyric volcanic rocks begin to appear in the conglomerates, and the arenitic matrix also becomes moderately feldspathic, recording the sudden and increasing importance of a volcanic epiclastic component in the provenance. The chert clasts in the conglomerates are tuffaceous (full of microlites of igneous minerals in a depositionally laminated cryptocrystalline matrix) and radiolarian-bearing (rather recrystallized but distinct), and are identical to both the clasts and the chert beds seen in the McGill Canyon Formation in the southwest Jackson Mountains. These chert pebble conglomerates have undergone suturing along the grain boundaries, and can exhibit overly-close packing. Further up in the section, vesicular augite-phyric flows and associated flow breccias and proximal volcaniclastics are interbedded with the sediments and rapidly (over several tens of meters) predominate. The Happy Creek Formation lower contact is put at the base of the first such unit; this contact is exposed several times in different imbricate thrust sheets and is constant in nature. Thin-bedded to laminated calcarenitic beds - thin, grey well-sorted fossiliferous grainstone to rudstone, usually recrystallized - are fairly abundant throughout the section. Russell (1981) also mentions black, grainy chert interbedded with the argillite and thin bedded arenite of this region; these beds may simply be more silicified and resistant argillite beds, or could be radiolarian-bearing or tuffaceous lithologies. This whole interval has been assigned to the lower member.

Age Bivalves collected from the headwaters of Boulder Creek in 1984 were identified by N.J. Silberling (pers. comm., 1986) as the pelagic clams Halobia or Daonella, of Middle or Late Triassic age. In addition, Russell (1981) reported Norian conodonts (Neogondolella navicula and Epigondolella mutlidentata) from this area that were identified by B.R. Wardlaw and collected by N.J. Silberling (both in 1979). Both of these samples were from carbonate clasts in the volcanogenic breccias of the lowermost Happy Creek, implying that the carbonate sourceland was of Norian age and that the boundary between the two formations in this area is Norian (and probably not significantly later, as the carbonate debris was apparently being shed off an active carbonate bank, not reworked-see next section).

Sedimentary Facies Using the classification of turbidite facies of Mutti and Ricci Lucchi (1972), as modified and explained by Walker and Mutti (1973) and Stow (1985), the A2 (organized conglomerate - the pebble- and cobble- conglomerates with a sandy matrix), D (classical medium-grained turbidites with base-cut-out Bouma sequences - the arenitic facies), B2 (organized sandstone - the calcarenites) and G (laminated argillite and fine arenites) are all present in the Bliss Canyon Formation in the Boulder Creek area. The graded and stratified coarse, proximal turbidite conglomerates of the A2 facies, with no inverse grading and with stratification and imbrication (the R_3, S_1 and S_3 sedimentary divisions; the S_2 was not seen), were probably deposited from an evolving gravelly (R_3 suspension-sedimentation) to sandy (S_1 traction sedimentation) high-density turbidite current on a low slope (Lowe, 1982) in the terraced, lower portion of a feeder channel to a submarine fan (Walker, 1978). A terrace origin seems more likely than one within the thalweg because of the relative lateral continuity and thinness of the conglomerate beds.

These conglomerates grade upwards into finer-grained sediment deposited from the residual high-density current by suspension-sedimentation (the S_3 division). The presence of the D and G facies and the overall thickening and coarsening upwards of the Bliss Canyon Formation indicate basin plain to outer fan depositional environments and slope progradation. Taken with the A2 lower channel terrace facies and the abundance of soft-sediment deformation, these strata (in the vicinity of the headwaters of Boulder Creek) are more probably levee deposit suspension-sedimentation turbidites (which can be very similar) near the feeder channel within the upper fan to basin plain and/or slope facies (Walker, 1978). Soft-sediment deformation (noted throughout the eastern side of the range) seems to positively correlate in the literature with slope steepness and with turbidite events as causative mechanisms. The rest of the area would more likely have been deposited in lower to upper fan facies, bypassing the mid-fan as no suprafan lobe deposits with proximal medium- to coarse-grained turbidites of quartzose provenance were observed. The strata could also be upper fan or slope facies, more distant from the feeder channel system. The sediment source was probably ultimately fluvial, as the extrabasinal clasts and sediment are fairly mature; this is seen in good rounding of clasts, high sand/shale ratios and good sorting in the conglomerates and arenites (even the coarsest and least-evolved turbidite beds have sorted, shale-poor populations, implying the source was similar), and composition (quartz and chert). In addition, Russell (1981) reported numerous plant fossils. The limestone and volcanic clasts, both inferred to be intrabasinal as both parental facies are in stratigraphic contact with basinal sediments (the limestone on the west side of the range, and the volcanic lithologies just up-section in this vicinity), are much more angular. The rounded tuffaceous and radiolaria-bearing chert clasts are inferred to have been reworked out of the McGill Canyon Formation or derived from the same source terrane, because of their pronounced similarity within the two formations.

Carbonate turbidite facies can be treated in the same fashion as clastic ones in terms of facies analysis (Stow, 1986). The calcarenite beds of facies B2 are too thick generally (5 cm) to be grain-flow products; instead they may represent the S_1 division with traction-sedimentation in a sandy high-density turbidite (Lowe, 1982) with an origin in an unconsolidated medium- to coarse-grained fossiliferous carbonate sand deposit. Deposition was probably from fairly proximal turbidite flows near a carbonate platform (Wilson, 1975).

The existence of such a carbonate bank or marginal reef system along this margin seems likely, as the calcarenite turbidites of facies B2 represent discrete turbidite mass-flow events in a basinal setting with a distinct provenance in the carbonate facies. The only mixing of calcarenite and quartz arenite sediment-types is in the A2 turbidite facies, where some angular limestone clasts are present; these clasts (which are Norian) could originate from already-cemented portions of the active carbonate bank (early cementation in carbonate facies is very common - Bathurst, 1975). It is unlikely because of their age and angularity that these are extrabasinal in origin. The distinct modes of deposition and provenance in the the calcarenite (B2), conglomerate (A2), and quartz arenite (D) turbidite facies assemblages taken together indicate separate extrabasinal terrigenous fluvial and intrabasinal carbonate bank sediment sources for turbidites being deposited in an upper fan and basin plain to lower slope setting. The hemipelagic background sedimentation is expressed in the argillite facies. The quartzose and cherty turbidites were probably fed in through gaps in the carbonate bank, or were reworked form shelf and slope deposits; clastic sedimentation rates on the shelf could not have been too high or the carbonate facies would not have developed.

The extrusion of the vesicular and pillowed basaltic flows in the immediately superjacent basal Happy Creek Formation took place at a depth of several hundred meters water or less, based on the size (several mm minimum) and relative abundance of the vesicles (Moore, 1965; Jones, 1969; Williams and MacBirney, 1979; Cas and Wright, 1987). This provides a constraint on the depth of the basin floor late in Bliss Canyon Formation deposition. The appearance of the Happy Creek Formation volcanics and volcaniclastics in the section is gradational but fairly abrupt, probably marking the sudden initiation of volcanic activity within this area of the basin. No depositional hiatus is thought to exist between the Bliss Canyon and Happy Creek Formations.

(3) Bottle Creek

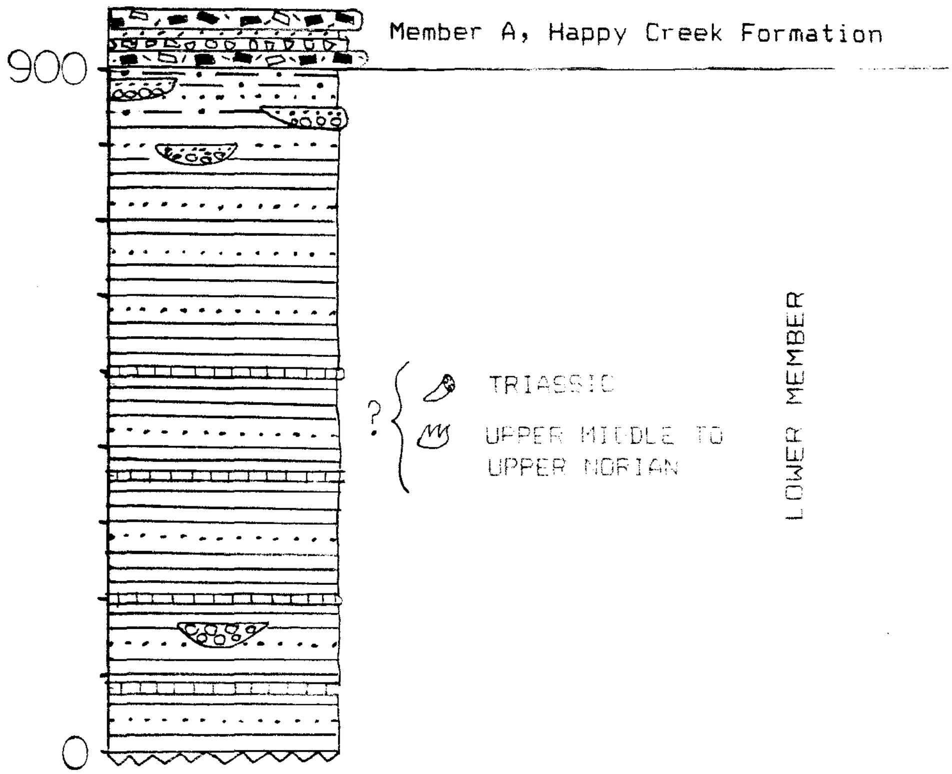
The Bliss Canyon Formation north of Bottle Creek is also involved in the west-vergent thrusting, and in several places within the nappes is tightly folded on an outcrop scale. A minimum of approximately 550 m and potentially as much as 900 m of section is found here (depending on whether one particularly thick thrust sheet is internally folded or not-no such folds were mapped, but exposure was poor). The formation in this area is again entirely put within the lower member. See fig. 10 for the stratigraphy.

Stratigraphy As elsewhere along the eastern side of the Jacksons, the dominant lithology is grey thin-bedded to laminated argillite. The next most abundant is thin-bedded grey to light grey fine-grained sandstone, then light grey calcarenite and calcrudite (carbonate conglomerate), then green chert-pebble conglomerates. The first three lithologies are thinly interbedded throughout the unit, though the calcarenite facies becomes less abundant near the upper few hundred m of section. The chert pebble conglomerates are rarer, and occur in laterally discontinuous channel-fill sequences and thick beds in two positions, near the top and less abundantly near the bottom of the section. At the top of the section the argillite and interstratified thin-bedded fine-grained sandstone passes fairly suddenly upwards to augite- and plag-phyric basaltic flows, flow breccias, coarse-grained green feldspathic arenites and interbedded argillite and fine-grained grey arenites (the latter similar to those in

FIG. 10

(3) Bottle Creek

Stratigraphy of the Bliss Canyon Formation, Jackson Mountains



the underlying sequence). The Bliss Canyon - Happy Creek contact is put at the base of this volcanogenic interval, which is seen in several thrust sheets. The base of the Bliss Canyon is not exposed in this locality or any others, due to structural truncation and stacking.

The lowermost section of the unit contains crinoidal biosparite in a matrix of grey argillite and fine-grained sandstone. Throughout the lower half of the section are thin to thick beds of light grey fossiliferous (and internally laminated with convolute bedding) calcrudite to calcarenite (sorted grainstone, recrystallized to the extent that original textures were sometimes hard to distinguish in the field). The argillite is rather featureless aside from depositional lamination except at the top of the sequence just below the Happy Creek contact, where it is more silty and bioturbation of the sediment is substantial. This latter occurrence is unusual for the lower member of the Bliss Canyon on the east side of the range.

An arenite from the upper part of the formation was petrographically examined. The rock was highly immature texturally (poorly sorted and subangular, though with little clay matrix) and consisted dominantly of volcanic rock fragments (trachytic and plag-phryic) and plagioclase grains, with additional tuffaceous and radiolarian chert, crinoidal biosparite allochems, and monocrystalline to mosaic metamorphic quartz grains. It is best classified as a feldspathic volcanic-arenite (Folk, 1980). Russell (1981) also noted the feldspathic nature of some of the sandstones in this area, but held that the overall nature was more quartzose and cherty, exclusive of volcanogenic zones I interpret to lie at and above the Happy Creek contact. Sandstones lower in the section were also observed in the field to be feldspathic to lithic arenites. The detrital grains have all undergone extensive pre-depostional alteration and silicification. The assemblage suggests a source in the upper member of the McGill Canyon Formation (or the same source as that unit) - the combination of radiolarian tuffaceous chert, crinoidal biosparite, volcanogenic rock fragments and feldspar and metamorphic quartz is distinctive. That there was a nearby, early phase of Happy Creek volcanism shedding feldspathic and volcanogenic sediments into the basin seems less likely, as (aside from the multilithologic nature of the assemblage) the first pulse of volcanism recorded in the Happy Creek is everywhere augite-phyric, and this volcanic component is not.

The chert-pebble conglomerates in the section have abundant rounded chert and quartzite clasts, as well as smaller argillite, volcanic and sandstone clasts, all also rounded (and hence in this case inferred to be extrabainal - the source again be in the McGill Canyon Formation, or in a common source). The conglomerates also contain intrabasinal angular blocks of sparry to micritic grey limestone. The matrix is arenitic, and there is clast-support. The more argillite-rich intervals throughout the section frequently display soft-sediment deformation, with very ductile tight and even isoclinal and rootless folds overprinted by the regional cleavage; this deformation of unconsolidated sediment thus predates thrusting. Flame structures were also noted at the top of the unit, just below the contact, and could have been caused by rapid loading by the volcanic pile above.

Age A boulder of grey biomicrudite was found in the basal King Lear Formation just to the north of this area, where the King Lear unconformably overlaps thrusted Bliss Canyon sediments. The boulder was in massive fluvial chert- pebble conglomerates that had a source in the Bliss Canyon and McGill Canyon Formations, judging by the remarkable resemblance of the chert pebbles and cobbles to those in these units. This contention is supported by chert-pebble conglomerates to the south in the range, which have radiolaria-bearing clasts, of probable Triassic and of Carboniferous and Permian age - identified by D.L. Jones (1980) in Russell (1981). Scleractinian corals from the boulder were dated by G.D. Stanley (written communication, 1986) as belonging to the family Volzeiidae, of Triassic age. In addition, D.L. Clark (pers. comm., 1986) identified conodonts from the same boulder as being intermediate between Epigondolella humboldtensis and E. postera, and of upper middle or upper Norian age. The sample certainly was reworked from the Bliss Canyon Formation locally, and implies that in this region it contained middle to upper Norian limestone strata.

Sedimentary Facies The sorting, parallel lamination, coarse grain size, thickness (10 cm to 1 m) and compositional purity of the calcarenites and calcrudites points to an origin from a gravelly or sandy high-density turbidite current, with deposition by traction-sedimentation (Lowe, 1982), as inferred from calcarenites in other areas already discussed. They would belong to turbidite facies A4 to B2 (Walker and Mutti, 1973). The background hemipelagic argillite (facies G) to fine-grained classical Bouma $T_{(C)DE}$ base-missing quartzose turbidite (facies D) sedimentation is as already discussed earlier, and took place either in a lower fan setting quite far down-current or as levee deposits near the head of the fan. The infrequent chert-pebble conglomerates - graded beds of Walker (1978) - were deposited in a feeder channel setting in the upper fan, by suspension-sedimentation from gravelly high-density current (the R_3 division) and then traction-sedimentation (S_1) from a sandy high-density current as the turbidite flow evolved (Lowe, 1982). At the very top of the unit, the sudden appearance of burrowing and of cross-lamination and lenticular bedding, and the more homogenous (in terms of bedding and grain size) nature indicate that the sedimentary facies may have suddenly become shallower, or perhaps just passed above storm wave-base, and were no longer so anoxic. The disappearance of the calcarenite facies might mean the calcarenite turbidites could not reach these shallower regions, perhaps because they lay up-slope.

The overall setting seems to be similar to that observed to the south - two distinct sources (carbonate platform slope-apron and terrigenous turbidite) in an upper fan setting (the association of facies A_2-B_2, D and G for the terrigenous component), and perhaps overall shoaling or regression (the apparent presence of shallower, shelfal facies at the top of the section).

(4) Buff Peak

Just south of Buff Peak is an area of thrust-faulted Bliss Canyon to Happy Creek lithologies with large, angular folds. Intrusion by rhyolitic hypabyssal dikes related to local King Lear volcanic centers (such as Buff Peak) is extensive. The Bliss Canyon in this area is particularly fragmented and obscured, and correspondingly uninformative and hard to piece together. The section must be at least on the order of 600 to 700 m thick. See fig. 11 for the stratigraphy.

Stratigraphy The Bliss Canyon section near Buff Peak is similar to that to the south, with the principal exception that in the limited outcrop exposure no chert-pebble conglomerates were noted. The section is composed dominantly of grey to black thin-bedded to laminated argillite, with frequent thin interbeds of grey coquinoidal (biosparudite) limestone and less common green micrite and cross-laminated very fine-grained grey sandstone. A calcarenite from the lower part of the interval was found to be an intraclastic biosparudite (dominantly broken, abraded and sorted bivalve and crinoid skeletal fragments with abundant intrabasinal intraclasts of argillite and biosparite, and rare clasts of plag-phyric volcanics and chert, in sparry cement). The calcarenite facies is more common than in the Bottle Creek area, and the sandstone facies less so. The section grades at the top into the basal green feldspathic sandstone and then the overlying plag-phyric volcanic flows and intercalated grey argillite of the Happy Creek Formation. This section has also been assigned entirely to the lower member of the Bliss Canyon Formation.

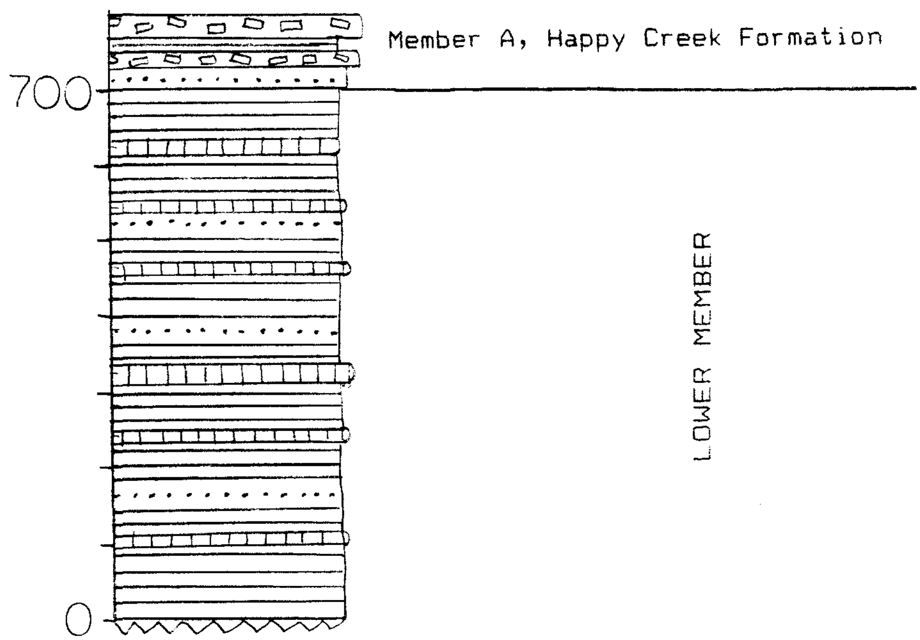
Age No faunal age constraints were found within this outcrop area (one sample was collected, but did not yield conodonts (D.L. Clark, pers. comm., 1985).

Sedimentary Facies The argillitic nature of the section and the general lack of coarser terrigenous clastics implies a distal shelf, slope or

FIG. 11

(4) Buff Peak

Stratigraphy of the Bliss Canyon Formation, Jackson Mountains



basinal facies. The carbonate grains in the calcarenite definitely had a shallow marine origin, but could also have been laid down on the shelf in a storm deposit or redeposited as a turbidite at deeper levels in the basin; I favor the former interpretation based on regional stratigraphic reconstruction (see later discussion). The nature of origin of the rare micrite interbeds is uncertain, and could even be diagenetic or repacement, or they could be shelf carbonate muds. Other constraints within the section are insufficient to distinguish between basinal and turbidite facies and more shelfal environments; the cross-lamination observed in the sandstones could form in either environment. Given the lateral variation within the other sections of the Bliss Canyon, slope to distal shelf environments with occasional calcarenitic storm or turbidite events, but bypassed by terrigenous turbidite currents, would be the favored origin.

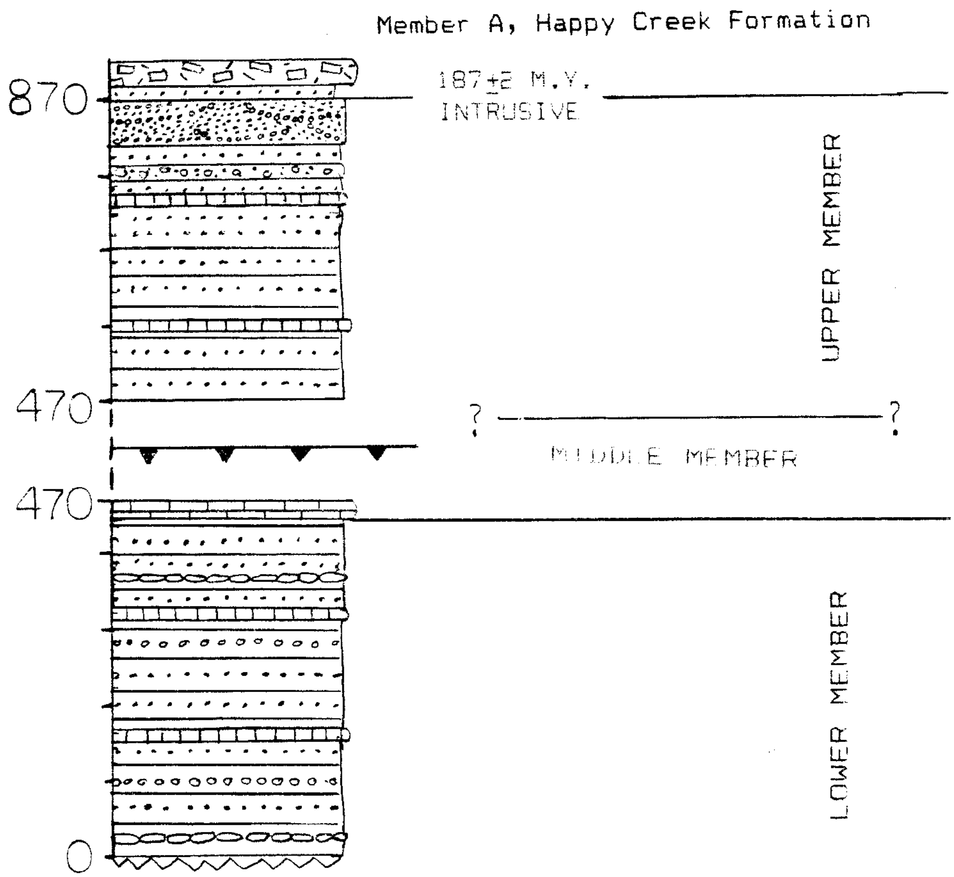
(5) Jackson Creek

This outcrop area is on the west side of the Jackson Range, just north and south of the mouth of Jackson Creek along the range front. The strata are involved in east-vergent thrusting and have undergone extensive contact metamorphism to high grade adjacent to several sub-volcanic intrusive bodies of Happy Creek and Early Mesozoic intrusive suite affinity. There is probably several hundred meters of section present. See fig. 12 for the stratigraphy.

Stratigraphy In the highest structural (and inferred lowest stratigraphic) position is an interval of meta-quartzite, meta-

FIG.12
(5) Jackson Creek

Stratigraphy of the Bliss Canyon Formation, Jackson Mountains



conglomerate, pelitic and psammitic schist (muscovite- and biotite-bearing), meta-chert and marble. Metamorphic grade is high (albite-epidote to hornblende hornfels facies), and will be treated in detail in chapter 6. Some of the original sedimentary characteristics can still be discerned. The quartzites were originally grey fine-grained well-sorted thin-bedded graded quartz arenites or subarkoses. Bedding style ranges from planar-laminated in the finer-grained, thinner beds to low-angle cross-laminated in the sandier, thicker units to massive beds several m thick. The clasts in the conglomerates were cherty and quartzose (with locally large detrital plagioclase grains) with a quartz arenite to subarkosic or wacke matrix and clast-support. The cherts were originally depositionally laminated and some contain recrystallized radiolaria, in a silty to argillaceous cryptocrystalline matrix. The limestone is extensively recrystallized to marble and can be sheared and brecciated, but some original textures are preserved. Crinoid and bryozoan skeletal fragments are discernable, but depositional texture is not. Bedding was probably thin but the extremely ductile behaviour of marble in high-temperature tectonite zones makes this uncertain. A large marble lens caps the sequence on one ridge crest. This interval is put in the lower member, with the capping limestone in the middle member. The lower member here is about 470 m thick.

In a structurally lower and stratigraphically higher position is a slightly different section. The metamorphic grade is somewhat lower here. The upper stratigraphic contact with the Happy Creek Formation is preserved, with green volcaniclastic sandstone and conglomerate and plag$\pm$augite-phyric flow rock interbedded with quartzite superposed over

Bliss Canyon metasediments. The latter unit contains phyllite, marble, quartzite and light green thin-bedded cross-laminated calcsilicate beds, passing upwards on the north side of Jackson Creek to quartz-pebble conglomerate (rounded, poorly sorted but compositionally mature with a quartz-arenite matrix, very low-angle cross-bedding and clast-imbrication, with scour and channel-fill sequences) with some lenses of green volcanic arenite. Because these volcaniclastics are still only a minor component of the section, this section is left in the Bliss Canyon Formation. On the south side of the creek the upper part of the structural section is composed of quartzite (low-angle cross-lamination to planar lamination), recrystallized marble, chert- and quartz-pebble conglomerate (rather angular, with matrix support), and pelitic to psammitic schist. This interval on both sides of Jackson Creek is assigned to the upper member. Clast imbrication within a cross-bedded quartzose and cherty conglomerate at the top of the section in the upper part of New Years Canyon suggests a source to the north and east. The sandstones were originally quartz arenite to chert arenite. The upper member here is thought to be about 400 m thick.

Age No faunal age constraints are known for the Bliss Canyon Formation in this outcrop area; a recrystallized limestone was collected for conodont analysis but did not have any fauna (D.L. Clark, pers. comm., 1986). Preliminary U/Pb zircon dating on a pluton intruding the Bliss Canyon and Happy Creek Formations in this area gave an age of 187±2 m.y., so the upper limit on the formation is Early Jurassic.

Depositional Environment The portion of the sequence placed in the lower member of the Bliss Canyon Formation is noticeably coarser, thicker bedded and more proximal than the argillite/fine-grained arenite facies of the lower member on the eastern side of the range. The sedimentary structures observed in the quartzites (grading, with planar-lamination in the finer-grained beds to cross-lamination in the coarser-grained and thicker lithologies) are compatible with traction- to suspension-sedimentation from a low-density sandy turbidite current - the classical Bouma T_{CDE} sequence (Lowe, 1982), and with facies C to D of Walker and Mutti (1973). The metaconglomerates (though sedimentary textures are mildly overprinted in the field) probably represent distal (down-current) distributary channel scour and channel-fill of facies A2. The laminated phyllite and metachert represent the hemipelagic sedimentation of facies G.

The best facies interpretation for this set of lithologies is the smooth to channelled transition area on the supra-fan lobe of the mid-fan region (Walker, 1978). Too little is preserved within the limestone lens placed in the middle member to make sedimentary facies assignations.

The strata placed within the upper member are quite distinct from all those so far discussed from the Bliss Canyon. On the north side of the creek they are much coarser overall, thicker-bedded and higher-energy in origin, and conspicuously lacking in phyllite in the upper portion. The rounded and compositionally mature metaconglomerates and quartzites or metachertarenites, with pronounced clast imbrication, channel scouring and low-angle cross-sets and lacking in phllite seem more typical of foreshore facies in a wave-dominated high-energy coarse-grained shoreline or bar environment (Elliott, 1986). The lower part of the interval, with thinly interbedded marble, phyllite, quartzite and cross-laminated calcsilicates (originally marl), might be of an offshore shelf facies. On the south side of the creek, the upper part of the section was sandier (with low-angle cross-lamination to parallel lamination), the chert-pebble metaconglomerate beds seen had arenite matrix support (and hence were more likely to be debris flows), and there is interbedded phyllite and marble. The sequence is most likely transition zone, farther off-shore but still fairly high-energy and shallow. Some idea of paleogeography can be gained from this; the shoreline in this area lay to the north, which agrees with the observed clast-imbrication.

The upper member of the Bliss Canyon in the area of the mouth of Jackson Creek was deposited in a shoaling sequence from shelfal to foreshore in a high-energy, coarse-grained, wave-dominated environment with a shore line to the north.

(6) Bliss and Alaska Canyons

The best exposures of the Bliss Canyon sequence lie in this area, and hence the proposed type section. However, there is rather complex, tight folding and thrust faulting of the section. Metamorphic grade is much lower than in the region to the north, and more of the depositional characteristics are preserved. The stratotype for the Bliss Canyon Formation is here designated to lie in the area in and south of Alaska Canyon (north of an east-west high-angle fault, which juxtaposes these rocks against the McGill Canyon Formation). There is approximately 900 m of section present; the base is not exposed but the upper stratigraphic contact is present. See fig. 13, for the stratigraphic section.

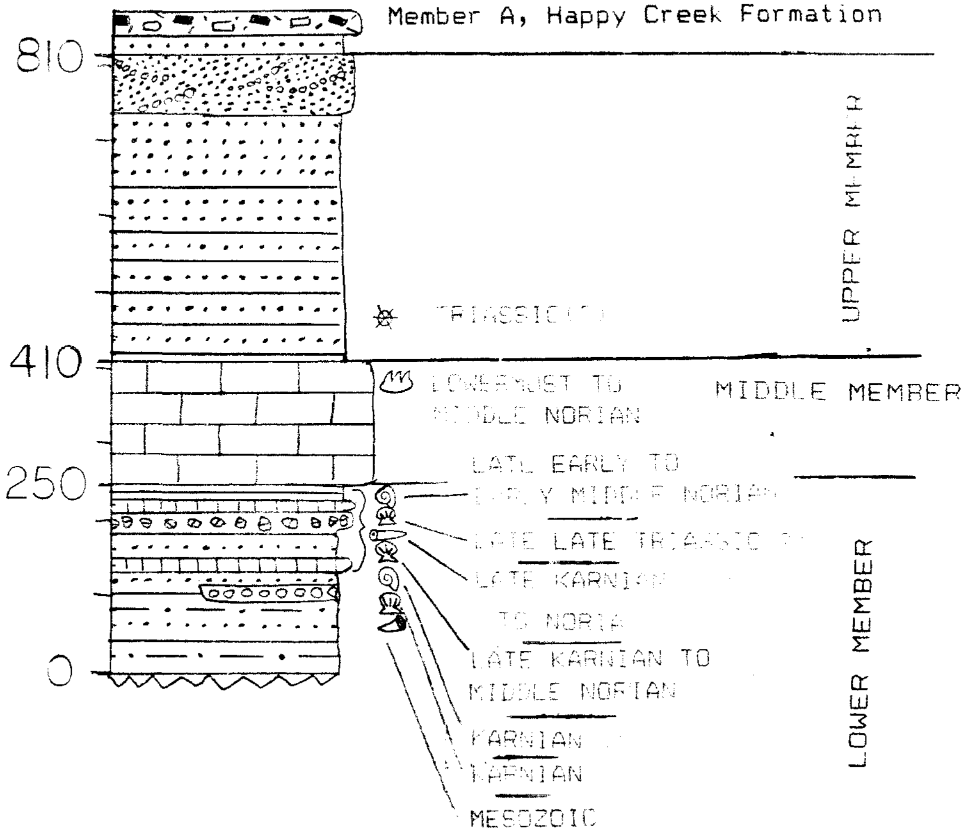
<u>Stratigraphy</u> In the lower part of the section there is approximately 100 m of grey laminated argillite and siltstone with oscillation ripples and lenticular cross-lamination, thin- to thick-bedded green and grey quartzose gritstone and arenite (immature, subangular and poorly sorted, with intrabasinal lithic fragments), and thin-bedded chert. This passes upwards to an assemblage of carbonate breccias and conglomerates (with a quartz arenite or gritstone matrix to arenaceous micrite matrix), thin-bedded and laterally continuous extremely fossiliferous biosparite and biomicrite, grey argillite and shaly micrite, and quartzose and cherty conglomerates (mature, well-sorted and very clean). All lithologies are arenaceous (chert and quartz grains are abundant, even in the carbonate facies). The carbonate component increases volumetrically up-section, and the limestone clast-bearing beds coarsen and thicken. These carbonate breccias commonly have matrix-support (floatstones), while the biosparite and biomicrite exhibit clast-support (packstones). The clastic component seems to be predominantly chert and quartz. This entire sequence is assigned to the lower member, and is about 250 m thick.

Fuller (1986) described in detail a portion from the top of this sequence from the north slope near the mouth of Alaska Canyon - the following is a summary. Among the the lithologies mentioned were quartzose intramicrite (wackestone), bioclastic intramicrite (wackestone), and biopelmicrite (packstone) beds interbedded with graded laminated to bioturbated calcareous mudstone and quartzose micrite, from

FIG. 13

(6) Bliss Canyon (the stratotype)

Stratigraphy of the Bliss Canyon Formation, Jackson Mountains



the upper 50 to 60 m of the lower member. Bioturbation was abundant through the interval. Textures include floatstone (in the conglomerates) and wackestone to packstone (in the carbonate sand facies). The sandstones commonly had sharp tops and bottoms and weak grading, with basal scour and a high sand-shale ratio. The floatstones are channelized and laterally variable in thickness, have scoured and abrupt bottoms, are massive and normally (though poorly) graded and have some clast imbrication. Though not noted in the text, one photograph in Fuller (1986) also shows clear reverse grading at the base of one such bed. Quartz and chert detrital grains were ubiquitous, but feldspar and lithic fragments were absent, with the exception of one unusually rounded and clean conglomerate facies with feldspar and volcanic, plutonic and metamorphic lithic clasts. Some of the clasts in one intramicrudite included probable Paleozoic corals and silicified fusulinids, implying a late Paleozoic source terrane (the McGill Canyon Formation?). Paleocurrent analysis by measurement of clast lineation in several of the floatstones (Fuller, 1986) indicates current flow in a NNE-SSW direction (possibly along the basin axis, or alternatively directly down-slope).

Overlying this sequence there lies a very continuous and distinctive massive, recrystallized light grey to tan limestone bed, from ten m or so to as much as several hundred m thick (and thought to be 160 m in the area of the stratotype). The variability is due to ductile thinning and thickening in areas of tight folding related to thrusting (see chapter 7). This limestone locally preserves depositional features, with biosparudite (a coquina with micritic intraclasts) and biomicritic textures, but generally is now medium- to coarse-grained sparry calcite in the areas of greater deformation. Bedding was not preserved, but was probably very thick. No other lithologies were seen interbedded with the limestone except within a few meters at the upper and lower contacts. This limestone represents the middle member of the Bliss Canyon Formation, and appears to thin to the north (especially near Jackson Creek, this could be due to either depositional lateral variation or tectonics). The bioclasts and intraclasts in one intraclastic biosparudite had micrite coats or "sweaters", implying a shallow-marine stage in their history (Bathurst, 1975). Fuller (1986) also examined the lower part of this interval, and described it as dark grey massive pelmicrite and crinoidal biopelmicrite (crinoidal, peloidal and intraclastic wackestone and packstone) with detrital chert grains and crinoidal biomicrite, conformably but suddenly overlying the finer-grained and less massive facies of the lower member.

The limestone of the middle member is overlain abruptly by grey argillite and interbedded quartzose and cherty gritstone and sandstone of the upper member. These facies coarsen and thicken gradually upwards. The upper part is made up of light green to grey quartzose and cherty conglomerate and sandstone with good rounding and sorting, thick-bedded with small scoured channels and channel-fill, and cross-lamination and cross-bedding in the thicker beds. This passes fairly suddenly upwards into the Happy Creek Formation - several meters of green arkosic arenite overlain by olivine-amphibole-plag-phyric volcanic flows. This whole interval is assigned to the upper member, and is about 400 m thick. The upper member may thin somewhat to the north, unless this apparent change is caused by later tectonic factors.

Age Fragmentary conodonts collected in 1986 from the massive limestone of the middle member were identified by D.L. Clark (written communication, 1987) as either Epigondolella primitia or E. postera, and hence are of lowermost to middle Norian age. In addition, a collection of ammonites, belemnites and bivalves was found in 1986 from the very top of the lower member south of Alaska Canyon and identified by N.J. Silberling (pers. comm., 1986). The collection included Juvavites, a haloritid ammonite of late early or early middle Norian age; Aulacoceras, a ribbed coleoid ("belemnite") of late Karnian to Norian age; and pectenacid bivalves (not Monotis) of possible late Late Triassic age, as well as bone and crinoid fragments. The specimens were probably reworked into this deposit from unconsolidated sediment within the basin.

Russell (1981) also reported biostratigraphic data from this outcrop area: (1) Late Karnian to middle Norian Halobia cf. H. cordillerana found by R.C. Speed in 1978 and identified by N.J. Silberling in 1979, and coming from the top of the lower member; (2) Mesozoic coral in boulders in a conglomerate, identified by N.J. Silberling in 1978, and probably from the upper lower member; (3) Early Mesozoic and probably Triassic radiolaria in bedded chert and cherty limestone, identified by D.L. Jones in 1979, with an origin in the bottom portion of the upper member.

Fuller (1986) presented additional fossil age control from Alaska Canyon. The assemblage came from what I interpret as the upper portion of the lower member, within the core of a tight anticline (but at a slightly lower stratigraphic interval than the collection mentioned above), but which Fuller originally thought to be a stratigraphic rather than structural succession. Halobia cf. H. superba and H. austriaca (pelagic bivalves, both of Karnian age), Arcestes pacificus and A. carpenteri and Discotropites mojsvarensis (ammonites, also of Karnian age) were reported by Fuller, who pointed out the very close resemblance of these specimens to those of the Hosselkuss Limestone (see later regional discussion for details). The Halobia are too fragile to have been transported.

The lower member is well-dated as Karnian to early or early middle Norian, and the middle member is of middle Norian age (possibly extending into the late Norian, as the base of the upper member is probably still within the Triassic). A minimum age on the upper member was not determined from this outcrop area.

Depositional Environment The stratigraphy and lithologic characteristics in this area reveal a great deal about the nature of the sedimentary facies, which differs from that inferred on the east side of the range. Even the base of the lower member here has some evidence that the sequence was laid down above storm wave-base (the presence of well-developed and extensive oscillation ripples within one horizon). Though the rest of the features of the lower part of the lower member indicate hemipelagic (argillite, siltstone, chert) and proximal (up-current) Bouma T_{A-E} turbidite current sedimentation, the basin must have been fairly shallow. The presence of turbidite currents, for example, means only that the environment was below normal wave-base so that the sediment was not systematically reworked; such events are storm-related tempestites and basinal depths are not necessarily implied. In the upper part of the lower member, the abundant carbonate floatstone beds (also called diamictites and calcidebrites - Stow, 1986) were deposited by a cohesive debris-flow mechanism, as indicated by the matrix-support, lateral variability, and poorly-developed clast imbrication and normal grading. These flows might easily evolve into high-density gravelly turbidite currents. That this did happen is indicated by the presence of such very proximal (e.g., upcurrent) turbidite beds in the section - clast-support conglomerates with a sandy matrix, reverse grading sometimes present at the base and normal grading otherwise, and parallel lamination in the sandier overlying beds (a R_2-R_3-S_1 succession of sedimentary divisions- Lowe, 1982). Some of the thin clean and well-sorted biosparite beds might in addition represent a grain flow depositional mechanism. If so, this would imply quite steep slopes.

Two sources were involved - an active (Karnian to Norian) carbonate platform and a terrigenous supply of quartzose and chert-rich sediment, at least partly late Paleozoic (the McGill Canyon Formation?). A significant slope-angle is concluded. The mixed terrigenous and carbonate facies were also probably deposited from such a composite source, by more normal downslope reworking sedimentation processes. The carbonate platform was not a reef proper, but more a Bahamas-style composite of various very shallow-water carbonate facies (with peloidal, micritic, skeletal and intraclastic sedimentary facies). Much of these facies would then be redeposited down-slope. Boundstone facies (and the organisms that might cause them) and oolitic lithologies are not seen in the debris, so energy levels were probably moderate and not high. This coarsening-upwards sequence in the lower member near Bliss and Alaska

Canyons was shed off a mixed carbonate barrier platform and clastic shelf onto a fairly shallow basin floor to slope-apron environment (a depth of a few hundred meters or less seems likely, based on the oscillation ripples). The upper part of the lower member would represent arpon-slope debris being shed from the carbonate platform as it prograded into the area from an unknown direction. No evidence for volcanism exists in this section.

The middle member - the massive, largely recrystallized limestone unit - was the carbonate platform proper. As noted in the discussion on carbonate debris, the unit does not imply a boundstone coral-style reef. Instead, a very shallow moderate-energy set of facies was found (subtidal to lower intertidal, with water depths probably less than 10 m). Oolitic and reefal high-energy facies were missing, but peloidal, skeletal and intraclastic lithologies are more common than micrite (which is indicative of low-energy conditions - Bathurst, 1975). Both sparry cement (implying the allochems were washed clean) and micritic matrix (no such current working) were found. The diverse and abundant fauna also indicates normal marine conditions (crinoids, ammonites, belemnites, pelagic and benthic bivalves, bryozoa, and conodonts all existed in this time period). Bioturbation and peloid formation point to substantial organic activity. The great thickness (several hundred meters) points to high sedimentation rates, which is standard for such carbonate facies. The carbonate platform was apparently an offshore barrier bank, but could not have been too distant from the clastic shoreline because of the presence of the fine-grained terrigenous fraction. The bank does seem to thin and disappear to the north, and not all of this is inferred to be tectonic in origin.

Overlying these carbonate platform facies abruptly is the upper member - argillite with thin interbeds of sandstone and conglomerate, shoaling upwards to thick-bedded conglomerate and sandstone with good rounding and sorting, frequent small scour-and-fill structures, and cross-lamination to cross-bedding in the thicker and coarser lithologies. The best interpretation for these deposits is an upwards evolution from offshore shelf to transition facies to shoreface, perhaps within a very large back-platform lagoonal setting. This is similar to the interpretation for the Jackson Creek area. The cross-bedding, lack of finer lithologies, textural and compositional maturity and scour-and-fill in the upper part of the upper member are particularly characteristic of the shoreface (Elliott, 1986). There is no evidence of active volcanism in this part of the section either. The contact with the underlying carbonate platform of the middle member could be either transgressive or could be caused by continued migration of the carbonate bank offshore, and deposition of the upper member in the lagoon or marginal seaway behind the platform. Such instances are reported from Honduras in the present and the Early Jurassic of Europe (Sellwood, 1986). No back-reef facies were found in the back-platform area.

The Bliss Canyon Formation in the type locality exhibits an overall shoaling and regression from a shallow-floored basin with hemipelagic and distal turbidite sedimentation, through a combined clastic and carbonate slope-apron with debris flows and proximal turbidites, to a moderate energy offshore carbonate bank, and upwards through shelf and transition zone to shore-face environments. Paleoslope may have been from northwest to southeast, judging by facies relationships and the paleocurrent analysis of Fuller (1986).

Depositional Environments and Stratigraphic Framework for the Bliss Canyon Formation

From the stratigraphic, paleogeographic and sedimentary facies information developed above, a coherent depositional framework and paleogeography for the Bliss Canyon Formation can be developed (fig. 14, 15). A number of distinct environments are represented within the formation. On the eastern side of the range, the southern three outcrop areas display basin plain and various turbidite fan complex settings, starting in the Ladinian and extending into the Norian. Section 1, in the southern Trout Creek Spur, was deposited in a lower, distal terrigenous turbidite fan setting (fine-grained, low density turbidite current and hemipelagic sedimentation) adjacent to an active carbonate bank from which coarse-grained, high-density turbidite currents were also coming. The latter were, however, considerably more proximal in terms of down-current evolution, reflecting proximity to the carbonate bank slope-apron. Section 2, around Burro Bill Springs, Boulder Creek and the southern part of Bottle Creek, was laid down in the prograding upper reaches of a clastic turbidite fan - specifically in the terraced portion of the feeder channel and the adjacent lower slope/basin plain, as recorded by the up-current proximal terrigenous coarse-grained high-density turbidite currents (though thin-bedded and laterally continuous) which appear over more distal, finer-grained turbidite facies. Both are interbedded with hemipelagic sediments. The carbonate platform-source

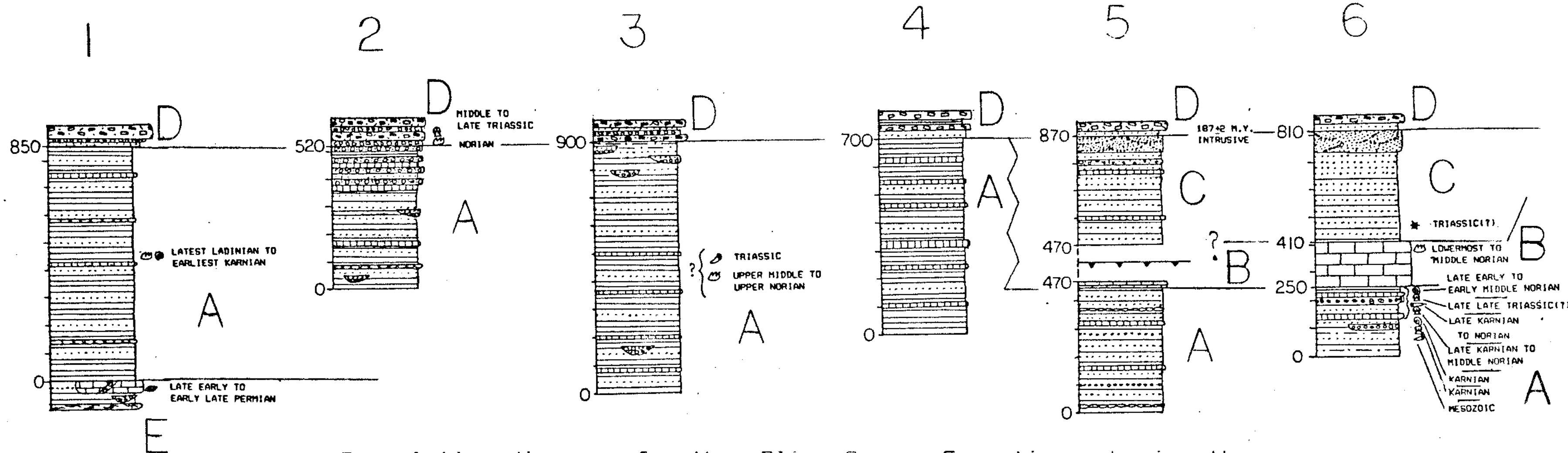


FIG.14 Correlation diagram for the Bliss Canyon Formation, showing the composite stratigraphic sections constructed for that unit and the proposed correlation of the lower, middle and upper members. Section locations are given on the accompanying map. (1) Trout Creek Spur; (2) Boulder Creek; (3) Bottle Creek; (4) Buff Peak; (5) Jackson Creek; and (6) Bliss Canyon (the stratotype).

A - Lower Member B - Middle Member C - Upper Member
D - Happy Creek Formation E - McGill Canyon Formation

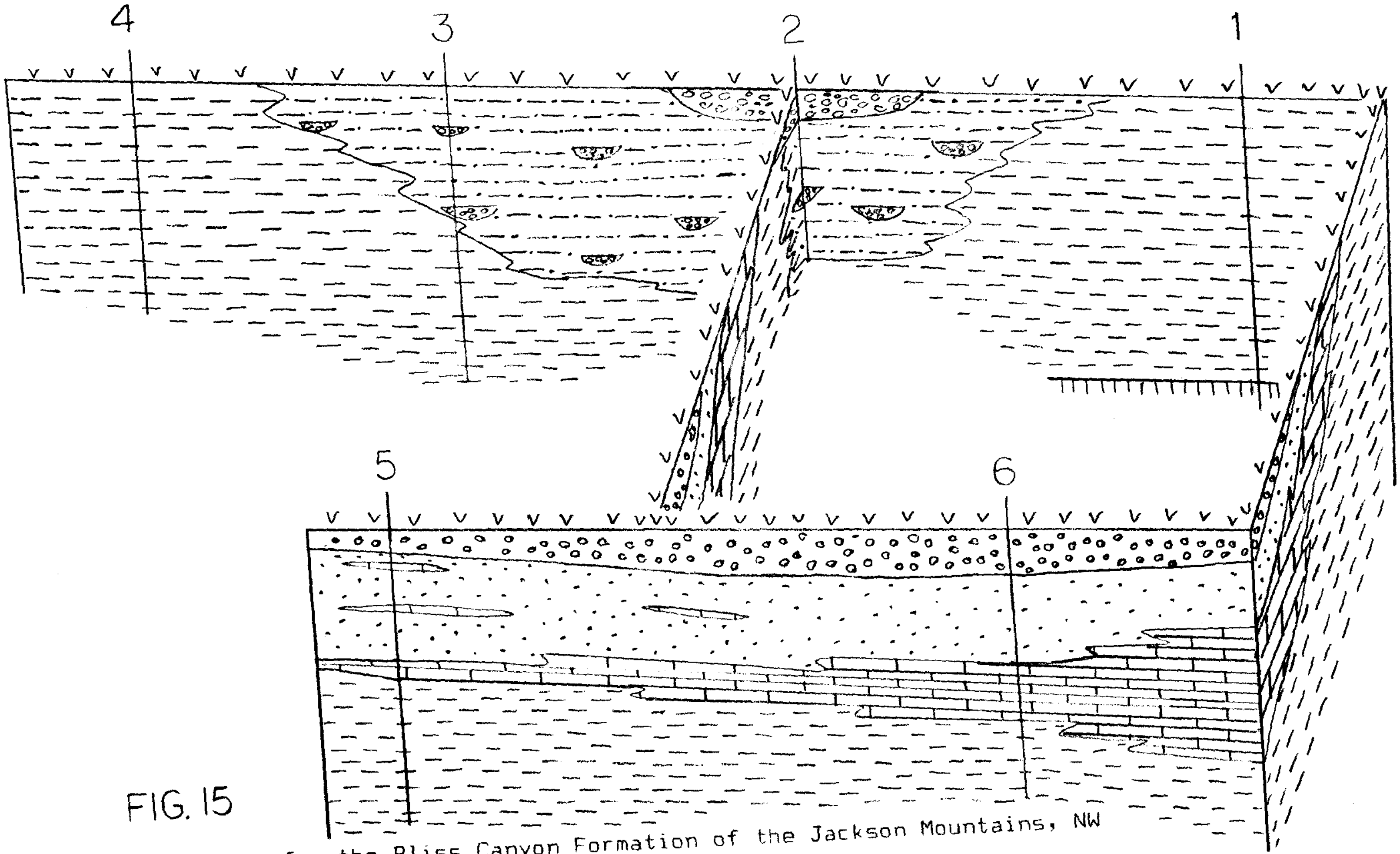


FIG. 15

Fence diagram for the Bliss Canyon Formation of the Jackson Mountains, NW Nevada. Section locations are shown on the fence.

facies symbols

oooo supratidal and littoral

carbonate platform

turbidite feeder channel

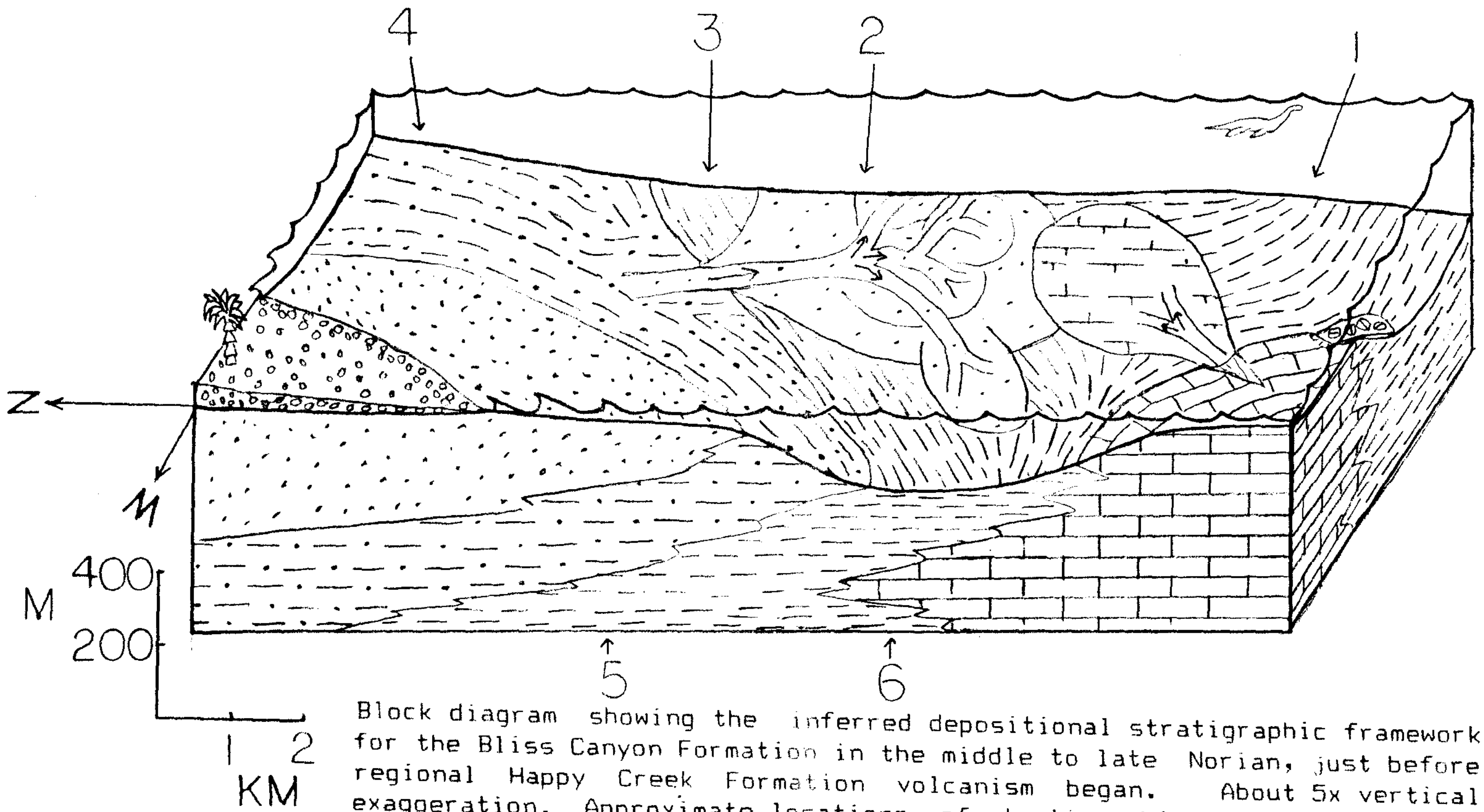
McGill Canyon Formation

shoreface to transition zone

lower and middle turbidite fan

slope and basin plain

VVVV Happy Creek Formation



Block diagram showing the inferred depositional stratigraphic framework for the Bliss Canyon Formation in the middle to late Norian, just before regional Happy Creek Formation volcanism began. About 5x vertical exaggeration. Approximate locations of stratigraphic sections given by the numbers (see map for positions).

FIG. 16

facies symbols

- supratidal and littoral
- distal shelf
- carbonated platform
- carbonate debris flow
- shoreface to transition zone
- lagoonal or basin plain
- carbonate turbidite fan
- clastic turbidite fan

proximal turbidites (similar to those in section 1) are also present here. Section 3, from north of Bottle Creek, is characterized by upper fan sedimentation in the terrigenous clastics, and by features in the carbonate turbidites similar to those already described, although the latter are missing in the upper part of the section (perhaps cut off by the prograding submarine fan suprafan). All three sections (except the top of 3, which is distinctly shallower with bioturbation recording a more normal and oxygenated environment) were deposited in an anoxic basin; the depth of the basin could, however, have been quite shallow. Continuing to the north, section 4, near Buff Peak, appears to be somewhat shallower in depositional environment, with a shelfal location inferred. The calcarenites here might be shelfal tempestites (storm deposits).

On the western side of the range, in section 5 from the mouth of Jackson Creek, the strata are much shallower yet in origin. They shoal up to shoreface settings (with clean and well-sorted cross-bedded, scoured and channelled conglomerates and sandstones) near the top from a middle fan setting (hemipelagic argillite and chert with thin quartzose turbidites and larger chert-pebble conglomerate lenses) at the bottom. To the south, around Alaska and Bliss Canyons in section 6, the section records sedimentation in locales shoaling from basin plain (though sedimentary structures indicate a position above storm wave-base) to carbonate platform "fore-reef" apron-slope to carbonate platform to shelfal to shoreface settings. The basin plain facies is characterized by hemipelagic accumulation of argillite, chert and siltstone, and proximal turbidite facies sandstone and fine-grained conglomerate beds.

The fore-reef deposits are characterized by carbonate diamictite breccias (debris flows) and clast-support very proximal calcarenite turbidites. The carbonate platform (middle member, in sections 5 and 6) dates from early to middle Norian; elsewhere in the region it must be older (Karnian and maybe even into the Ladinian) as it acted as a sediment source to the basin (sections 1 to 3 and 6). The offshore carbonate barrier is distinguished by moderate-energy facies (apparently no framework developed and peloidal, intralcastic, and skeletal grains dominate, with both micrite matrix and spar cement present), and must have been quite near a terrigenous sediment source because of the high proportion of arenaceous sediment found in the carbonate facies. The overlying interval (upper member - argillite and terrigenous clastics coarsening and thickening upwards to cross-bedded littoral sandstone and conglomerate) probably records a back-barrier lagoonal to shoreface transition.

Paleogeographic indications also aid in restoration of the local stratigraphic framework: in section 5, the paleocurrent direction may have been from the north and east; and in section 6, carbonate debris flows came from the northeast or southwest (Fuller, 1986; I favor the latter interpretation based on other aspects of the paleogeography). The source for the proximal carbonate debris in sections 1 to 3 almost certainly lay to the west in the carbonate platform facies seen in section 6. The source for the clastic sediment in the submarine fan facies of sections 1 to 3 is inferred to have been to the north and west in the direction of the coeval shallower facies of upper sections 5 and 6 (and 4?). The formation on the western side of the range shows a change in depositional environment to littoral or shoreface facies by the end of deposition in the late Norian. This was followed by the sudden and regional inception of very shallow marine to subaerial Happy Creek volcanism. Shoaling is also seen on the eastern side of the Jacksons in sections 2 and particularly 3; section 4 might also have been quite shallow throughout.

There appears then to have been a clastic sourceland and shoreline to the north and west of the Jacksons, with regional regression, offlap and progradation (Brown and Fisher, 1977) during slowly rising sea level (Haq and others, 1987). A barrier carbonate platform was located to the west and south (in this area at least; elsewhere it might have been directly fringing on the coastline), and started developing by the Karnian and possibly earlier. Both sources shed sediment via turbidite currents to the east into a shallow-floored but anoxic and restricted basin, whose record goes back as far, perhaps, as the Ladinian in the Jacksons. This basin was oversupplied. Oversupplied basins have high sediment supply and rapid infilling by progradation, and typically are small and somewhat confined (Stow, 1985). Late Norian volcanism of the Happy Creek Formation overwhelmed pre-existing sedimentation patterns, and established a new stratigraphic framework (see chapter 4 for details).

What the stratigraphic situation at the base of the Bliss Canyon was is not certain, as that contact is not typically present. A disconformable or even conformable relationship is possible in the area of section 1, implying basin continuity from at least the Mississippian (McGill Canyon Formation) to the middle Norian.

Sedimentation rates for the Bliss Canyon Formation can be estimated at about .02 m/1000 yrs for the basinal part of the lower member (from the upper lower member of section 1) and at about .05 m/1000 yrs for the shallower facies (upper lower member to the upper member of section 6). The former value is low for anything but a continental shelf or slope or cratonic basin - especially low for a trench or a successor basin (Schwab, 1976). The latter rate is very high for a cratonic basin, and falls within the range proposed for the continental shelf and foreland basin setting, and is low for any other likely environment (including successor basins). Though hardly conclusive, these rates add to the theory that the Bliss Canyon Formation was probably deposited in shallow basin to shelf setting. They also provide some negative evidence for a location in certain petrotectonic environments, such as trenches (implying no major suture exists between these rocks and the craton), cratonic basin (no underlying continental crust) and successor basin (which agrees with the lack of evidence for the Sonoma orogeny in this area).

CHAPTER FOUR:
THE LATE TRIASSIC TO MIDDLE JURASSIC STRATIGRAPHY OF THE JACKSON MOUNTAINS: THE HAPPY CREEK FORMATION

The Happy Creek Formation is here proposed to contain the latest Triassic to Middle Jurassic age volcanogenic strata of the Jackson Mountains. The unit was originally defined by Willden (1963; 1964) as the Happy Creek Volcanic Series, consisting of Permian or older age and grading upwards into unnamed Permian and Triassic beds, and composed dominantly of volcanic with some sedimentary rocks. The unit was redefined by Russell (1981; 1984) as the Happy Creek Igneous Complex. Russell established that the age range was in the span Late Triassic to Jurassic, and that Willden had originally inverted the stratigraphic sequence. As the formal definition of a complex requires "highly complicated structure to the extent that the original sequence of the component rocks may be obscured" (I.U.G.S., 1976), and the stratigraphic and intrusive and structural relations of the Happy Creek Igneous Complex of Russell (1981; 1984) are now well characterized, it seems more appropriate at this point to redefine the unit as the Happy Creek Formation. As used here, the Happy Creek Formation also includes volcanogenic exposures assigned to the Boulder Creek beds by Russell (1981; 1984); detailed structural mapping in the area has established that these exposures are tectonically and not stratigraphically intercalated with the basinal sediments of the Boulder Creek beds (renamed the Bliss Canyon Formation - see chapter 3).

The contact between the two units is defined as the point at which the volcanogenic fraction of the section becomes volumetrically the most significant. The contact is distinctive, constant in nature and readily mappable through the range. Though there are exposures of the Happy Creek Formation in Happy Creek in the northern Jacksons, better exposures exist of both the basal portion and lower contact with the Bliss Canyon Formation (around Boulder Creek and Bottle Creek on the east side) and the overlying parts and upper contact with the King Lear Formation (the Jackson Creek to King Lear Peak region, in the central and southern area of the range). However, as the term Happy Creek has seen extensive usage in the literature it will be retained as the formal stratigraphic name of the unit of formation rank defined here. No stratotype will be given as no one section fully or adequately represents the unit. Instead the enitre area of the Jackson Mountains mapped in this project will be designated as a type locality (I.U.G.S., 1976).

Outcrop of the Happy Creek Formation is quite extensive, and it is by area of exposure the most abundant unit in the range, with about 180 km^2 of outcrop. The unit is found in the two allochthonous thrust-faulted zones on the eastern and western flanks of the range, and also takes up a large portion of the autochthonous central buttress of the range that the allochthonous thrusts packages are imbricated against. Because of the massive and resistant nature of the bedded volcanics, internal deformation in a section is usually not intense. Generally, suitable stratigraphy exists so that structures can be mapped. Bedding placement and thickness estimates in the following discussion are therefore more precise than for the Bliss Canyon or McGill Canyon Formations. A distinct and coherent stratigraphy was found to exist in the Happy Creek

Formation, with significant and mappable lateral facies variations. No one unfaulted section displays the entire interval, and several large plutons intrude and locally metamorphose the unit, but with this stratigraphy the nature of the depositional, intrusives and structural relationships can be well characterized. The formation also has a complex relationship with an episode of late syn-volcanic east-west trending high-angle faults. Also associated with the Happy Creek Formation is a cogenetic suite of hypabyssal thick sills (on the east) and dikes and plutons (on the west) of gabbroic to dioritic to monzonitic composition, which are included in the Early Mesozoic Intrusive Suite and discussed in chapter 6. Locally the metamorphic grade is high (metamorphism in the range will be discussed in chapter 7). A more significant problem is very low grade, but regionally pervasive, auto-hydrothermal alteration, which is not uncommon in such volcanic piles and which made petrographic analysis difficult due to the rarity of fresh samples.

The Happy Creek Formation has been divided up into a series of informal members, defined on the basis of facies and lithology. These include (from oldest to youngest): (A) augite $\pm$ plagioclase porphyry basaltic andesite flows and basal green coarse-grained epiclastics (also containing thinly interbedded argillite, quartzose and cherty arenite, and chert-pebble conglomerate and limestone breccia at the contact with the underlying Bliss Canyon Formation on the eastern side of the range); (B) oolitic to crinoidal limestone, stromatolitic micrite and siltstone and flat-pebble conglomerate (interfingering with the augite-phyric flows of member A in the north of the range); (C) green immature volcanic wacke and arenite and conglomerate, siltstone and shale; (D) red coarse-grained volcanic arenite and wacke and pebble to cobble conglomerate (both the latter sedimentary units present as a thick interval between the andesitic to dacitic flows and related rocks of units A and E, and similar but smaller lenses exist in members E, F, G and H but have not been differentiated); (E) green to grey plagioclase ± hornblende andesite porphyry flows, flow-breccia and diamictite (volcanic conglomerate with matrix-support), with some subaerial weathering and red bed conglomerate lenses; (F) light grey plagioclase + quartz ± hornblende ± biotite porphyritic dacite and andesite flows, flow breccia and diamictite, and green volcanic arenite lenses; (G) green to red (oxidized, or with paleosols) plagioclase + hornblende ± augite andesite flows, flow-breccias and diamictites with lenses of volcanic epiclastics sediments; and (H) red plagioclase ± hornblende ± augite andesite flows, and very coarse- to fine-grained highly immature red and green volcanic epiclastics, deposited in late structural basins bounded by east-west trending high-angle faults that were characterized by local Fe-mineralization.

In addition, there are the components of the Early Mesozoic Intrusive Suite: gabbro and diorite sills, and gabbro, diorite, tonalite and granodiorite stocks. See chapter 6.

The contact with the underlying Bliss Canyon Formation is gradational and conformable, with the relatively sudden interbedding of augite-phyric flows, flow breccias and volcanic epiclastics in with argillite, quartzose sandstone and chert-pebble conglomerate in a zone tens of meters thick, before the latter sedimentary beds disappear altogether.

The base of the Happy Creek is put at the base of this zone, which is exposed almost everywhere in the range. The contact with the overlying King Lear Formation is also widely exposed, but is much more complex as it involved syn-depositional activity on the east-west trending high-angle faults and local unconformable overlap due to later west-vergent thrusting. In places this contact is a buttress unconformity with growth-fault relationships across the edge of a basin, in other places a disconformity to angular unconformity (associated with syn-sedimentary thrusting during the deposition of the King Lear) with pronounced weathering of the Happy Creek underneath, and in yet other places the contact is gradational with interbedding. The top of the Happy Creek is placed at the bottom of the first coarse clastic bed of the King Lear Formation.

The Happy Creek Formation will be discussed by geographic area for convenience and clarity. The areas include, counterclockwise from the extreme north (see fig. 17): (1) the northwest tip of the Jacksons; (2) the contact metamorphosed and thrusted outcrop around the mouth of Jackson Creek; (3) the area north of Jackson Creek; (4) the area of Hobo Canyon; (5) the area of Bliss Canyon; (6) around King Lear Peak, (7) the southern tip of Trout Creek Spur; and (8) DeLong Peak region (including the Iron King Mine).

The discussion of the Happy Creek Formation in each area will include: (1) description of the field and petrographic observation of lithology, volcanic and sedimentary features and relationships, and provenance and paleocurrent data (if any) by area; (2) age data when available, both biostratigraphic and geochronologic; and (3)

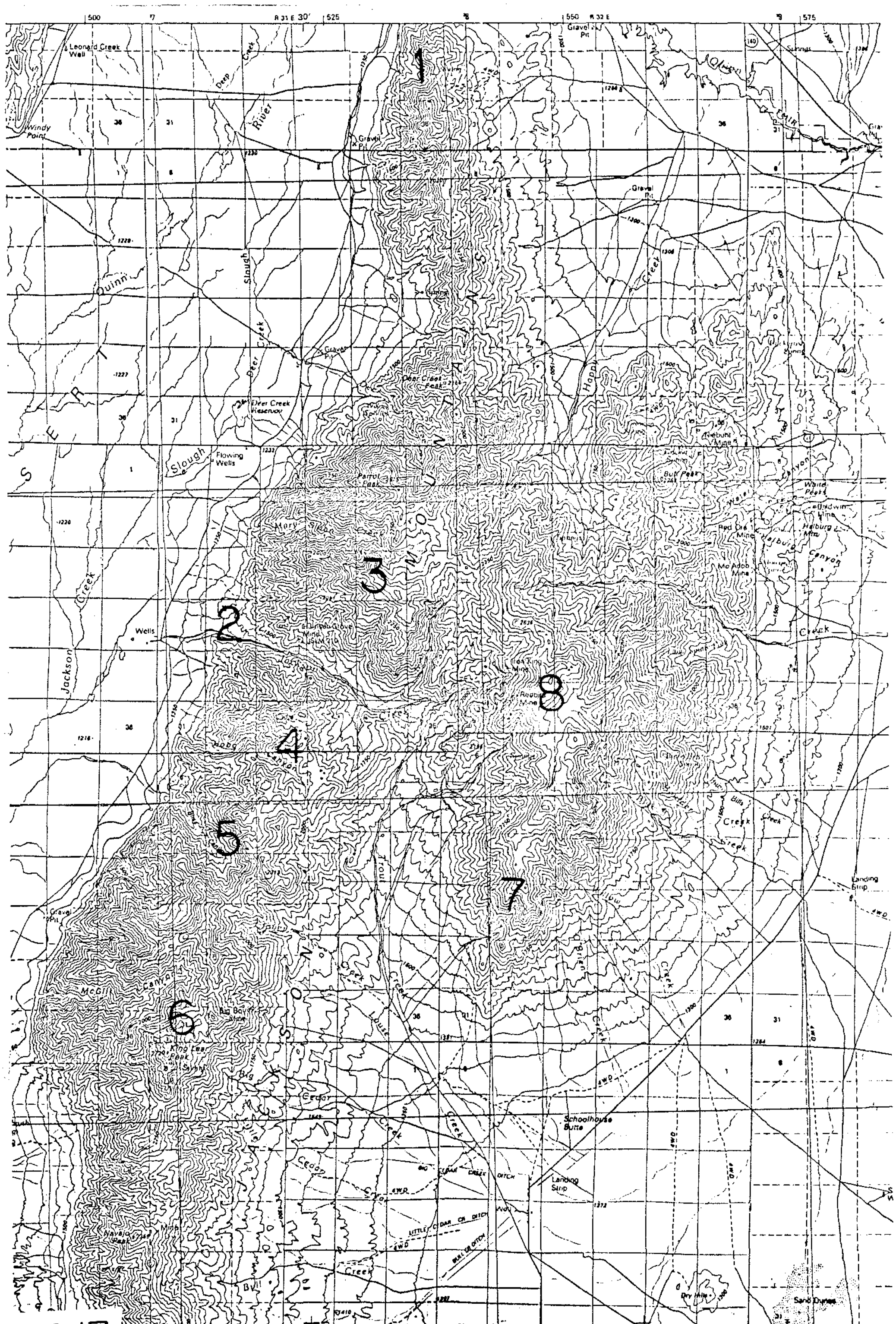


FIG.17

Map of the Jackson Mountains, NW Nevada showing the geographic areas discussed in the text for the Happy Creek Formation. These areas are: (1) the northwest tip; (2) the mouth of Jackson Creek; (3) north Jackson Creek; (4) Hobo Canyon; (5) Bliss Canyon; (6) King Lear Peak; (7) Trout Creek Spur; and (8) DeLong Peak.

interpretation of the important characteristics of the volcanic and sedimentary facies for the strata of that area. A final section will be concerned with the inferred regional depositional and volcanic, stratigraphic and structural framework, and the evolution and paleogeography of the Happy Creek Formation in the Jackson Mountains as a whole. Later chapters will discuss the geochemistry, geochronology, regional correlation and tectonic evolution of the unit.

(1) Northwest tip of Jackson Mountains

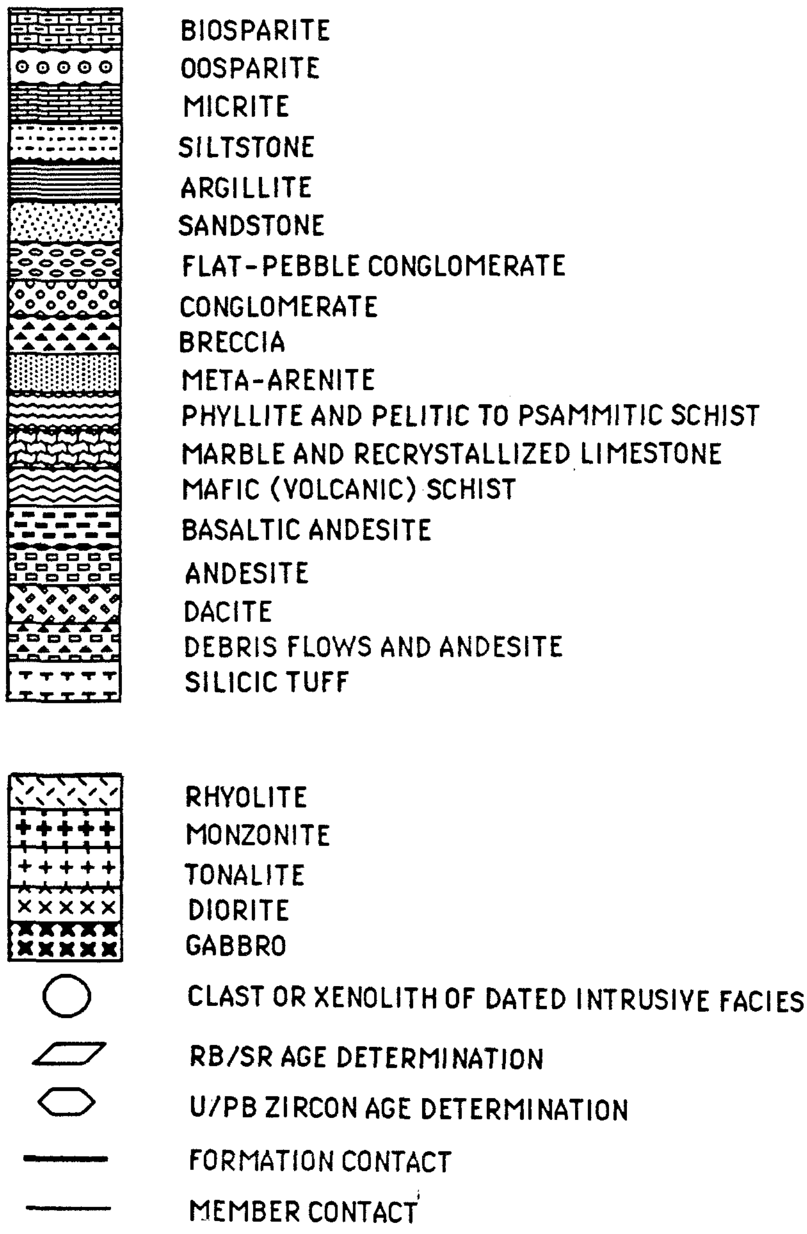
At the tip of the northwestern spur of the Jackson Mountains is outcrop of the Happy Creek Formation. This sequence of strata is intruded by a large cross-cutting granitic pluton (the Happy Creek pluton). As a result, units cannot physically be traced continuously from north to south, though the stratigraphic correlation is good. This area is not thrust faulted or metamorphosed (except right at the contact with the pluton), but has been internally cut by high-angle faults. The beds are unconformably overlain by Tertiary basalt flows and sediments; no other contacts are present. See fig. 18, 19 for the stratigraphic section. Though the section of the Happy Creek Formation shown was not measured because of difficult topography, the distances involved and the frequent interruption by faults and intrusions, they still record the stratigraphy satisfactorily as structural attitude and contact placement were both generally well determined in the field.

<u>Stratigraphy</u> The section of Happy Creek Formation in this area is about 1930 m thick. Top and bottom contacts are not exposed. The basal 510 m

FIG. 18

LITHOLOGIC SYMBOLS USED IN STRATIGRAPHIC COLUMNS

HAPPY CREEK FORMATION



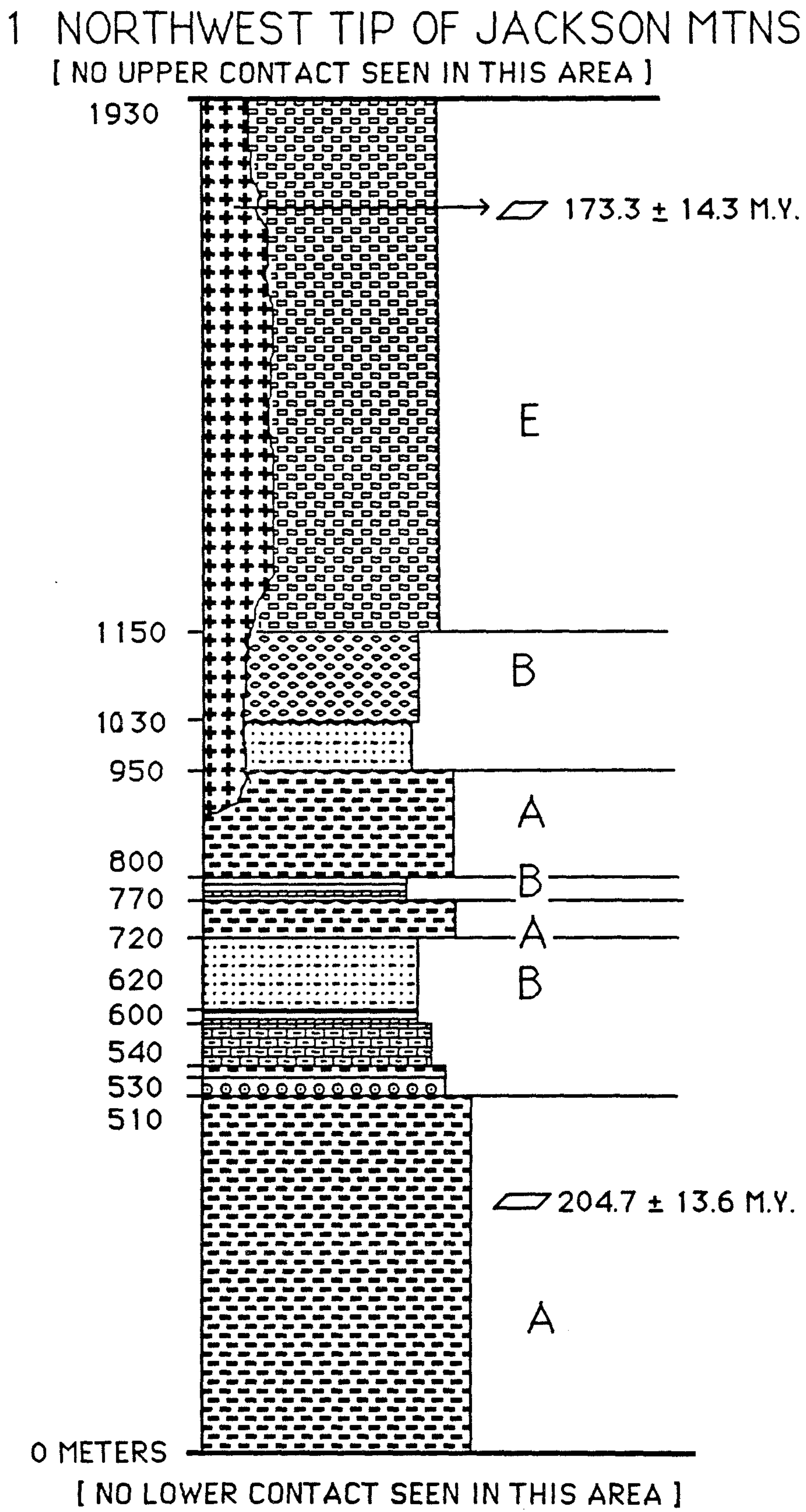

1 NORTHWEST TIP OF JACKSON MTNS
[NO UPPER CONTACT SEEN IN THIS AREA]
1930
173.3 ± 14.3 M.Y.
E
1150
B
1030
950
A
800
B
770
720
A
620
B
600
540
530
510
204.7 ± 13.6 M.Y.
A
0 METERS
[NO LOWER CONTACT SEEN IN THIS AREA]

FIG. 19

consists of dark green augite $\pm$ plagioclase phyric basaltic andesite, with abundant augite phenocrysts commonly up to 1 cm in size in a fine-grained groundmass. The flows are highly amygdular, with cavities up to 1 cm or more in size. Interbedded with the flows are flow breccias with angular clasts of the basaltic andesite; the interstices between the clasts have either a volcanic matrix or grey coarse calcite spar cement. The matrix to the augite porphyry flows contains mm scale plagioclase laths in an intersertal texture; the groundmass has been recrystallized to randomly oriented microlites of plagioclase, quartz, chlorite and mafic granules, and was originally quite fine-grained to glassy. The plagioclase phenocrysts have been extensively replaced by plates of a white mica (sericite?). The augite crystals are large and equant and display a disequilibrium relationship with the groundmass, and locally have reaction rims of very-fine grained brown material (these coronae are probably composed of hornblende).

Overlying the augite porphyry flows is light grey, thickly bedded limestone. This facies consists of 20 m of recrystallized oosparite at the bottom and about 60 m of biosparite at the top; there is a 10 m thick augite porphyry flow in between. The biosparite is a crinoidal grainstone, with round to pentagonal and star-shaped crinoid columnals, well-sorted, abraded and disaggregated, and can be cross-bedded. No micrite matrix was seen in these lithologies. The skeletal grains in the biosparite are from crinoids and calcareous algae with coarse granular calcite spar cement and sutured boundaries where grains touch, and in thin-section can be sheared and broken with micritic alteration along the fractures. Thin green siltstone layers are found between the thick limestone beds. The andesites and carbonates display a peperitic relationship, i.e., the sedimentary strata immediately underneath a flow are observed to be convolutely folded and ductilely deformed before lithification of the sediment. They also have angular volcanic glass debris and shards and microlites (highly chloritized and devitrified, and with reaction rims) intimately intermixed with the carbonate grains in a zone adjacent to the flow. This zone also exhibits pillow forms of flow rock encased in ductilely deformed carbonate strata. Some of the individual carbonate grains were deformed during this event. Also, patches of carbonate sediment were caught up by and deformed and baked within and between the flows. Dikes of the augite porphyry andesite within the carbonate strata have similar characteristics associated with them.

The carbonate strata are overlain by a 20 m interval of thinly laminated to cross-laminated arenaceous micrite, micrite and black sandstone (angular black grains in a light, calcareous matrix). Thin stromatolitic mats are present, characterized by crinkly, laterally continuous very thin laminae. This is overlain by 100 m of silicified grey fine-grained sandstone, siltstone, and argillite, thinly and evenly bedded to cross-stratified (low-angle). There are also thin beds of green volcanic lithic arenite. The siltstone (which can have herring-bone cross-lamination - i.e., bimodal) is entirely composed of very angular, very fine-grained and poorly sorted volcanic rock fragments, devitrified glass and igneous microlites in a muddy matrix. The microlites include quartz, feldspar and a mafic phase. The provenance is entirely volcanic; no terrigenous component is present. Bioturbation is also present in these facies.

Overlying this sequence is about 50 m of augite-phyric andesite flow rock, and then more arenaceous micrites and siltstones (30 m), then 150 m of augite porphyry identical to that below. Next in the stratigraphic sequence is 80 m of green to grey laminated to thickly bedded silicified argillite, siltstone and fine-grained sandstone. This coarsens upwards to a thick (120 m) interval of intraclastic conglomerate, with a sandy matrix and well-sorted and rounded ellipsoidal flat pebbles up to 10 cm in size. The sand matrix is clean and well sorted, and there is clast-support with imbricate fabric (the pebbles are all parallel to one another and imbricated at a very low angle or parallel to bedding). The pebbles are green, aphanitic and quite uniform in nature and are composed of very fine-grained volcanic rock. This facies coarsens upwards at the top to a poorly sorted and subangular conglomerate with the same clast type. Overlying these conglomerates are grey plagioclase ± hornblende andesite porphyry flow rocks, at least 780 m thick.

Green andesite dikes very similar in lithology to the andesite flows are found in the upper part of the section, and may have been part of a feeder system to the flows. Commonly they have soft-sediment deformation and peperite textures associated with them, hence intrusion probably took place while the sediments were unlithified. There are also cross-cutting granitic dikes from the pluton to the south, truncating all the other lithologies and the high angle faults.

The basal augite porphyry flow sequence is placed in member A. The interfingering sedimentary facies (biosparites, siltstones and conglomerates) are put in member B. The plagioclase ± hornblende andesite flows above are assigned to member E.

Age Augite porphyry flows identical to those in the basal part of the section from the southwestern Pine Forest Range several miles due west (see chapter 11) were dated by Rb/Sr isochrons at 204.7±13.6 m.y. (R. Kistler, pers. comm., 1988). The two sections correlate well enough that extrapolation of that age data to this area is highly reasonable.

The limestones were sampled for conodonts, with no success (D. Clark, written communication, 1986). In addition, two large cross-cutting monzonite (the Happy Creek pluton; KJd on the state map - Stewart and Carlson, 1978) and diorite (the Parrot Peak pluton; Mzgr on the state map) plutons to the south are colinear on an isochron at 173.3±14.3 on a Rb/Sr plot (R. Kistler, written communication, 1988). A maximum age of between latest Triassic to Early Jurassic (Norian to Pliensbachian) and the Middle Jurassic to early Late Jurassic (Aalenian to Oxfordian) is inferred for this section.

Depositional Environment A fair amount can be inferred from the nature of the augite porphyry flows of member A. The presence of abundant and large amygdules indicates extrusion subaerially or subaqueously at depths of several hundred m or less (Moore, 1965; Jones, 1969). The textures (porphyry with a very fine-grained to glassy intersertal to hyalopilitic groundmass) are characteristic of extruded lava flows. The associated autobreccias, which have either a volcanic matrix or calcite cement, are also flow-related, and could have formed in flow-front, top-of-flow and flow-foot settings. The calcite cement - bladed or equant mosaic and coarse and sparry - is most characteristic of the freshwater to mixed phreatic zone (below the water table), which agrees with the syntaxial calcite overgrowths on the echinoderm fragments (Scoffin, 1987). The nature of these flows is most compatible with subaerial to shallow submarine origin as thick, blocky, somewhat viscous aa or blocky flows. Less viscous pahoehoe flows (not present in the section) would be thinner and might have associated lava tubes and pillows, while thick, columnar-jointed horizons would be more characteristic of massive, flood occurrences.

The finer-grained clastic sediment in the lower occurrence of member B is highly indicative of a very shallow subtidal to intertidal marine environment. The bimodal herringbone cross-lamination in the siltstones, for example, is characteristic of rapid sedimentation in a high-energy setting with frequent and regular reversal of current - i.e., a tidal flat setting (Reineck and Singh, 1980). The stromatolitic mat (laterally linked hemispheroid) was formed in a supratidal to intertidal environment, while the presence of bioturbation would indicate normal marine waters. The siltstones and sandstones, with good sorting but angular particles and compositionally immature (as they consist entirely of glassy to very fine-grained volcanic debris) are probably hyaloclastite debris reworked locally in the tidal to subtidal environments. Such hyaloclastite particles are formed as lava flows encounter a body of water, spalling and fragmenting off glassy particles as the surface of the flow is quenched. The micritic and marly facies probably also represent intertidal to subtidal facies. The overall setting for this set of facies could very well have been in a lagoon between a carbonate barrier bar and a basaltic andesite stratovolcano (see next paragraph).

The thick-bedded carbonate bioherms, composed of crinoidal and calcareous algal skeletal particles, were deposited in a high-energy environment (good sorting, cross-bedding, with abraded and broken grains and no micrite mud). An offshore barrier bar or very shallow submarine dune field to shoreface or even subaerial dune field origin would be possible. However, given the relatively coarse grain-size (making aeolian reworking difficult), the evidence that the sediments were quite wet (peperite textures) and the overall geometry, an offshore bar is most likely. The carbonate facies exhibits substantial lateral facies variation within the outcrop area, and thins and lenses out into the augite porphyries and the intertidal sediments locally.

In the upper strata of member B, the flat-pebble conglomerate is extremely diagnostic of a very high-energy beach or supratidal setting. Note the clast support, rounding and sorting of the pebbles, the sandy and well-sorted clean matrix, and very low-angle imbrication. The pebbles were derived from nearby flows and reworked and deposited in the surf zone as the flows interacted with the marine realm, and perhaps redeposited by storm events in the supratidal zone. The coarsening and decrease in rounding upwards at the top of the section was probably due to slight regression or progradation, caused by relative sea-level drop or by an increased influx of volcanic flows and debris, or even by a decrease in energy.

The peperite textures seen throughout the middle part of the section record the contemporaneity of deposition of the very shallow to supratidal sedimentary facies and the basaltic to andesitic volcanism of members A (the flows) and E (the dikes that also cross-cut and deform the rocks of member B before lithification). The carbonate and clastic facies were being deposited in a very active volcanic region, along the shoreline interface of marine- and volcanically-dominated facies domains.

The top of the section - the grey plagioclase $\pm$ hornblende andesites of member E - represent subaerial flows. No interbedded sediments were seen, and few diagnostic features. Pillows or other indicators of marine extrusion are lacking, although flow-related autobreccias are present.

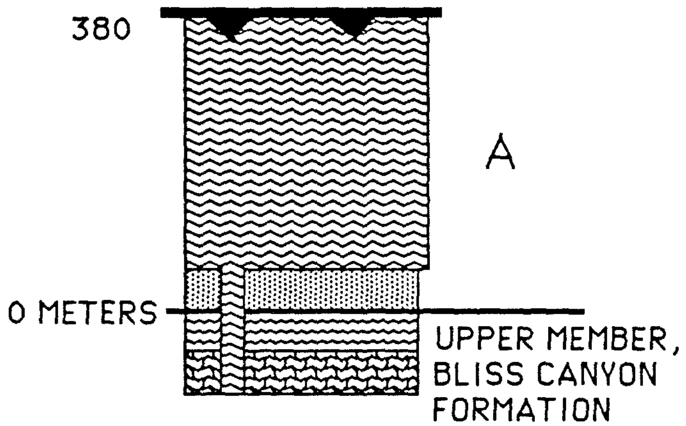
To summarize: The section records subaerial to shallow marine aa flows in the lower part. These are overlain by a carbonate barrier bar and lagoonal facies (the barrier bar probably lay to the present-day north of the lagoon), with occasional interruption by additional flows. The lagoon was strongly tidally dominated, with a proximal sediment source in the volcanic flows (including hyaloclastite debris - the flows were interacting with the marine realm) and some contribution from the carbonate facies further offshore (the marls and micrites). Higher in the section, the beach to supratidal flat-pebble conglomerates record overall regression or progradation, with the overlying andesite flows probably forming in a subaerial locale. Abundant peperitic relationships document synchronicity of volcanism and very shallow marine sedimentation.

(2) Mouth of Jackson Creek

Of all the outcrop areas, this one has been the most extensively affected by contact metamorphism related to several large adjacent and inferred subjacent plutons. The outcrop area includes the strata in several thrust sheets south of the mouth of Jackson Creek and north of Bliss Canyon. Virtually all of the original textures and stratigraphic relationships within the northern extent of these structural slices have been obscured and overprinted by later events. The metamorphic grade decreases rapidly to the south so that in the southern extent of the nappes more of the original features are preserved. See fig. 20.

Stratigraphy The depositional contact with the underlying Bliss Canyon Formation is identifiable in this area, though it may have been tectonically transposed. Green plagioclase porphyry andesite dikes of Happy Creek-affinity intrude and locally metasomatize the Bliss Canyon, which consists of marble, meta-quartzite and meta-conglomerate in this area. The Happy Creek Formation in contact with these metasediments of the Bliss Canyon in the northern part of the area consists of metamorphosed green volcanic arenites and is structurally overlain by dark green mafic phyllite and schist. The meta-arenites, where depositional textures are preserved, have thin bedding (1 to 2 cm) with low-angle cross-lamination to planar lamination and appear to have been moderately well-sorted. In thin-section there was no preservation of depositional texture, with the exception of some compositional banding inferred to be relict bedding, and of quartzose clots inferred to be siliceous pebble ghosts. The mafic schists are totally recrystallized; the only remnants of the original texture seen were large relict plagioclase grains. These latter may be preserved relatively intact, or in other cases are recrystallized around the edges with some replacement

FIG. 20
2 MOUTH OF JACKSON CREEK



by sericite in the cloudy interiors. It is possible that mafic schists and phyllites were volcanic epiclastics, with detrital plagioclase grains, or that the plagioclase were phenocrysts in a flow rock. These lithologies are included in member A. In the southern part of the area amygdular, mildly metamorphosed green volcanic flows of member A directly overlie the upper member of the Bliss Canyon.

Original thicknesses were difficult to estimate in this area, particularly as the section may have been tectonically thinned. The best estimate for the structural thickness of member A in the northern part of the area is about 240 m. The lower-grade metavolcanic interval to the south is about 380 m thick.

Age No age constraints exist on this section of the Happy Creek Formation.

Depositional Environment A shallow marine origin is most likely for the meta-arenite, given the parameters that were observed (the cross-laminated to paralell-laminated thinly bedded sorted sandstone with no finer lithologies), and the very shallow marine nature of the immediately underlying Bliss Canyon Formation.

Nothing much can be said of the overlying mafic schist horizon, except that it was at least plagioclase-bearing. Augite was probably also present but was completely metamorphically recrystallized. The amygdular lava flows to the south were probably extruded in a shallow marine to subaerial setting, given the lack of pillows and the size relative abundance of the amygdules. They could also have been augite-phyric originally, but the low-grade metamorphism obscured field identification of the mafic phenocryst population.

(3) North of Jackson Creek

The area described in this section extends from just north of Jackson Creek on the south (bounding area 4), northwards to the southern extent of the Happy Creek pluton. It is bounded on the east by the depositional contact with the King Lear and on the west by the east-vergent thrusts (above which lies area 2). The Happy Creek Formation in this area has been affected by syn-volcanic east-west trending high-angle faults. These faults possess an early phase with a substantial component of vertical slip, with north-side down overall along a complex fault system in the mutual headwaters of Happy Creek, Mary Sloan Creek and the north fork of Jackson Creek. Later, post-depositional faulting of the same geometry also cut the unit. The formation has, in addition, been affected by later west-vergent thrusting and folding, and has been intruded by several cogenetic plutons and by minor later bodies as well. Around these comagmatic intrusive centers contact metamorphism is locally significant. There has been extensive jasperization and iron mineralization along the syn-depositional high-angle faults during their initial phase of activity. The stratigraphy is distinct and mappable, and totals about 4300 m. Both the lower (with the Bliss Canyon) and upper (with the King Lear) depositional contacts are preserved. The former is conformable, while the latter varies in nature from conformable to buttress unconformity. See fig. 21 for a stratigraphic column.

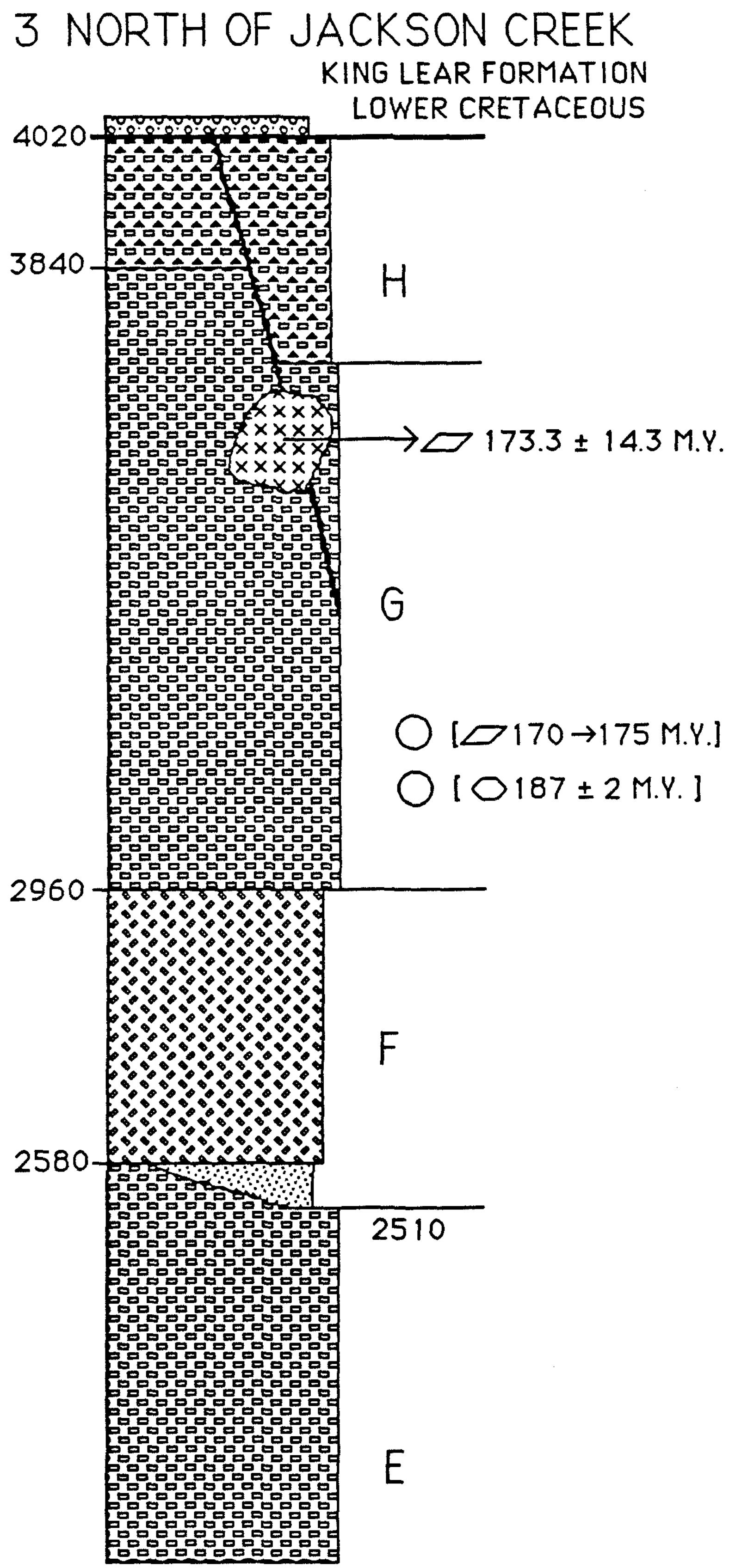
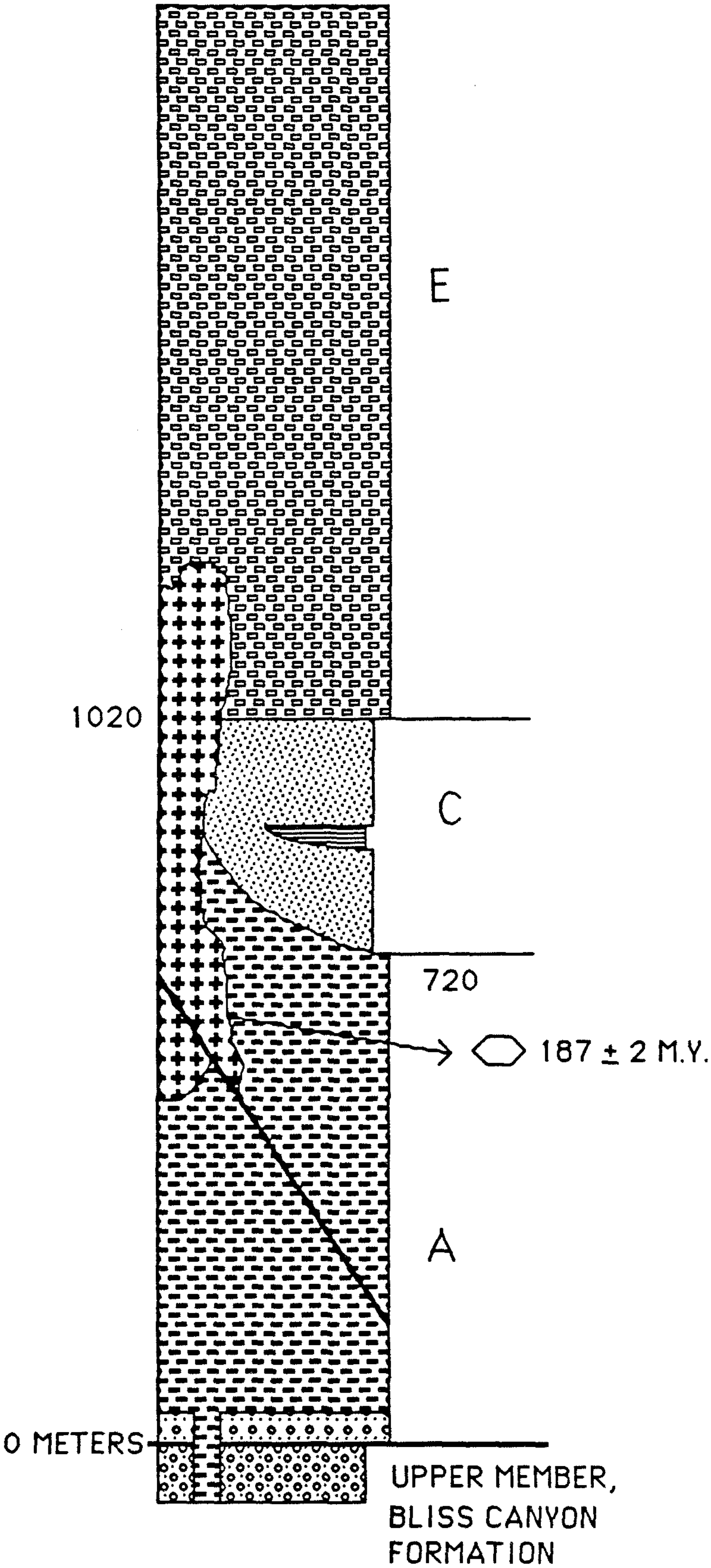

3 NORTH OF JACKSON CREEK
KING LEAR FORMATION
LOWER CRETACEOUS
4020
3840
H
173.3 ± 14.3 M.Y.
G
[170→175 M.Y.]
[187 ± 2 M.Y.]
2960
F
2580
2510
E

FIG.21



E
1020
C
720
187 ± 2 M.Y.
A
0 METERS
UPPER MEMBER, BLISS CANYON FORMATION

Stratigraphy Member A crops out near several large plutons and is correspondingly contact metamorphosed. The upper member of the Bliss Canyon Formation (rounded quartzose pebble conglomerate with cross-bedding, imbricate clasts and scouring and containing lenses of green volcanic arenite) is overlain by approximately 30 m of epiclastics at the base of member A of the Happy Creek Formation. This interval is composed of coarse-grained green volcanic arenite and conglomerate. The beds are thick with normal grading. There are thin interbeds of light green sugary-textured calcsilicates (containing quartz, calcite and epidote), recrystallized light grey calcarenite and tuffaceous metachert. The epiclastics coarsen upwards, and are overlain by dark green augite and/or plagioclase porphyry basaltic andesite and andesite flows and flow breccias. Some pillows observed at the base of the sequence. Textures are trachytic, and the augite phenocrysts up to 6 mm in size. Augite abundance decreases upwards overall throughout the member, and hornblende is present in the upper part as well. The member is 720 m thick just north of the western part of Jackson Creek. The Bliss Canyon Formation in this area is intruded by similar green porphyritic andesitic dikes with augite, andesine and quartz in an intergranular matrix of feldspar and quartz. These are probably feeders to the flows.

The plagioclase in the flow facies is An_{30} to An_{40} (as measured on Carlsbad and albite twins), and so is andesine in composition. Apatite and quartz are common accessories, with chlorite, epidote and elbaite tourmaline growth overprinting the volcanic textures. The mafic phases are often completely replaced (by a combination of chlorite, an opaque, calcite and epidote, and rarely by hornblende). The groundmass is recrystallized (and contains chlorite, opaque, plagioclase, quartz and epidote) and the plagioclase phenocrysts extensively sericitized. Veins of epidote, quartz, chlorite and calcite are common. Original volcanic textures are, however, well-preserved. The groundmass is felty to trachytic, and when it is compared to the augite and andesine phenocryst population the overall texture is highly bimodal.

The tuffaceous meta-chert is thinly laminated to homogenous with angular plagioclase and mafic fragments floating in a cryptocrystalline cherty matrix. The mafics have been replaced by epidote and chlorite, and the plagioclase sericitized and silicified. Breccias associated with the volcanics have angular andesite xenoliths in an andesitic matrix of similar composition. Higher in the member, hornblende, chlorite and quartz appear as important components in the volcanic mineral assemblage. The crystallization sequence is plagioclase hornblende + chlorite quartz augite. The texture is intergranular texture, and oligoclase is the plagioclase (An_{28}).

Overlying member A in the south, but disappearing northwards is a lens of light green volcanic arkose and phyllite. The sandstone facies is thinly well-bedded with cross-stratification and is locally conglomeratic, with andesitic volcanic clasts. The phyllite is very fine-grained and homogenous, aside from cleavage, and is a minor part of the epiclastic horizon. The unit is 310 m thick at its most extensive, just north of the western part of Jackson Creek, and belongs to member C. This unit is better developed to the south.

These epiclastics are overlain in turn by another volcanic succession, which is distinct from member A. This member, E, is 1490 m thick and is made up of massive fine-grained green to grey plagioclase-phyric andesite flows and breccias. Amygdules up to 6 to 8 cm in diameter are present in the flows in this interval, and have been filled by radial growths of chlorite. Locally in the upper half of the member the flows are oxidized and reddened on their tops in zones up to several m thick. The plagioclase is substantially more calcic, with An_{56} (labradorite). Augite was observed as granules in the groundmass, but is quite rare as a phenocryst phase in this member. Sphene is also present. Scattered diorite dikes of the Early Mesozoic Intrusive Suite cut the section.

The next unit, member F, has at its base a lens of green massive volcanic conglomerate with clast-support and a poorly sorted and angular volcanic arenite matrix. This lens is 70 m thick, and thins and disappears to the north and south. The superjacent bulk of the member is light grey or bluish-grey dacite. The dacite is plagioclase- and hornblende-phyric, and outcrops as flows and as associated breccias with very coarse and angular clasts of the dacite in a light-colored glassy matrix. Some of these breccias have reaction rims on the fragments, implying disequilibrium with the matrix. The flows are very thick and massive, with reddened and oxidized zones at their tops. The dacite part of the member is 590 m thick.

Petrographic examination found that the dacite flows have plagioclase + quartz + biotite and/or hornblende $\pm$ alkali feldspar (sanidine) as major phases, and zircon, opaques and apatite as minor phases. Barite was found during mineral separation efforts to extract zircon. Sphene, epidote and chlorite have replaced the original mineral assemblage in dispersed patches. The observed crystallization sequence in one flow was plagioclase hornblende biotite alkali feldspar quartz. The plagioclase is slightly zoned from An_{32} in the cores to An_{20} at the rims. The alkali feldspar is sanidine and locally has cryptoperthitic and myrmekitic textures. The texture of the dacite flows is granitic (subhedral granular) in the interiors, to the more common bimodal porphyritic fabric with a fine-grained intergranular matrix of feldspar and quartz. Some of the flows are also pilotaxitic to trachytic, reflecting varying degrees of flow-alignment of the phenocrysts. The dacites are also notably rich in Fe-oxide opaque phases. Less commonly, flows have plagioclase in the labradorite composition field with An_{60}, instead of andesine/oligoclase, and also have augite as one of the mafic phases.

The dacite member is succeeded by another andesitic interval, member G. This member, 880 m thick in the north fork of Jackson Creek, is characterized by green to grey andesite flows, associated breccias and abundant volcanic diamictites. The flows are tabular and well-bedded, between 3 and 22 m in thickness, and almost always have oxidized and reddened zones between 1 and 6 m thick at their tops. The breccias, which commonly occur as a framework at the bases of the flows, make up the entire flow body, or are just scattered through, have angular clasts of the andesite flow facies in a very fine grained andesitic matrix. The clast population includes both oxidized and unoxidized clasts, implying oxidation took place in between eruptive events. The diamictites contain a variety of lithologies, including the angular andesite clasts and the comagmatic intrusives of the Early Mesozoic Intrusive Suite (tonalite, granodiorite and particularly diorites). The clasts are supported by a siliceous, aphanitic matrix. The flows are plagioclase-phyric porphyries with random to trachytic fabrics or are aphyric. They are amygdular, particularly at their tops. The amygdules are filled by calcite, quartz and chlorite, and are irregular to ellipsoidal voids or vertical pipes.

In thin section, the assemblage plagioclase $\pm$ hornblende or augite $\pm$ quartz + opaque is present in the andesites. The plagioclase is andesine (An_{35}) and in some cases has thin oscillatory zoning. The groundmass is a cloudy brown to clear green glass with granules of opaques and microlites of feldspar, with the phenocrysts randomly oriented (hyalopilitic) to semiparallel (pilotaxitic). The andesites are highly bimodal (porphyritic) in grain size distribution. The plagioclase is somewhat sericitized, the mafics are in some instances replaced by chlorite, calcite and epidote, and there is some recrystallization of the groundmass, but the extent of this process is noticeably less than in lower members of the formation. Augite is substantially less common then in member A, and not evident in the field. Some flows have xenoliths of slightly different volcanic lithologies; the degree of alteration and replacement in these xenoliths was observed to be substantially greater than in the host rock.

Small intercalated occurrences of epiclastic sediments are present in member G. They range from lenticules several m thick of cross-bedded dark grey volcanic arkose and conglomerate with intraformational volcanic and intrusive (granodioritic to dioritic) clasts to one lens north of Parrot Peak 60 m thick and 1.8 km wide in outcrop. This body is made up of massively-bedded reddish-brown coarse-grained volcanic conglomerate.

Red, well-bedded sedimentary breccias, with support by a wacke matrix, are also found at several places high in the section north of Parrot Peak, and are associated with red cross-laminated arkose lenses.

The basal King Lear in the fault-bounded high south of the high-angle fault complex mentioned earlier rests directly on member G, and consists of a very thin basal red conglomerate with volcanic and plutonic clasts and overlying green chert-pebble conglomerates. The uppermost member G is highly weathered and reddened, with caliche carbonate veins and some incised channels filled with King Lear boulder conglomerate.

Member G and all underlying stratigraphic units are cut by the family of east-west high-angle faults. The paleotopography caused by this system indicates that a substantial amount of vertical slip took place, and stratigraphic relationships show that Happy Creek volcanism was ongoing during the first phase of fault activity. Member H of the Happy Creek Formation fills and even overlaps some smaller graben structures (which cut member G) in the headwaters of Happy, Mary Sloan and Jackson Creeks and was deposited on the floor of the large fault-bounded basin in the northern half of area 3. It is missing on the structural high in the southern part of area 3. The faults are characterized by jasperization and by hematite and magnetite mineralization along their surfaces, and the strata near and within the grabens are similarly affected. Sedimentary debris within member H contains already mineralized clasts, and flows contain Fe-mineralized xenoliths. Member H (and perhaps also the upper part of member G, which could be laterally equivalent in places distant from the faulting) was thus laid down during the initial activity of these faults. The faults also continued this initial phase of movement during the deposition of the basal King Lear Formation (see appropriate section).

Member H is 180 m thick in the main basin in the northern part of the area, and could be 300 m thick or more in some of the smaller basins. This latter value is only a rough estimate because of the complex structural relationships. The northern section is characterized by red andesitic flows with interbedded distinct massive and homogenous red breccia beds. The breccias exhibit poor sorting, clast- to matrix-support in a muddy to sandy matrix, and angular to subangular clasts (up to 1 m in size) of Happy Creek volcanic and Early Mesozoic Intrusive Suite facies. Member H pinches out further to the north. The lowermost King Lear is composed of similar breccias without the andesite flows. There is no evidence for a hiatus in deposition between H and the King Lear Formation, and in fact the transition is gradational.

The graben-fill sequences in the central, headwaters portion of area 3 have a more complex stratigraphy. The lithologies present include red plagioclase porphyry scoriaceous andesite flows and related breccias, red fine- to very coarse-grained epiclastic sediments, and green volcanic arenite. The red andesites have the assemblage plagioclase $\pm$ hornblende $\pm$ augite $\pm$ chlorite, with accessory apatite and no quartz. The plagioclase is andesine with An_{33} to An_{36} and is in some cases highly sericitized. The texture is porphyritic with weakly aligned (pilotaxitic) to parallel (trachytic) arrangement of the phenocrysts in a felted plagioclase-rich matrix with some glass in the interstices. Some of the flows are so completely replaced by hematite that no vestiges of the original textures survives except for phenocrysts. Other flows contain xenoliths that have already undergone such replacement before their entrainment. Some of the breccias are directly flow-related facies containing angular fragments in a matrix of the same lithology. Some of the volcanic breccias are quite different, with angular fragments and brittlely broken grains in a foliated to vesicular glassy matrix. These volcanic facies have commonly undergone replacment by epidote, sphene, quartz, calcite and chlorite.

There are additional breccias and conglomerates with angular to subrounded intraformational and comagmatic intrusive clasts (including tonalite and Fe-mineralized andesite) up to several m in size in a red volcanic and arkosic wacke matrix, commonly in matrix-support. The clasts have already undergone the Fe-mineralization, and the matrix is highly hematitic as well with detrital plagioclase grains and volcanic rock fragments, themselves in a Fe-rich muddy matrix. Sorting is extremely poor, and all detrital fragments are highly angular. Beds of similar red mudstone and volcanic arkosic wacke are also present in the graben fill. In addition, there are interstratified green coarse-grained volcanic arkoses, clean but poorly sorted and angular. These sedimentary facies make up a major portion of the graben-fill sequence. Bedding in the grabens is very highly laterally and vertically variable.

Age The base of the section (members A, C and E) is intruded by a monzonitic pluton (the Harrison Grove pluton) with a U/Pb zircon age of 187$\pm$2 m.y. This pluton is cut by the high-angle faults. In addition, the Happy Creek and Parrot Peak plutons are colinear on a $^{87}Sr/^{86}Sr$ vs. $^{87}Rb/^{86}Sr$ diagram at 173.3$\pm$14.3 m.y. (R. Kistler, written communication, 1988). The Happy Creek and Parrot Peak and Happy Creek intrusives also cut members A, C and E, plus members F and lower G, as well as some of the first phase of east-west trending high-angle faults. Xenoliths and fragments identical to the Harrison Grove pluton are found in members G and H in this area. Zircon extraction was attempted on the two different dacite flows of member F, but with no success. No fossils were found in the Happy Creek Formation anywhere in the Jackson Mountains.

The underlying Bliss Canyon Formation in this area has no direct biostratigraphic constraints. Pagiophyllum found in the overlying King Lear Formation has been assigned the range of Triassic to as young as Lower Cretaceous by S. Ash (pers. comm., 1986).

The Happy Creek Formation in area 3 is Lower Jurassic or earlier at the base, as members A, C and E are intruded by a pluton straddling the Early-Middle Jurassic boundary. Members F and lower G are Middle to early Late Jurassic or older, as they are intruded by a body of that age. The early generation of high-angle faults is constrained to the interval of middle to early Late Jurassic by cross-cutting relationships with the intrusives. The members G and H must have been deposited after the intrusion of the Harrison Grove pluton, and so are Middle Jurassic or younger (but older than Early Cretaceous, which is the minimum age of the overlying King Lear Formation in this area). H, however, was deposited in active basins whose bounding faults are cut by the younger set of plutons, and so must be as old or older than these plutons. Member G must also be as old as or younger than 170 to 175 m.y., as that is the preliminary U/Pb zircon date obtained on diorite sill to the east (and identical to the xenoliths found in G).

In summary, the best estimate for the range of the Happy Creek in this area is then Lower Jurassic or older (members A to F) to late Middle Jurassic (members G and H).

Depositional Environments The augite-phyric flows and flow breccias are, at the base of the member A, interbedded with sedimentary lithologies. These facies include the calcarenitic marble (probably a recrystallized biosparite), calcsilicate (with a marly or a quartzose carbonate protolith), and tuffaceous chert (whose thin lamination might be due to water-deposition). All of these facies are inferred to be of a marine origin. The basal green volcanic arenite and conglomerate of member A, thick-bedded, lacking fines and with normal grading and no observed internal sedimentary structures, is best assigned to a fluvial setting where such normal grading probably represents a decrease in transport velocity and channel abandonment (Reineck and Singh, 1980). A turbidite origin seems highly unlikely, as there are no indicators such as were described from the lower member of the Bliss Canyon Formation. Supporting this is the fact that the underlying upper member of the Bliss Canyon is very shallow marine, as is member A to the north in area 1.

The augite-phyric basaltic andesite and associated breccias of member A were deposited as blocky, thick volcanic flows, judging by the characteristic textures of the bodies. These textures include the highly bimodal grain size distribution with a very-fine grained groundmass, and the trachytic flow-alignment of the phenocrysts. The breccias, with angular monolithologic clasts in a similar andesitic matrix, are flow-related as well. The observation of pillows at one locality does point to extrusion of at least part of the section subaqueously (whether marine or nonmarine is not clear). The comagmatic andesite dikes in the subjacent upper member of the Bliss Canyon are feeder dikes to member A.

The depositional origin of Member C is not very clear. The combination of thin, well-bedded and cross-stratified sandstone and conglomerate with shaly lenses seems most likely to have originated as braided channel sand flat deposits (and not braided channel scour and fill, as such beds would be thicker and have erosive bottoms - Reineck and Singh, 1980). Such deposits would require steep slopes and a coarse ediment load (implying proximity to the sourceland), such as in a lower middle to distal alluvial fan or alluvial plain setting. The detritus within this member is entirely volcanic and intraformational in origin.

Member E, amygdular andesite and related breccias, is interpreted as another flow sequence. Evidence for subaerial eruption is provided by the oxidation and weathering zone observed on many flow tops in the upper half of the unit in particular, and by the size range of the amygdules (6 to 8 cm), which is much greater than that observed for submarine flows of any significant depth (Moore, 1965; Jones, 1969; Williams and MacBirney, 1979; Cas and Wright, 1987).

The epiclastic interval at the base of member F is interpreted in a fashion similar to member C, as braided stream deposits (though in a different setting). The clast support and sandy matrix imply fluvial sedimentation, while the angularity of the detritus, the extreme coarseness, the lack of fines and the poor sorting point to extreme proximity to the source; the volcanic strata above and below are subaerial as well. The body appears to be large channel feature cut into member E and filled with the epiclastics. A braided stream system is preferred to a meandering river because of the probable very high sediment load in the system. The sediments appear to be andesitic and not dacitic, implying deposition before the overlying dacitic phase of volcanism had begun.

The origin of member F, thick dacite flows and flow breccias, is judged from the observed textures. These include that of the pilotaxitic to trachytic and highly porphyritic flows, and the monolithologic breccias in similar aphanitic matrix. Quartz, alkali feldspar, hornblende and biotite are significant phases in these flows in addition to the usual andesine plagioclase, and they are unusually rich in Fe-oxide phases. The mineralogy is the culmination of a trend started in members A and E of increasing FeO*/MgO ratio and silica and alkali content, as seen in the increasing quartz and Fe-oxide content, the appearance of alkali feldspar and in the transition of the mafic phase mineralogy from augite   augite + hornblende   hornblende   hornblende + biotite   biotite. Subaerial eruption is again indicated by the weathering of the flow tops.

The overlying andesitic succession of member G appears to contain a reversal of this evolutionary trend, implying perhaps a new phase of magmatism. The mineral assemblage is much more like that of member E (lacking quartz and alkali feldspar and with augite and hornblende present). The flows are trachytic andesite with flow-foot and basal breccias and pipe amygdules at the tops. These features are characteristic of blocky or aa flows (Cas and Wright, 1987). Extrusion was subaerial, as almost every flow has a subaerially weathered top. The diamictites are lahars - volcanic debris-flows with polylithologic but intraformational clasts supported in a siliceous aphanitic muddy matrix (though whether hot or cold is not known). The observed alteration of the flows (sericitization of plagioclase, replacement of mafics by chlorite, epidote and calcite) took place very early on, as some flow xenoliths underwent this process before entrainment in later flows. The presence of xenoliths in the flows and clasts in the lahars of plutonic facies could be explained by their entrainment in the feeder system to member G, or by their exposure in the area and reworking into the volcanic facies; in either case those particular intrusives had already been emplaced.

The large lens of volcanic conglomerate north of Parrot Peak represents a canyon incised into the volcanic succession, filled by fluvial gravels and conglomerate, and covered by more flows. The thin cross-bedded arenite occurrences could be similar smaller fluvial channel-fill, sheet-flood deposits on a sloping volcanic surface, or even aeolian deposits. The matrix-support red breccias in the upper portions of the member North of Parrot Peak are debris flows, and the red cross-laminated arenite are small fluvial or sheet-flood lenses; these may actually be distal equivalents of member H, but do not form a distinctive mappable horizon.

Member H was closely related depositionally to the high-angle faults. In the northern part of the area the member is composed of red subaerial andesite flows and flow-breccias, and red sedimentary breccias with very poor sorting and some very large clasts, clast- and matrix-support, and wacke to sandy matrix. The sedimentary breccias are interpreted as very proximal debris flows coming off the fault scarp caused by the high-angle structures, in a talus slope to uppermost alluvial fan setting. All the clasts are from within the Happy Creek Formation or the related intrusive bodies. The unit tongues out to the north, away from the fault, until the King Lear directly overlies member G.

In the marginal block-faulted system just to the south, H fills the grabens with the subaerial andesitic flow facies and debris flows observed to the north and with additional facies. The breccias with broken grains and clasts in a foliated glassy groundmass are tuff breccias, caused by phreatic explosive events. Other sedimentary facies include conglomeratic wackes with matrix-support and red volcanic and arkosic mudstones, all very immature (very poor sorting, dirty, and very angular) and probably upper alluvial fan or talus in origin, with deposition by debris-flow or sheet-flood processes. The green arkoses, clean and with better sorting, are probably stream channel facies. All facies are laterally and vertically discontinuous, as would be expected in this structural and depositional environment. The Fe-mineralization was caused by epithermal systems along the faults during their activity and during the filling of the basins, and affected strata near those faults as well as in the graben-fill.

(4) Hobo Canyon

This outcrop area includes the occurrences of the Happy Creek Formation on the western side of the Jacksons below and east of the thrusts that bound area 2, south of Jackson Creek and area 3, and north of the ridge just south of Hobo Canyon and area 5. Metamorphism is not significant, there are no cross-cutting plutons, structure was relatively simple, and the stratigraphy was distinctive and mappable. See fig. 22.

Stratigraphy The stratigraphic contact with the Bliss Canyon Formation is not present (due to an overthrust by the east-vergent thrust system), but the contact with the overlying King Lear Formation is quite prominent. In this area the two formations are conformable (with andesite flows and coarse-grained red beds interstratified in the lowermost King Lear) to disconformable (bedding in both units parallel, but with weathering and caliche development evident in the uppermost Happy Creek at the unconformity). The contact is put at the base of the lowest clastic bed of the King Lear Formation.

In the south, the lowermost Happy Creek Formation is green andesite flows, stratigraphically below the green volcanic epiclastics, with a minimum thickness of 380 m. These volcanics are placed in member A, and the lower contact was not exposed. No augite-phyric lithologies were noted in this lower interval, which could be consistent with an origin in the upper part of A.

In the northern part of this area the lowermost exposed strata consist of 960 m of volcanic coarse-grained sediments. The lowermost 200 m is green coarse-grained arkosic volcanic arenite and wacke and volcanic conglomerate with andesite pebbles to boulders. The sequence is thick-bedded (5 to 10 m thick beds). Above 200 m some andesite flows are interbedded with volcanic wackes. Some thick matrix-support diamictite volcanic breccias with a mud matrix are also present. At about 600 m there is a thin bed of rhyolitic crystal tuff with broken plagioclase

4 HOBO CANYON

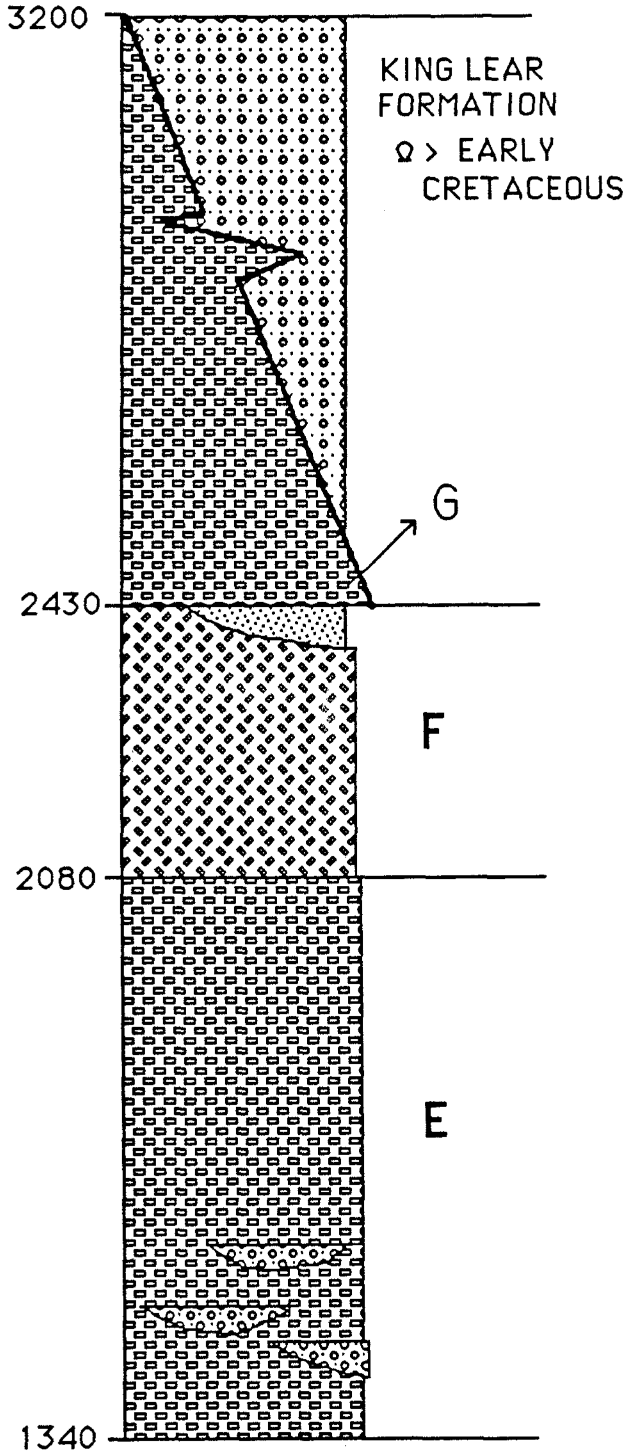
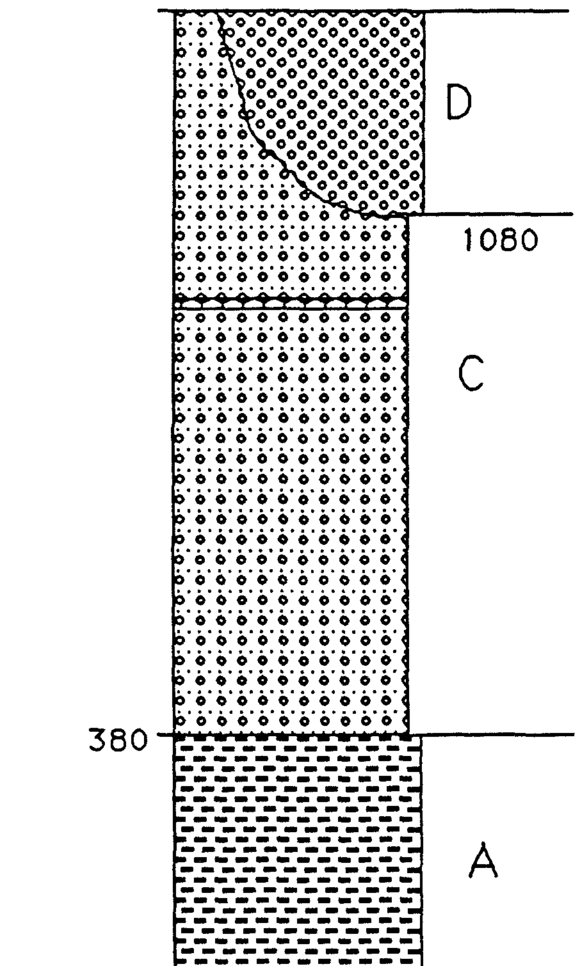


FIG.22



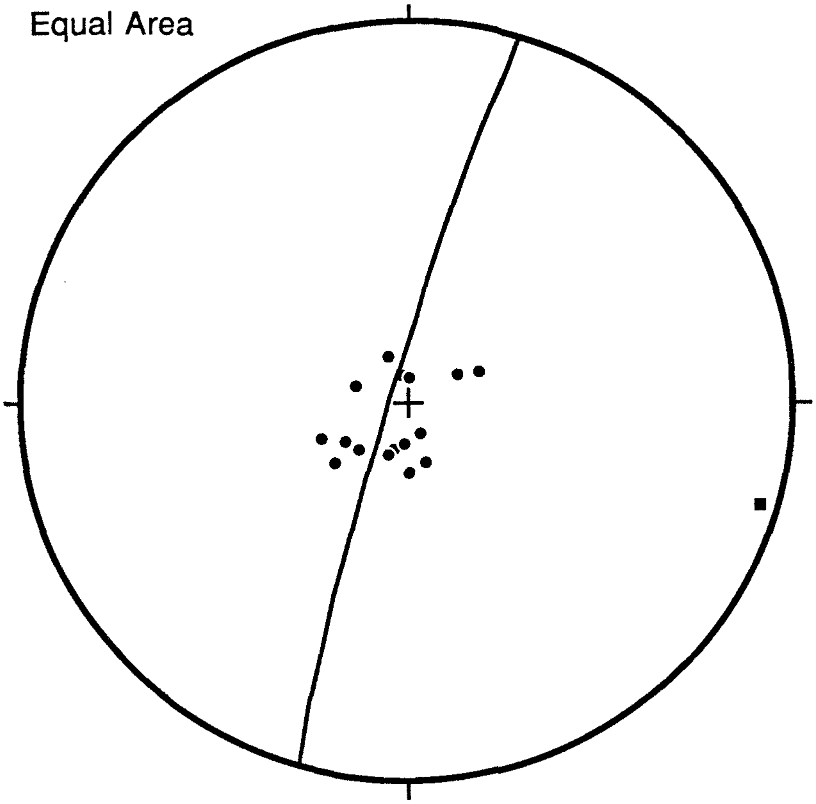
[BASE NOT PRESENT IN THIS AREA]

grains in a glassy groundmass that has been highly altered to calcite. Many of the epiclastics in the upper portion of this sequence also contain abundant reworked highly siliceous volcanic lithologies - tuff breccias, tuffaceous conglomerates, and tuffaceous sandstones are found at a level of about 800 to 900 m (to the north of and stratigraphically above the pinchout of the red bed lens discussed below). The tuffaceous clasts are white and rhyolitic, with glassy to flow-banded or lapilli tuff textures.

The dominant lithology throughout this entire epiclastic interval is green thick-bedded fine- to coarse-grained arkosic volcanic arenite and wacke with small lenses of volcanic pebble to cobble conglomerate. There are only rare andesitic flows. Fining-upwards sequences from cobble conglomerate to fine-grained arenite with thicknesses of 20 to 50 m were observed at several localities. The conglomerates commonly have a volcanic wacke matrix with clast-support of various volcanic lithologies (all recognizably Happy Creek or related intrusive phases). Clasts are often as large as 8 cm and can be boulder-sized. The wackes are green but weather orange, are coarse-grained, poorly sorted, and angular- extremely immature. This epiclastic interval is placed in member C.

Above this and interfingering with C (in the northern part, in the headwaters of Hobo and Christiorsson Canyons) is a very thick (260 m) and laterally extensive (2.7 km wide) lens filled with conglomeratic red beds. These red beds are very poorly sorted, subangular to angular, with clast-support and a bright red arkosic and volcanic arenite or wacke matrix. Hematitic cement gives the red color to the rocks, which contain volcanic rock fragments and detrital sericitized plagioclase, quartz and chloritized mafics. A secondary calcite cement fills in the pore spaces left by the hematite. The clasts (pebble to boulder) are entirely intraformational, consisting of andesitic, rhyolitic, and vein quartz lithologies; no lithologies, including intrusives, with a provenance uniquely outside the Happy Creek Formation are present. The siliceous volcanic clasts include a plagioclase phyric rhyolitic flow-breccia. The lens is composed of numerous scour-and-fill channels, with scoured and abrupt bases, normal grading, clast imbrication and trough and tabular cross-stratification. Rare interchannel red siltstones to fine-grained red arkosic volcanic arenite are preserved, in interchannel areas. A field estimation of the overall clast imbrication of pebbles in the lens indicated a source to the east. The trough cross-stratification surfaces were used in paleocurrent analysis, and the results indicate an east-west axis and paleocurrent direction (fig. 23), with paleoslope possibly dipping to the west. This lens on its northern margin is overlain by and lenses out into the green thick-bedded volcanic arenites of member C. In the central (thickest) and the southern portions of the lens it is directly succeeded by green plagioclase phyric andesite flows, with baking observed at the contact. The red bed lens is defined as member D.

To the south, around Hobo Canyon, the green coarse-grained volcanic epiclastics of member C are somewhat thinner (310 m) and finer-grained, characterized by thick-bedded fine- to coarse-grained arkosic and volcanic arenite and small (several m) lenses of green volcanic pebble conglomerate, with some green to red siltstone at the base. The overlying lens of red beds of member D is also of decreased thickness (170 m decreasing to zero as it lenses out to the south), and is somewhat



POLES TO CROSS-STRATIFICATION, MEMBER D,
HAPPY CREEK FORMATION

THE GREAT CIRCLE IS THE GIRDLE TO THE CROSS-BED POLES (197 86W), WITH A POLE AT 107 4 (FILLED BOX) - PEBBLE IMBRICATION INDICATES A SOURCE FROM THE EAST, SO THE PALEOCURRENT DIRECTION WAS FROM ABOUT 110.

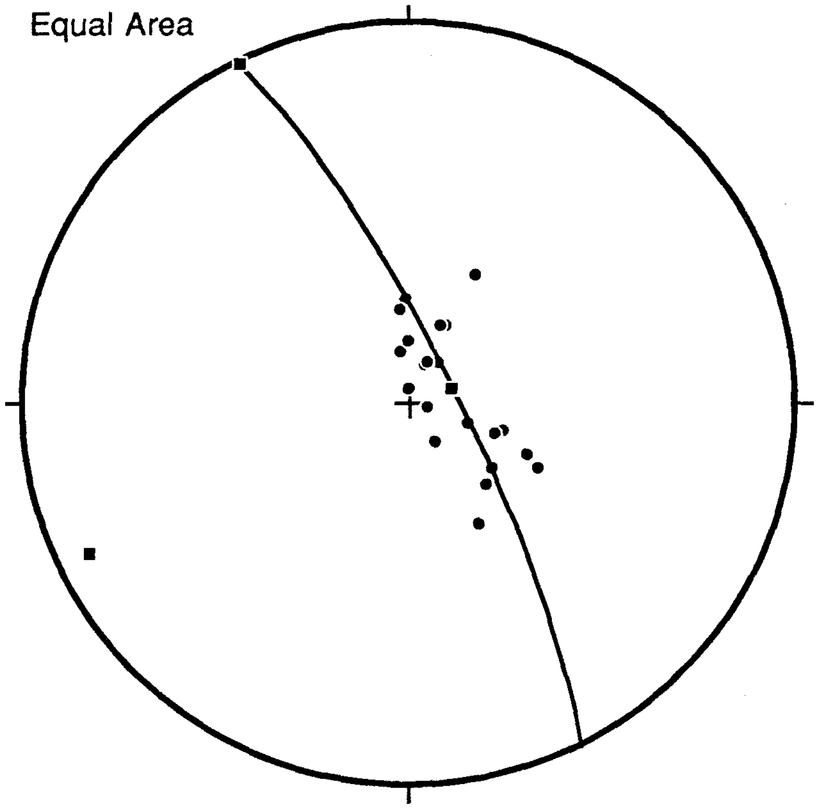
FIG, 23

finer than to the north, consisting of fine- to coarse-grained red arkosic volcanic arenite to moderately rounded red volcanic pebble conglomerate.

Overlying both members C and D is a horizon of green plagioclase phyric to aphyric green andesite flows. The flows are about 580 m thick in the north to about 735 m thick in the middle to 670 m to the south. The flows, where thickness could be determined, are about 10 m thick with finer grained tops and glassy bottoms, and some have scoriaceous textures. A few of the flows are hornblende or clinopyroxene phyric. There are also lahars - volcanic diamictites with volcanic clasts supported by a green siliceous matrix. These lithologies are part of member E. In addition, there are a number of small concave lenses of coarse-grained red beds identical to those of member D already mentioned. These small lenses (hundreds of meters wide and tens of meters thick) have volcanic pebble to cobble conglomerates, very red and exhibit clast support with a wacke matrix, with subrounded to subangular clasts. These epiclastics are rich in feldspar and chloritized mafic detrital grains and volcanic rock fragments. Texturally they are also very immature, with a clayey to silty matrix, very poor sorting, very angular grains, and matrix support of the larger volcanic rock fragments. The andesite flows in a section beneath two of these lenses were also locally oxidized and red.

The next interval upwards stratigraphically is composed mainly of light to medium grey or reddish-grey, fine- to very fine-grained plagioclase phyric dacitic volcanic flows. The plagioclase grains locally are stained pink or red. The flows are very massive and have associated breccias, and are somewhat less resistant, greyer and lighter in color than the green andesites; the color is the outstanding characteristic that made them mappable in the field. The associated breccias are composed of angular grey dacite clasts in a light grey glassy matrix with microlites of feldspar and granules of quartz. The matrix has been partially devitrified in a spherulite pattern and is rather cloudy and altered. There are also thick interbeds of light grey diamictite breccia with large angular clasts of the same grey dacitic lithology supported by a grey aphanitic siliceous matrix. The matrix in the tuff breccia has some devitrification spherules as well. The dacite interval is 350 m thick, and is assigned to member F.

Overlying the dacite flows with an irregular and channelled bottom is a laterally extensive lens (1.2 km) of green cross-bedded arkosic wacke as much as 40 m thick, which thins and pinches out to the south. These epiclastics are texturally immature - very poor sorting, angular grains, wacke matrix (mostly silt-sized) but grain support, and compositionally unevolved. They contain sericitized plagioclase, magnetite, and chloritized mafic detrital grains and some volcanic rock fragments but no quartz, with some hematitic staining of grain surfaces. There is substantial soft-sediment deformation in this epiclastic horizon, with boudinage of beds and rootless isoclinal folds. Trough cross-bedding is a common feature. STernet analysis of the cross-bedding (fig. 24) indicates that the paleocurrent direction was to the west, and that paleoslope dipped in that direction. This interval is intruded by a steep NNE-trending green foliated dike, with devitrified glassy borders. The intrusion of the dike, syn-sedimentary fault activity, or the



POLES TO CROSS-STRATIFICATION FROM MEMBER F, HAPPY CREEK FORMATION

TROUGH CROSS-BEDDING MEASURED. THE BEST FIT GIRDLE IS 334, 80 E WITH A POLE (FILLED BOX) AT 244 10. THIS POLE IS THE INFERRED PALEOCURRENT DIRECTION; THE 10 DEGREE DIP TO THE WEST MAY INDICATE PALEOSLOPE DIPPED IN THAT DIRECTION.

FIG. 24

emplacement of the overlying volcanic flows of the next unit up could have caused the soft-sediment deformation. For simplicity, the epiclastics have been assigned to F.

In-between the dacite unit and the overlying King Lear Formation is a tongue of uppermost Happy Creek lithologies up to 770 m thick. The tongue thins and disappears to the south (as did the lens of member D lower in the section). In the headwaters of Hobo Canyon the dacite is directly overlain by the King Lear. The lithologies in this wedge are mostly green, red and grey fine-grained plagioclase phyric andesite flows and flow breccias. The texture of the plagioclase porphyries is felty, and they are locally trachytic to pilotaxitic. These flows are thick and massive (where contacts are discernable, around 20 m) with aphanitic tops. The red coloration is caused by alteration and staining of the groundmass. There are thin (30 cm) red plag-phyric andesite dikes intruding some of the flows. Lahars (the volcanic matrix-support diamictite conglomerates and breccias) are also present in the section. It is significant that a portion of the clasts in these lahars are of the Early Mesozoic Intrusive suite; previously volcanic facies of the Happy Creek made up the overwhelming majority of clasts of the coarser epiclastics. This facies has been assigned to member G.

Where member F is in contact with the King Lear Formation, F is oxidized and weathered to a depth of up to 10 m (but most intensely in the top few m), with seams of gray caliche carbonate in cracks and crevices. Member G has less extensively developed weathering, and is locally interbedded with the basal King Lear Formation.

Age There are no direct age constraints from the Happy Creek Formation of this area. Some constraints are provided by the overlying King Lear Formation, which contains middle Lower Cretaceous freshwater molluscan fauna in this vicinity (Willden, 1963). The Happy Creek in area 4 is thus older than middle Early Cretaceous (and probably significantly older, as these fossils are from a position high in the King Lear).

Depositional Environment The lowermost Happy Creek Formation in this area is represented by the green andesite flows of member A in lower Hobo Canyon. Little can be said of their origin; no diagnostic features were observed.

The overlying green epiclastic interval of member C is more distinctive. It is characterized by extremely immature and proximal first-cycle sediments with a volcanic provenance. Clasts as large as boulders are present. The detrital fragments include volcanic rock fragments (all identifiably intraformational), sericitized plagioclase grains, and chloritized mafic grains. Bedding is thick, with pebbley lenses and thick fining-upwards trends.

The set of characteristics observed in C indicates a proximal alluvial fan setting, with deposition by fluvial, debris-flow and sheet-flood processes in a semi-arid environment (Davis, 1983; Collinson, 1986). More specifically, the matrix-support diamictites are muddy debris flow deposits, and the pebble to boulder conglomerates and arenites are stream channel facies. Some of the arenite and wacke beds may also represent stream channel and/or sheet-flood deposits. No sieve conglomerates (with empty interstices between clasts) were seen - these would be most indicative of the apex of an alluvial fan. Limited finer-grained facies are present, and then only in the thinner, finer-grained and more distal (in terms of slightly better sorting and rounding) southern portion. The siltstone facies there may represent overbank facies in a distal or interlobe alluvial flood plain setting; the coarser lithologies were probably laid down in a braided stream setting. The likely setting for member C in the northern part of area 4 would be the proximal middle portion of an alluvial fan; the southern part of the area was laid down in somewhat more distal to interlobe setting in the alluvial fan facies pattern. The 20 to 50 m fining-upwards sequences most likely were caused by large-scale events affecting the sediment supply and fan development (either tectonic or related to evolution of the source volcanic terrane).

The red bed lens of member D intertonguing with the upper strata of member C is also quite distinctive. This large composite channel body is an amalgam of trough to tabular cross-bedded pebble-imbricated pebble to cobble conglomerate and arenite with clast support, filled framework and a general lack of fines. This set of features is diagnostic of filling of a major channel within the alluvial fan by channel lag and point-bar deposits in a fluvial environment. Member D appears to have been a major channel incised into the alluvial fan. The less abundant thin-bedded sediments may have been levee or top-of-point-bar deposits preserved by chance from fluvial scour and erosion. The general coarseness of the lithologies, lateral and horizontal variability, and abundance of erosional surfaces and cross-stratification indicate a braided stream environment (Reineck and Singh, 1980). Once again all clasts are intraformational volcanics, and the coarse-grained sediments are highly immature. Volcanic rock fragments, detrital and altered plagioclase and mafics, quartz grains, and silt and clay are the constituents. The red hematitic cement and staining is also stongly indicative of a subaerial environment. The paleocurrent direction appears to have been from the east.

Volcanic facies are a minor component of the predominantly epiclastic members and are seen only in member C. They include andesitic flows and rhyolitic facies (including foliated flow and flow breccia and lapilli and crystal tuff). Their presence and the fact that the debris in units C and D is entirely volcanic in provenance indicates that the highlands and source for the alluvial fan of member C and D are of volcanic constructional origin. Pebble imbrication indicates that these highlands may have lain to the east. The abundance of such siliceous volcanic lithologies in the clasts is interesting, as such facies have only rarely been observed bedded in the Happy Creek Formation, but must have been a locally significant part of the unit despite this.

The thick volcanic succession of member E overlying members C and D in this area consists of thick green plagioclase phyric to aphyric flows and lahars. Viscous mass flow deposits such as these can be either hot and directly related to explosive volcanic eruptions, or can be caused by cold sedimentary re-working of volcanic deposits; no diagnostic features such as charred tree trunks or gas escape pipes were seen, though in the field the matrix in most was thought to be sintered (indicating a hot origin - Cas and Wright, 1987).

The red bed channel-fill bodies within E have a concave geometry and apparently were deposited in channels eroded in the volcanic successsion by fluvial processes. The oxidized flows below the sedimentary lenses were probably weathered during the existence of the superposed fluvial channel. The sediment was extremely immature and proximal. These channels were canyons cut into the volcanic pile itself, filled and overlapped.

The next overlying unit, the dacitic succession of member F, is a series of thick, blocky flows and auto-breccias. Flow-foot, flow-front and top-of-flow deposits are all probably present and indistinguishable. The monolithologic nature of the clasts and glassy matrix indicates that they are flow-related. Lahars are also abundant, though in this unit they are of a lighter, greyer nature similar to the host flows; local derivation seems likely. The superjacent green volcanic wacke near Jackson Creek has an erosive base on the dacites and is inferred to be another braided stream system within an upper alluvial fan setting. The source may have been to the north as the deposit thins and disappears to the south. The provenance for this epiclastic unit is again entirely first-cycle volcanic. The soft-sediment deformation in the epiclastics is thought to be caused by the intrusion of the andesitic dike, given the spatial tie. The dike may have been a feeder to the overlying flows.

The wedge-shaped flow, flow breccia and lahar sequence at the top of the Happy Creek is placed in member G. Although it is somewhat similar to member E it differs from the stratigraphically distinct and lower occurrences of E. The pronounced lateral thickness variability, the oxidation of a number of the flows, the fine-grain size generally seen in the phenocryst population, the increasing abundance of augite and the presence of intrusive clasts within the lahars are distinctive features. These are more similiar in nature to the red andesites (with the color due in this case to Fe-mineralization) of member H within the grabens at the very top of the Happy Creek in other areas (specifically 3). These characteristics within member G lithologies may reflect that activity on the east-west trending vertical fault system had already begun at this time. One of the larger faults of that family bounds the northern extent of this outcrop area, where the tongue is at its thickest. The Fe-mineralization and scarp-related sediments that are also typical of member H are absent, however. It is more probable therefore that the fault dates from later in Happy Creek time and cross-cuts the unit, with the thickness distribution due to earlier volcanic constructional evolution. I consider this the more likely possibility, as strata best correlated to this unit elswhere in the Jacksons (area 3) are quite thick and do pre-date the faulting. Another alternative is that the thinning of the unit to the south was caused by erosion at the base of the King Lear. This seems unlikely as facies characteristics of the two formations are interbedded at the contact.

(5) Bliss Canyon

This area extends across the southern Jackson Mountains north of King Lear Peak from the ridge south of Hobo Canyon (and area 4) to south of Bliss Canyon (where a large high-angle fault is defined as the boundary between this and area 6). The setting is similar to that for 4 - eastward vergent deformation (with low-grade metamorphism in the structurally highest sheets but insignificant metamorphism elsewhere), distinct and mappable stratigraphy, and only a single pluton of the Early Mesozoic Intrusive Suite intruding the strata. The contact with the King Lear is, however, somewhat obscured in the southern part of the area by intrusion of a later hypabyssal rhyolitic sill complex of King Lear affinity into the strata of both units. See fig. 25 for details.

Stratigraphy The lower depositional contact with the Bliss Canyon Formation is preserved in several localities within the thrust belt on the western flank of the range. The quartzose and cherty thick-bedded, sorted and rounded, scoured and cross-bedded conglomerates and arenites of the Bliss Canyon are immediately overlain by 10 m or so of green coarse-grained arkosic volcanic arenite and then by mildly metamorphosed greenstone and phyllitic volcanics. The metavolcanic lithology, where best-preserved just south of Alaska Canyon, is a green amygdular augite-plagioclase porphyry basaltic andesite. The amygdules are filled by calcite and chlorite. The groundmass is pilotaxitic to intersertal, with plagioclase and Fe-oxide microlites in a banded glassy matrix, which has been extensively replaced by chlorite. In most places shearing and low-grade metamorphism has obliterated original textures. This set of facies is at least 270 m thick - the actual thickness is greater, but the interval is affected by thrusting so that no uninterrupted section of it exists. The interval is assigned to member A.

The next unit upwards is epiclastic green coarse-grained volcanic arenites which are thick-bedded and homogenous, with few sedimentary structures. There are also some interbedded green amygdular andesite flows. Though red volcanic wackes are also seen in the section, their

5 BLISS CANYON

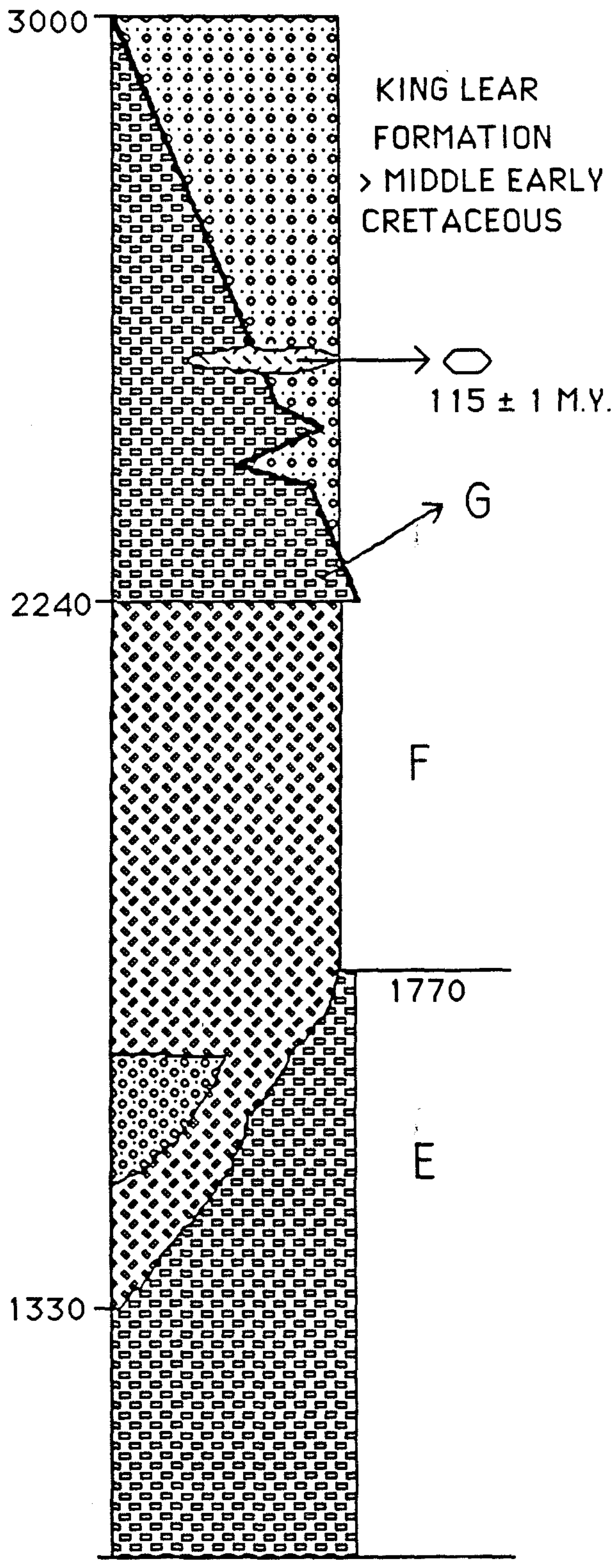
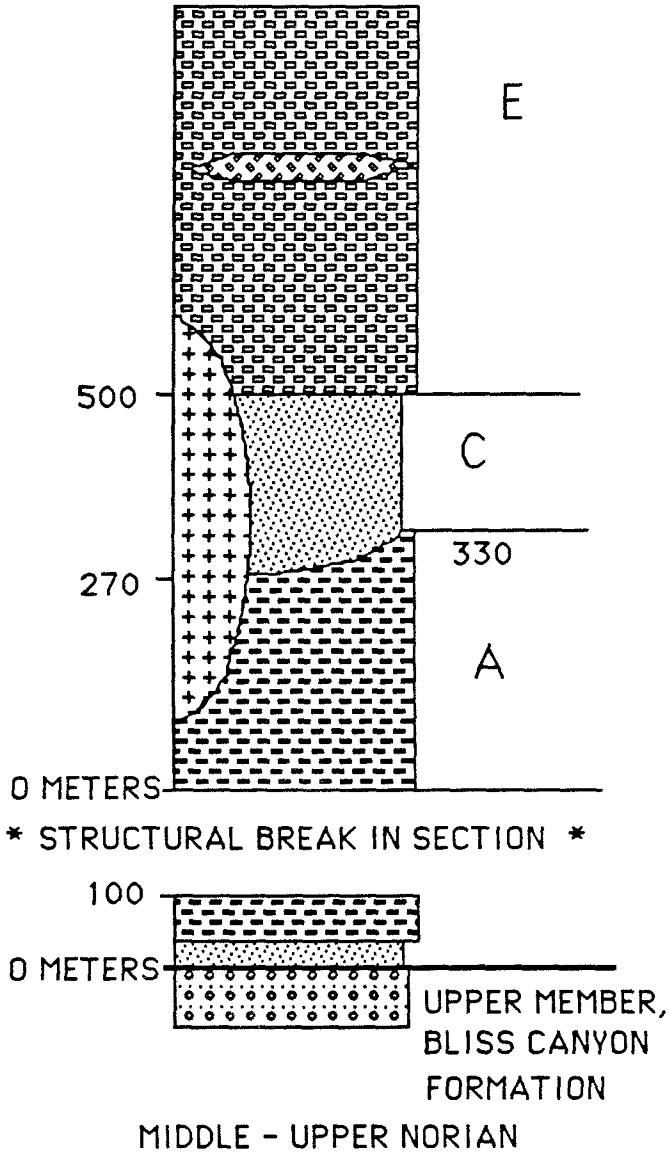


FIG. 25


E
500
C
330
270
A
0 METERS
* STRUCTURAL BREAK IN SECTION *
100
0 METERS
UPPER MEMBER,
BLISS CANYON
FORMATION
MIDDLE - UPPER NORIAN

relative abundance is low. This epiclastic interval is 170 to 225 m thick and is assigned to member C; it is much thinner and finer-grained than the same unit several kilometers to the north in area 4.

Overlying the epiclastics is another volcanic succession. Fine-grained green and red plagioclase-porphyry andesite flows and flow breccias are the dominant lithology, containing hornblende phenocrysts at the bottom of the section. In addition, there is a horizon of light grey volcanics in the middle part of the section at about 300 m. Though correlated to member E, this area has significantly more weathering and oxidation of the flows (as estimated in the frequency with which red coloration to the flows is present) than exposures of member E to the north. The oxidation of the flows is especially evident in the thinnest part of the section, where a 30 m thick lens of red volcanic cobble conglomerate is also present. This clastic bed has sub-rounded to rounded and sorted clasts sourced within the Happy Creek in a volcanic arenite matrix. The thickness ranges from 1270 m in the south to 340 m to the north.

Strata assigned to member F are next in the section. Grey dacite flows and flow-breccias characterize the section. One lenticular conglomerate and breccia unit is extremely coarse-grained. The clasts are moderately rounded to subangular and moderately sorted fragments of dacite, up to boulder-size, in clast-support in a very fine-grained light grey siliceous matrix. This body is found at the bottom of the thickest portion of the unit and is as much as 200 m thick. In fact, the bottom of the member is not planar, but appears to be concave with the thickest part and the breccia lens above the thinnest part of member E below. The thickening in the member is quite sudden. Some green andesite flow facies are interstratified with the dacites. In the northern part of the area member F is directly overlain by the King Lear Formation, and the dacites, (which are plagioclase-rich with a felted texture) are stained red in the hundred meters or so just below the unconformity (the weathering and oxidation preferentially affecting the matrix). The member is 470 to 810 m thick.

Above the dacite interval is a tongue of member G, thickest at the south right against the high-angle fault and thinning and disappearing rapidly (over about 2.7 km) to the north. Maximum thickness is 760 m. The unit is characterized by red plagioclase-phyric andesite flows and flow-breccia, in individual units 7 to 14 m thick. A pipe of flow-foliated andesite was seen intruding the flows. There are also several thin and laterally extensive lenses of poorly-sorted red volcanic conglomerate with clast-support in an immature (poorly-sorted and subangular) arenite to wacke matrix; these lenses are interbedded with the red volcanic flow facies. Some of the clasts appear to have undergone Fe-mineralization before deposition, and contain magnetite veins. Near the top there are also green flows interbedded with green coarse-grained volcanic arenite. In fact, one of these flows is interbedded with the sediments of the basal King Lear Formation (red volcanic arenite and conglomerate, and green chert-pebble conglomerate and chert-arenite). The topographic dip on the top of this tongue at the initation of King Lear deposition (probably after some erosion) was about 10° to 12°, as the bottom of G is rather planar. This is a minimum value, as the exposed profile is not necessarily along dip.

Age The bottom of the Happy Creek Formation in this area (member A) is constrained by the underlying Bliss Canyon Formation to be middle Norian at the oldest, and probably slightly younger as that biostratigraphic data is from the middle and not the upper member of Bliss Canyon Formation. The upper part of the section (member G) is intruded by rhyolitic sills with a preliminary U/Pb zircon date of 115±1 m.y. Member G is overlain by middle Early Cretaceous sediments relatively high up in the King Lear Formation, and so must be somewhat older than middle Early Cretaceous. The section therefore dates to within the period late Norian to early Early Cretaceous.

Depositional Environment Not much can be said of the facies in member A because of the limited exposure and overprinting. The augite-porphyry rocks are flows, as they have a glassy matrix with intersertal texture and amygdules. Based on the size of the amygdules, the flows are unlikely to have been extruded in water of depths below several hundred meters. The basal green volcanic arenite could be shallow marine to littoral or alluvial plain; no distinguishing features were found.

Member C in the Bliss Canyon area is substantially thinner and finer-grained than to the north, lacking the conglomerates and matrix-support diamictites (debris flows) that were seen to the north. The volcanic wackes are less abundant as well. The slight increase in maturity (better sorting and cleaner), the lack of debris flows (usually a more proximal facies) and the finer grain size (no conglomerates) indicate that the deposits are most likely stream channel and sheet-flood facies from a sandy braided fluvial system in the distal alluvial fan or interfan alluvial plain. A braided is preferred over a meandering system because of the general coarseness of the sediment, the lack of well-developed channels and lack of interbedded inter-channel fines.

The superjacent strata of member E are another volcanic succession. Green and red andesite flows and flow breccias are the dominant facies. These flows are subaerial, as indicated by the textures, common oxidation and lack of subaqueous indicators. The percentage of mafics (hornblende, in this case) in the unit decreases upwards from the base, with none observed in the upper part of the section. The channel cut-and-fill lenses of red proximal intraformational conglomerate are fluvial in origin. They represent canyons cut in the volcanic strata at the edge of a volcanic constructional edifice, filled by very proximal and coarse-grained fluvial debris and overlapped by more flows. There seems to have been a very large valley in this area, with as much as 800 m or so of member E missing. The sedimentary lenses and oxidation are concentrated in this area, and the valley was later filled by member F, which is at its thickest in the paleovalley.

Member F is made up of subaerial blocky dacite flows and auto-breccias, which can be highly felsic at the top of the unit. They are interpreted as subaerial flows because of the fine groundmass, oxidation and lack of subaqueous features such as pillows or bedded hyaloclastite debris. The very thick and extremely proximal conglomerate lens is fluvial, and is not interpreted as a rock-fall deposit because of the rounding, channel-fill nature and presence of framework-fill matrix. In the bottom of the valley-fill sequence of member F is another fluvial channel deposit, overlain by more dacite flow facies. The dacites are also preferentially oxidized and red in color near the bottom of this valley feature. Where the King Lear directly overlies member F, more weathering and oxidation is evident; a depositional hiatus took place between the two units.

The tongue of member G is composed of subaerial aa to blocky flows and flow-breccias, similar to the tongue of the unit seen in area 4. Interbedded thin laterally extenisve beds of red immature (very poorly sorted, subangular and with muddy matrix) clast-support breccia and conglomerate are interpreted as very proximal stream channel or rock-fall facies. Though similar to an upper alluvial fan depositional setting, the relatively steep relief inferred to have existed during deposition of these beds was probably due to slopes caused by the building of the volcanic edifice and not by constructional alluvial fan development. Some minor high-angle fault activity during the deposition of member G in this area is indicated by the abundance of coarse clastics in the unit (shed, perhaps, from a fault scarp), and by the presence of sedimentary clasts with the Fe-mineralization characteristic of those faults. The high-angle fault might have contributed to the depositional morphology and wedge-shape of the member as well, but most of it is probably due to constructional relief. The conformity and interbedding of the uppermost Happy Creek (member G) and King Lear Formations in this area indicates that the transition between the two units is gradational.

To summarize, the Happy Creek Formation in the Bliss Canyon area records subaerial volcanism and related epiclastic sedimentation. The volcanics get less mafic upwards overall (the lowest volcanics in member A are augite-plagioclase-phyric, the base of member E has plagioclase $\pm$

hornblende andesites and the top only plagioclase-porphyry andesite, and member F is more siliceous and felsic yet. The lower epiclastic interval of member C was deposited in a distal, braided alluvial plain setting. A large valley was cut into member E (andesites) and partially filled by member F (dacites); in both members fluvial lenses and oxidation of the volcanics are most common in that valley. The position of the valley in E and F has shifted from the large channel and proximal alluvial fan of member C and D lower in the section seen in area 4 just to the north. Deposition of the tongue of member G in the southern part of the area was on the margin a volcanic pile, with some influence by high-angle fault-activity possible (the same conclusion applies to the tongue of member G to the north in area 4). The King Lear rests on and is interbedded with member G or, where G lenses out, overlies member F with prominent weathering in F.

(6) King Lear Peak

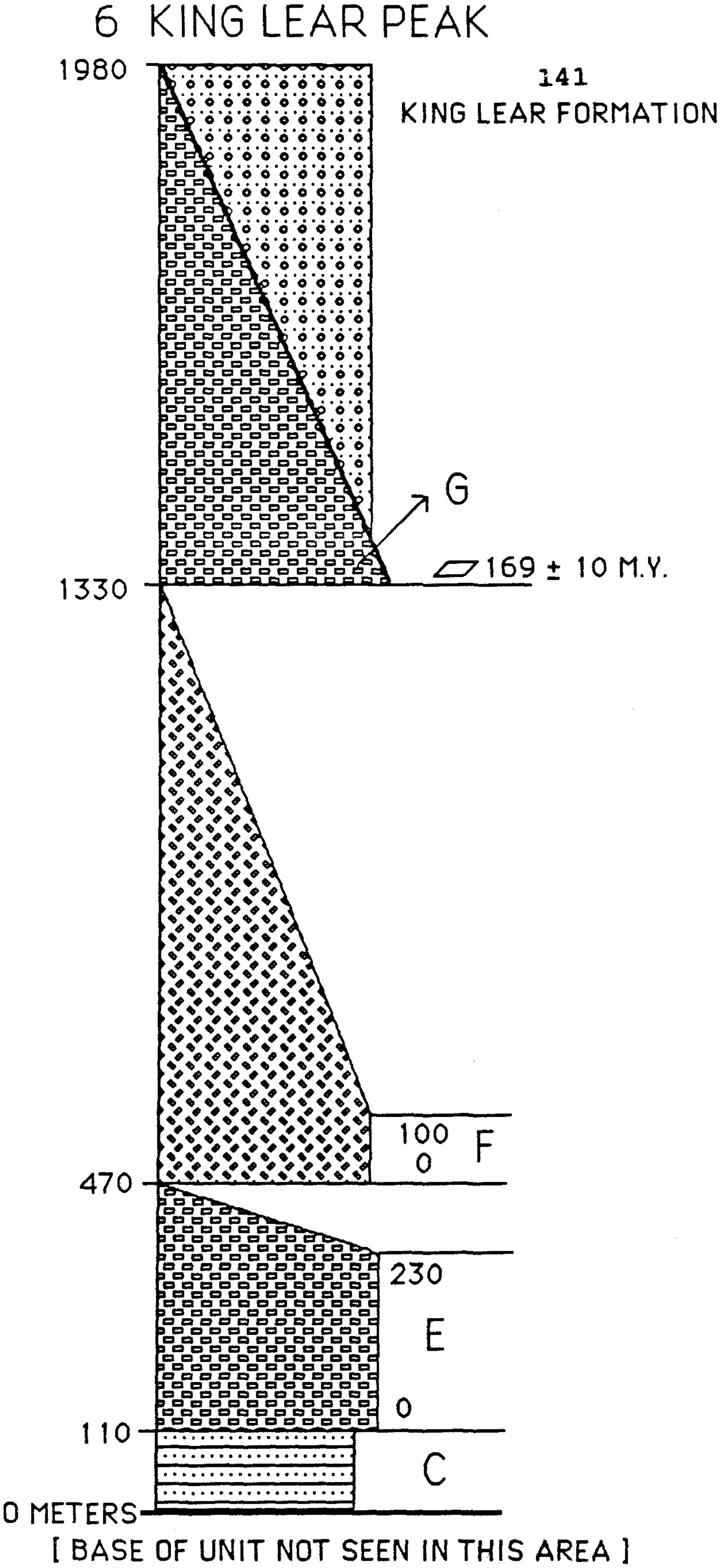
The Happy Creek Formation in this area of the Jackson Mountains (south of a large east-west high-angle fault and area 5, and north of 41°7'30") is folded and faulted by the east-vergent thrust belt and has been intruded, on the extreme south, by a large gabbroic to granitic composite pluton of the Early Mesozoic Intrusive Suite. In the northern part of the area rhyolitic sills of King Lear affinity intrude the uppermost Happy Creek. The base was not preserved in this thrust zone, and while the stratigraphy in the formation was less distinct and less information was gathered than to the north, it was still traceable. The section is unmetamorphosed, except for the contact aureole around the pluton. See fig. 26 for details.

Stratigraphy The lowest unit of the Happy Creek exposed in this area is member C, the green epiclastic interval. Green and red medium- to coarse-grained volcanic arenites and red shale were seen, with no sedimentary structures other than bedding. The bottom was not exposed so that a minimum estimate of the thickness is 110 m.

Member E overlies C and consists of homogenous green andesite flow facies. The thickness decreases to the south, from 360 to 230 m.

The overlying dacite succession of member F is highly variable in thickness through the area. It is 860 m thick in the north, west of King Lear Peak; 440 m to the south, just north of the Navajo Peak pluton; and as little as 100 m to the east of King Lear Peak. It is made up of the typical light grey to bluish-grey volcanic flow and flow breccia, in places oxidized and red. It also has epiclastic interbeds, as to the north, of grey volcanic clast-support cobble conglomerate.

Most of the exposure of Happy Creek in this area is member G. King Lear Peak is itself entirely composed of this member,and not of King Lear Formation, despite the name! Indeed, the crown on King Lear Peak is formed by a combination of bedding and jointing in these thick flows. Characteristically, the unit consists of well-exposed and well-bedded red plagioclase porphyry amygdular andesite flows and flow breccias in laterally persistent tabular bodies 5 to 20 m thick; some of the flows are also green, but red is the predominant color. A lens about 30 m thick of light brown volcanic conglomerate with well-rounded clasts, large-scale cross-bedding, clast support and a volcanic arenite matrix



6 KING LEAR PEAK
1980
141
KING LEAR FORMATION
G
169 ± 10 M.Y.
1330
100
0
F
470
230
E
0
110
C
0 METERS
[BASE OF UNIT NOT SEEN IN THIS AREA]

FIG. 26

with interbedded green siltstone was encountered low in the section north of King Lear Peak. This level in the member on the west side also has some thin red immature (poorly sorted, subangular) volcanic wacke interbeds as well. The overlying majority of the section is a massive pile of well-bedded red andesitic flows, diamictite with volcanic matrix-support, and minor red breccias (with sedimentary matrix-support). King Lear Peak is a homoclinal section of east-dipping beds of this character. Member G is about 750 thick at King Lear Peak, 220 m thick further south, and as little as 20 thick further south yet; if this was caused by positive volcanic constructional relief then the slope angle was about 10° (perhaps after erosion). The basal green siltstone, grey limestone and red volcanic conglomerate of the King Lear Formation rests conformably on the thinner, southern portion of the member, but with local channels incised into the Happy Creek and with red lisagon staining and light grey carbonate caliche veins in the few meters just below the contact.

Age Russell (1981) reported Rb/Sr geochronological data from four flows from the top of member G of the Happy Creek near King Lear Peak. He obtained an isochron on the four flows of 169±10 m.y. (2 Ú). No other age information is known for this area. The Happy Creek Formation is therefore late Middle Jurassic and older in this vicinity.

Depositional Environment Member C appears to be similar to exposures of the unit to the north in the Bliss Canyon area, with the addition of red shale interbedded with the volcanic arenites. Alluvial flood-plain deposition (sandy braided stream channel and shaly flood-plain over-bank deposits) is inferred, but substantially more distal than in areas 5 and, especially, 4. No new information can be added for members E and F beyond the conclusions already reached earlier from other locations.

The red andesite flows, flow-breccias, and related epiclastics of member G are subaerial. The andesite flows must have been relatively fluid flows (less viscous than is often the case for their andesitic composition, or were formed on a steep slope), due to their lateral continuity and constancy of thickness. The flows were aa to blocky in nature as indicated by the copious associated flow-breccias and range in unit thickness. The interbedded sedimentary diamictite breccias are probably debris flows (very poor sorting, no internal structure and muddy matrix-support) and the red volcanic wackes sheet-flood deposits on a sloping volcanic surface (thin non-lenticular beds in-between volcanic flows). The body of brown conglomerates and green siltstones is a fluvial channel (as indicated by the relative maturity of the deposit-rounded clasts forming a framework, with a sandy infilling matrix, and the large-scale cross-bedding). The siltstone might be of top-of-point-bar origin (though not levee or overbank - the slope is too steep). The channel was cut into the volcanic pile, at least partially filled by sediment and overlapped by more flows.

Member G has an intriguing thickness pattern, and appears to be part of a single small stratovolcanic edifice overlying member F, as it decreases rapidly in thickness to the north and south with slope angles of perhaps 10 to 12°on both sides. It was originally at least 12 km across and 800 m high, assuming minimial erosion, and acknowledging that this could be an off-center section. A similar pattern exists to the north in areas 3 and 4. The volcanic edifice may also have been affected and sliced across by the east-west high-angle faulting during the extrusional phase, and certainly before and during the deposition of the King Lear Formation (this will be discussed in detail later). These high-angle faults were oblique-slip, with both strike-slip and/or dip-slip stages. The construct might have been later overlapped and buried by King Lear sediments.

The King Lear Formation in this area overlies member G of the Happy Creek with evidence of significant weathering (the lisagon iron staining and caliche veins) of the latter before deposition of the former. The climate would have been semi-arid, judging by the kind of weathering and sedimentary facies (Collinson, 1986) and by the red bed nature (implying hematitic cement) of the sedimentary rocks interbedded with the volcanics, and even by the oxidation of the flows themselves.

(7) Southern Trout Creek Spur

This section, located at the southeast corner of the Jackson Mountains at the south end of Trout Creek Spur south of 41°16′, has been involved in extensive imbricate thrusting in the eastern, west-vergent thrust system. Several plutons also intrude the section and the thrusts. As a result, the stratigraphy within the Happy Creek Formation in this area has been difficult to work out. The conformable depositional relationship with the underlying Bliss Canyon Formation is, however, exposed and is one of the more obvious mappable contacts in the area. No King Lear strata are found in the area, and therefore the upper contact of the Happy Creek is fault-bounded. See fig. 27.

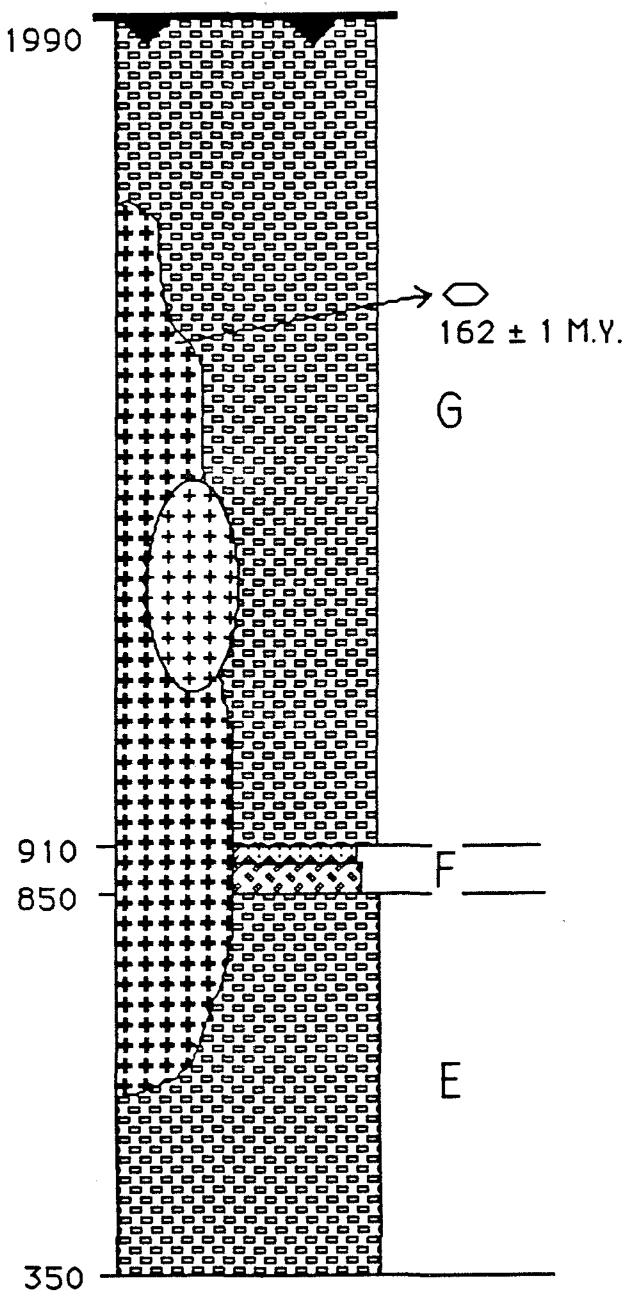
Stratigraphy The lowermost Happy Creek Formation, overlying the grey argillite of the lower member of the Bliss Canyon, in the southern part of area 7 is made up of 30 m of epiclastics. Massive green volcanic arkose coarsens upwards into green volcanic conglomerate with clast-support and an arenitic matrix. This is overlain by at least 320 m of thick-bedded, dark green, coarse-grained augite-plagiocase porphyry basaltic andesite flows and breccias. At the base of the sequence some beds have xenoliths of light grey micritic limestone. The top of this section is truncated by a thrust, but in another thrust slice member E overlies A. The sedimentary facies of members B, C, and D are all missing. The augite-phyric flows are not highly amygdular, and the amount of augite in the rocks decreases upwards in the section. A volcanic diamictite was also present in the section, and in one area barite-filled breccia dikes cut this member and the overlying member E.

Member E is comprised of grey and green fine-grained plagioclase-phyric andesite with only rare augite, and is significantly more amygdular, finer-grained and lighter in color than A. The groundmass is aphanitic with a sugary texture. Textures are commonly trachytic, particularly in the coarser-grained lithologies. At least 500 m of section E exists below member F on the western side of the range, but the base of E was not seen. In more easterly thrust slices E is 310 m thick, with inferred contacts with both A and F present.

The dacite interval of member F is very thin in this area. It consists of light blue-grey and grey siliceous very-fine grained

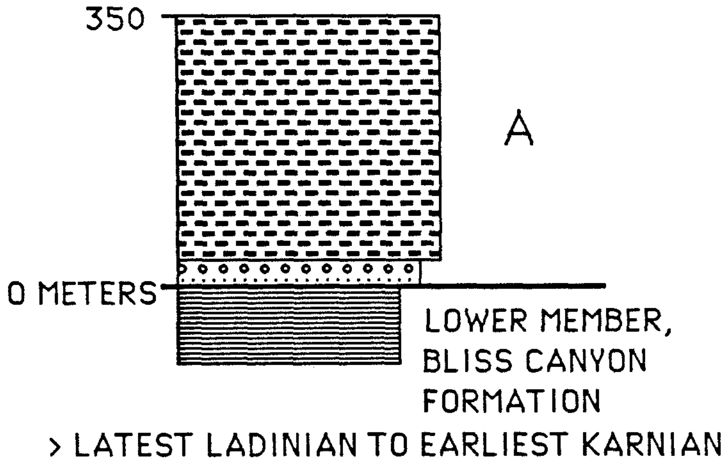
7 SOUTH TROUT CREEK SPUR

TOP NOT SEEN - STRUCTURAL TRUNCATION



THICKNESSES OF MEMBERS A AND E ARE
MINIMUM ESTIMATES DUE TO THRUST FAULTING

FIG. 27


350
A
0 METERS
LOWER MEMBER,
BLISS CANYON
FORMATION
> LATEST LADINIAN TO EARLIEST KARNIAN

plagioclase-rich dacite and dacite breccia in an aphantic light grey matrix. Flow textures are trachytic, with andesine plagioclase (An_{35}), and no mafics other than abundant Fe-oxide. The breccia is poorly sorted (sand-sized to cobbles), angular, and in matrix support. At the very top of the flow the breccia is finer-grained and in clast-suppport. The dacite is locally oxidized and reddened (particularly the feldspar). The entire member in the western, lowermost thrust is 60 m thick, and higher in the stack and to the east it is only about 20 m thick and is dominated by breccia.

Overlying the dacite is another andesitic succession, placed in member G. It is characterized by grey and green fine-grained plagioclase phyric andesites and associated flow breccias. The plagioclase, with An_{35}, is andesine, and hornblende is a common phase. The matrix is made up of a felted and intergranular network of plagioclase laths and hornblende and Fe-oxide granules. The flow breccias contain grey-green glassy clasts up to 30 cm in size. The clasts are even more amygdular than the matrix, and very highly vesicular (up to 40% of the rock). The flows exhibit hyalopilitic to trachytic textures and a very fine-grained matrix. Volcanic diamictites are also abundant in the section. Some flows are reddened and oxidized. Near the base the andesite flows are locally xenolithic breccias, containing up to 40% fresh angular diorite fragments. The amygdules are ellipsoidal to flattened (by flow shear), and filled by sulfides quartz chlorite in succession. Some of the voids are up to 14 cm in diameter, with sprays of epidote, quartz, azurite, malachite and specular hematite filling them. Some flows have augite, up to 8 mm in length, with fine-grained brown coronae indicating disequilibrium. Small lenses of volcanic arenite are also present. The top of the member is not exposed, and member H was not present in this area either. The thickness of member G is at least 1080 m.

Age The underlying Bliss Canyon Formation is latest Ladinian to earliest Karnian in the middle of the section in this area (Russell, 1981; and see earlier discussion, chapter 3). No age data exists from the Happy Creek Formation itself, but all members are intruded by the large monzodioritic Trout Creek Spur pluton. This pluton has a preliminary U/Pb zircon age of 162$\pm$1 m.y. The Happy Creek in this area must be younger than late Middle Triassic and older than Late Jurassic - i.e., Late Triassic to Middle Jurassic. The Happy Creek is probably significantly younger than the base of the Upper Triassic because several hundred meters of Bliss Canyon Formation overlie the dated strata and underlie the Happy Creek. The section is probably also significantly older than the pluton, as the pluton cuts the thrust faults, which cut the Happy Creek.

Depositional Environment The conclusion arrived at concerning the origin of member A in areas 1 and 3-5 apply here as well. The section is composed of blocky flows and auto-breccias with lahars (debris flows-the diamictite). Whether the environment was subaerial or not is not certain. The lack of pillows and related hyaloclastite debris may indicate a subaerial origin. The overlying andesite flows of member E are also subaerial by the same reasoning. The barite breccia dikes are due to hydrothermal activity at an unknown time.

Within member F, the dacite breccia is probably caused by both sedimentary reworking (at the very top of the flow) and flow-related autobrecciation, at the top of and within the dacite flow. Some length of subaerial exposure is indicated by the weathering of the unit as well as by the epiclastic cap. Judging by the thickness, this could be a single flow.

Member G is quite distinctive due to the very high degree of vesicularity of the flows. The sizes and volume percentage of the vesicles point strongly to a very shallow marine or subaerial origin, and the red weathering of some of the flows supports a subaerial extrusion. The section is composed of volcanic flows (as evidenced by the textures and nature of the groundmass and the shearing of the amygdules), flow-breccias (glassy and scoriaceous fragments in andesitic matrix) and lahars (diamictites) with rare small epiclastic channel scour-and-fill (volcanic arenite). The sulfides filling the vesicles indicate hydrothermal activity in the area, probably very soon after extrusion as the vesicles were still voids. The xenolithic breccias developed by the entrainment of large amounts of diorite in the feeder system to the flows; the fragments could have been concentrated by settling within a flow after extrusion.

(8) Iron King Mine and DeLong Peak

This area is situated in the Trout Creek Spur north of 41°16′ and within the eastern thrust domain. The section has been extensively imbricated during thrusting, and has also been pervasively hydrothermally altered. Several very thick and laterally extensive diorite and gabbro sills intrude the Happy Creek Formation here. The internal stratigraphy and structure have been quite difficult to work out as a result of the complex structure and a relative lack of distinctive and mappable horizons in this area. The basal contact with the Bliss Canyon Formation is very well exposed; the King Lear Formation is present at the top of several thrust sheets as thin sequences resting on the Happy Creek with angular unconformity and overridden by more thrusts. The King Lear Formation also rests conformably above member G of the Happy Creek. See section 8, fig. 28 for the reconstructed stratigraphy.

<u>Stratigraphy</u> Member A is the easiest unit to identify and examine because of the excellent marker horizon provided by the Happy Creek-Bliss Canyon contact. As elsewhere in the range, the dominant features are the abundance of augite and the dark green color. At least 660 m of section are present in one thrust sheet near Burro Bills Spring.

In the Boulder Creek area, the uppermost strata of the Bliss Canyon Formation (lower member) are chert-pebble conglomerate beds, argillite and calcarenite; these are interbedded with thin- to thick-bedded (1 to 2 m) vesicular augite + plagioclase phyric basaltic andesite flows and associated breccias and cross-bedded coarse-grained epiclastics at the base of the Happy Creek Formation. The flows have entrained large angular clasts of intrabasinal limestone. This limestone consists of thinly laminated packed biomicrite (Folk, 1962) or fossiliferous packstone (Dunham, 1962). The micrite matrix of the limestone consists of brown cryptocrystalline micrite, with allochems (clast-support) of gastropods, calcispheres (probably green algae - Scoffin, 1987), crinoid fragments, inarticulate brachipod (phosphatic shell fragments) and

8 IRON KING MINE - DELONG PEAK

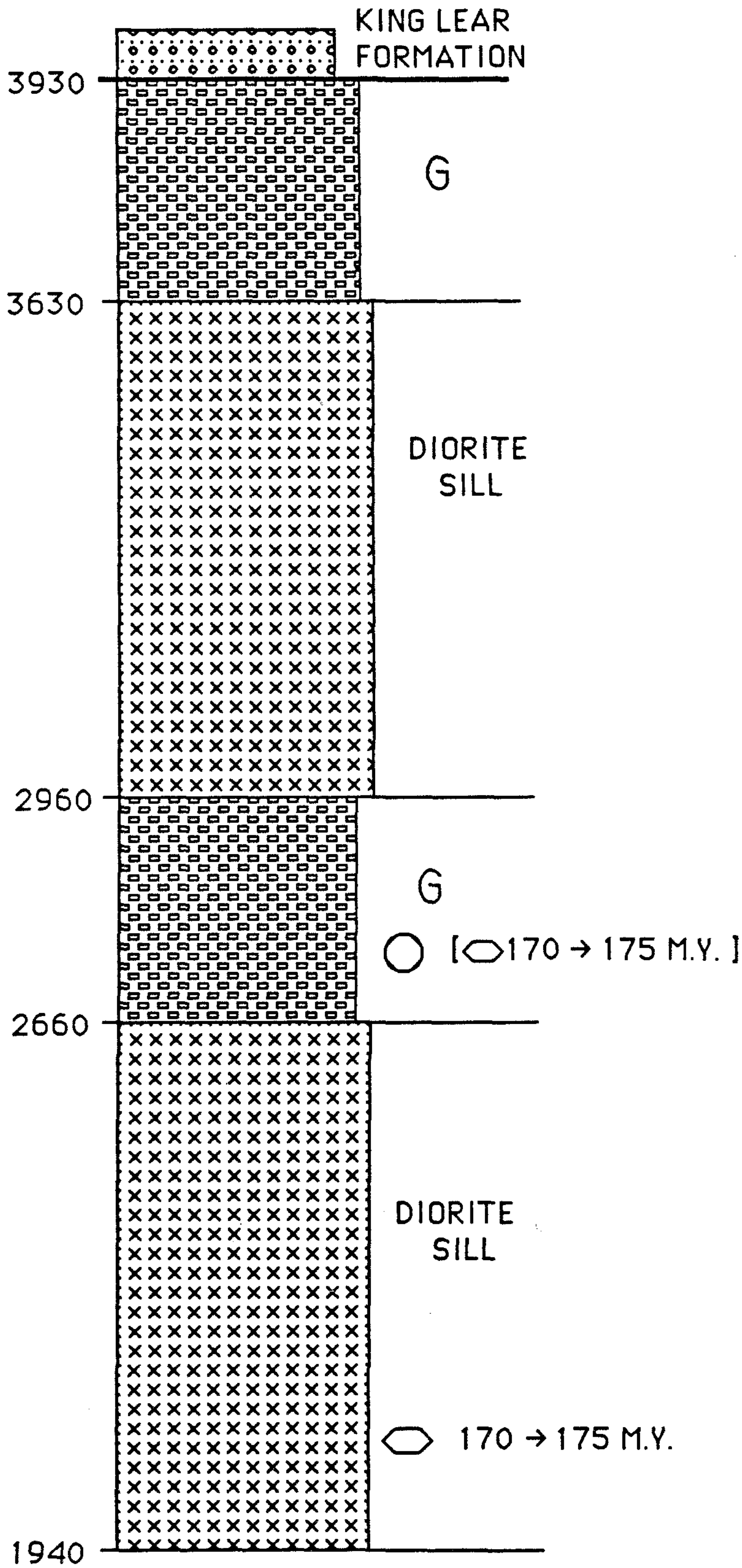
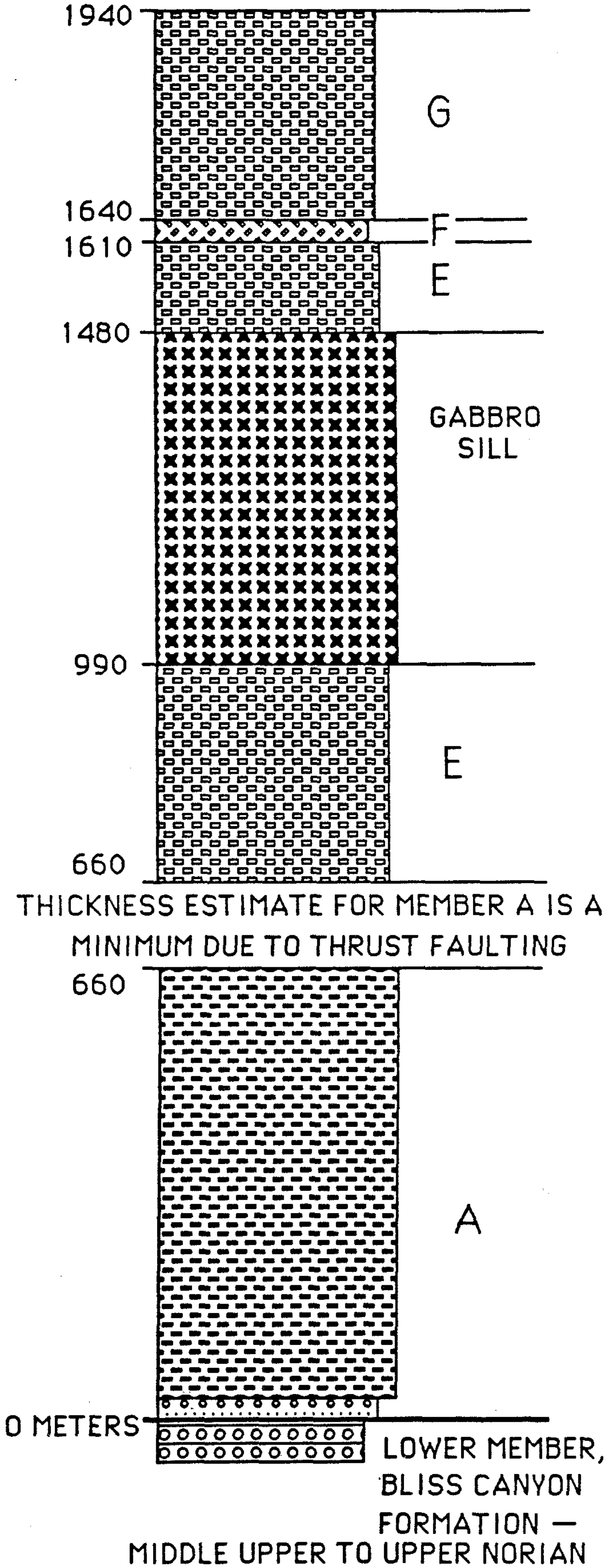


FIG. 28


1940
G
1640
F
1610
E
1480
GABBRO SILL
990
E
660
THICKNESS ESTIMATE FOR MEMBER A IS A MINIMUM DUE TO THRUST FAULTING
660
A
0 METERS
LOWER MEMBER, BLISS CANYON FORMATION — MIDDLE UPPER TO UPPER NORIAN

bivalves (both large thick fragments and thin, delicate valves of calcite; the latter predominate). The skeletal grains are typically small, whole and delicate, with the exception of the broken and abraded crinoid and large bivalve fragments. The clasts of limestone underwent ductile folding before entrainment in the flows (and perhaps related to this involvement). Over a stratigraphic distance of 40 to 50 meters the non-volcanogenic sediments disappear and the flow facies overwhelmingly predominate.

In the Bottle Creek area to the north, the contact with the Bliss Canyon Formation is more abrupt (this is a feature typical of the rest of the range). Here, dark green augite $\pm$ plagioclase flows and breccias (with pre-thrusting hydrothermal alteration and quartz and calcite veins) abruptly overlie the argillite of the upper Bliss Canyon. There are thin and rare lenses of argillite and quartzose arenite at the very base of the volcanic section and they exhibit baking due to the overlying flow units. The flows and breccias are also interbedded with thin lenses of green, poorly sorted and angular volcanic arkose with cross-lamination to lenticular bedding. The flows are very coarse-grained, with phenocrysts up to 1 cm, but have a very fine-grained matrix. Quartz is also present in phenocrysts.

Going up in the section, augite becomes less abundant and plagioclase more abundant. About the middle of the interval, some of the flows are reddened and oxidized. Trachytic to randomly oriented textures are present, and the groundmass in the basal (and non-oxidized) portion is much finer and glassier than higher in the section. Throughout the interval, breccias are commonly associated with the flows. An interbedded lens of volcanic conglomerate exists in the middle of the section. The conglomerate is made up of poorly-sorted but well-rounded pebbles in clast support in a volcanic arkose matrix.

In thin section, the euhedral plagioclase phenocrysts are andesine with An_{32} to An_{35} and are extensively replaced by sericite and by zeolites (one phase of which was identified as thomsonite). The glass has gone to calcite and chlorite. Zeolites are ubiquitous filling in the void spaces. The augite phenocrysts from the base of the member are large, euhedral and fresh and can have oscillatory zoning. The hornblende, however, though also abundant volumetrically, is in disequilibrium in the lower part of the member. It has thin brown coronae (composed of fine-grained radially oriented granules and inferred to be composed of augite) around the euhedral phenocrysts, implying dehydration. The plagioclase, augite and hornblende are all euhedral large phenocrysts floating in a much finer glassy groundmass. Fe-oxides are generally not present in any significant amounts. In one section interstitial mosaic patches of quartz and poikolitic plates of chlorite are seen as late-stage phases, filling in around the phenocrysts and glassy xenoliths. Some of the basal flows also have breccias with angular xenoliths of vesicular green glass caught up in basaltic andesite matrix. The vesicles are commonly small, nearly uniform in size and perfectly spherical to ellipsoidal (where flattened by adjacent phenocrysts) and are filled by zeolites. These glassy fragments look quenched.

Some labradorite (An_{54} to An_{68}) is present in one flow at the base of the member near Buff Peak as very large discoidal phenocrysts; this distinctive rock lacks mafics (other than very abundant Fe-oxides - the opposite of the usual arrangement of mafic phases in this area) and has a glassy amygdular groundmass that has been replaced by chlorite. Around Buff Peak the rest of the basal Happy Creek is a green plagioclase + quartz + augite $\pm$ hornblende porphyry andesite.

The top of member A (where seen south of Trout Creek) is composed of dark green coarse-grained plagioclase-hornblende-augite porphyry basaltic andesite with interbedded volcanic diamictites. The plagioclase is andesine (An_{36}). The proportion of hornblende to augite is greater than lower in the member, and the augite has been completely replaced by chlorite, quartz and Fe-oxide. The hornblende, however, is stable; this is the opposite of the relationship in the lower part of the member. The hornblende, plagioclase and augite occur as large euhedral phenocrysts in an amygdular very fine-grained intergranular groundmass of hornblende + Fe-oxide granules + plagioclase microlites. The large amygdules (up to several mm, and 15% of the rock by volume) were filled by radiating bunches of epidote, large crystals of calcite, and chlorite plates (in that order of crystallization).

In structurally lower, stratigraphically higher and more westerly thrust sheets the Happy Creek is more difficult to subdivide. The epiclastic horizons (such as member C) are missing, and the thrusting, pervasive hydrothermal alteration and sill emplacement obscure the original volcanic lithologic distinctions.

The contact between members A and member E is found at the base of the thrust sheet below and to the west of the lowermost slice containing the Happy Creek - Bliss Canyon contact. Member E is characterized throughout this area by green and grey fine-grained plagioclase porphyry andesite and associated breccias. The andesite has a trachytic, very fine-grained groundmass, and highly amygdular nature. There are also aphyric and coarse-grained facies, and hornblende is found (though not abundantly) as a phenocryst phase; augite and quartz were not noted. The flows are massive and thick. Interbedded sedimentary rocks are also present (again, not abundantly), consisting of a lens of brown volcanic conglomerate. There is about 460 m of member E; this can be determined as both the lower contact (with A) and upper contact (with G - F is missing this far east) are present in a single thrust-bounded interval.

Several very thin dacitic horizons are present within the thrust packages north of the Iron King Mine, and are placed within Member F. They are made up of blue-grey fine-grained plagioclase-phyric dacite and dark grey aphanitic glassy flows. Quartz and biotite are present as phenocryst phases in one light green facies. In addition, an extensively weathered crystal tuff occurs just north of the mine. The tuff has glass shards and angular plagioclase fragments in a siliceous and highly hematized matrix with botryoidal texture, and is put in member F. The Fe-enrichment is inferred to be caused by pronounced weathering, but could also be due to alteration associated with the near engulfment of the bed by several apophyses of the adjacent large diorite sill.

Member F is only about 30 m thick, and may (as in area 7) be a single eruptive unit, with a patchy distribution. F does seem to be missing in several localities on the western edge of the area and entirely further east. The presence of F in an area allows for easier differentiation of members E and G. Lacking this member, E and G are distinguished by the presence of diorite and other intrusive fragments, greater oxidation, better bedding, more abundant epiclastic lenses, higher vesicularity, greater variability and greater proportion of mafics in member G.

In the structurally lowest and stratigraphically highest thrusts, on the western part of the area, member G includes those strata overlying the dacite. The member is about 900 m thick in one thrust plate that contains the contacts with both the King Lear and with member F. Elsewhere it overlies member E directly. Basal chill zones or breccias and aphanitic tops with glassy fragments help delineate individual flows, which were observed to be 2 to 10 m thick and are relatively evenly bedded. The flows are highly vesicular and locally scoriaceous, particularly on top. The amygdules are filled by quartz, chlorite and calcite. In several flows, diorite xenoliths up to 2 cm in size were preserved in the process of being partially re-assimilated by the andesite host. Volcanic diamictites are a part of the section and have sub-rounded clasts, matrix-support, and are variable in composition, including intrusive debris. G is characterized by grey (and some green) plagioclase $\pm$ hornblende $\pm$ augite $\pm$ quartz porphyry to aphyric andesite flows. Typical assemblages are plagioclase + augite; plagioclase; plagioclase + hornblende; plagioclase + hornblende + quartz. The plagioclase is andesine (An_{35}) in the hornblende-bearing facies to labradorite (An_{58}) in the augite-bearing lithology. Chlorite is present as an alteration product of hornblende (along with calcite + Fe-oxide $\pm$ quartz), as a replacement for glass, and as a late-stage crystallization phase. The texture is commonly trachytic, or less commonly the phenocryst population can be randomly aligned. The groundmass is intersertal (glass or very small plagioclase, mafic and Fe-oxide grains in-between phenocrysts).

Epiclastic horizons are also quite significant and widespread, though thin and laterally discontinuous. This facies includes red sedimentary breccias and red sedimentary conglomerates and locally grade into a very coarse-grained red volcanic arkose. The breccias have sub-angular clasts of green plagioclase porphyry andesite in a framework with an arenitic matrix. The detritus in the conglomerates includes red and green plagioclase and hornblende porphyry andesite, aphyric and scoriaceous flow facies, diorite, and monzonite, and is similarly in clast-support with a sandy matrix. Diorite clasts and xenoliths are notable features of member G, and imply extrusion after or during the emplacement of at least some of the diorite bodies. Oxidation and reddening is a common feature in the flows, breccias and diamictites as well as in the sedimentary facies. The red color is caused by hematite-staining. Epidote + quartz veins are widely developed, are present as clasts in the epiclastics, and are deformed by and predate thrusting. Magnetite veins and hydrothermal alteration are characteristic near the intrusives. The alteration is characterized by brown and cloudy plagioclase and replaced by sericite, the mafics completely going to chlorite, calcite and Fe-oxide, and the replacement of the groundmass by abundant Fe-oxides (including hematite and magnetite, locally up to 15% of the rock by volume).

Member G is conformably overlain by the red poorly-sorted, thick-bedded volcanic pebble and cobble conglomerates of the King Lear, with a Happy Creek source for the clasts. Various levels of the Happy Creek

Formation in the thrust belt (including the diorite sill) are overlain unconformably by syn-orogenic sediments of the King Lear as well. Member H is not preserved in this area; though the Fe-mineralized high-angle fault family is developed, uplift and exposure during thrusting have probably removed any strata of member H that were present. These high-angle faults cut both the diorite sills and member G.

<u>Age</u> The underlying Bliss Canyon Formation in this area is as young as upper middle to upper Norian in age based on conodonts, and also Middle to Late Triassic scleractinian corals, in a boulder from the base of the King Lear Formation where it overlies the Bliss Canyon directly. The boulder is inferred to come from the Bliss Canyon. The unlithified carbonate clasts caught up in the basal augite-porphyry flows of the Happy Creek have also been dated on the basis of bivalves as Middle to Late Triassic in age, and by conodonts as being Norian in age. Member A and the base of the Happy Creek Formation, therefore, is upper Norian. In addition, the large diorite sills have a preliminary U/Pb zircon age of 170-175 m.y. and intrude members A, E, F and G, which therefore have to be Bathonian (middle Middle Jurassic) or older; xenoliths and clasts of the diorite are found in member G only, implying synchronicity of extrusion of member G and diorite intrusion. Member G is therefore probably the same age as the diorites, and belongs in the Bathonian, while A, E and F are upper Norian to Bajocian.

<u>Depositional Environment</u> The basal portion of member A is, in this area, distinct from the rest of the member. The glassy matrix, dark green

color and lack of oxidation, quenched scoriaceous xenoliths, and interbedding with marine sediments identical to those of the uppermost Bliss Canyon Formation together lead to the conclusion that the basal part of the flow sequence was laid down in a marine setting. The size of the vesicles (up to as much as 3 mm) and their relative abundance (averaging around 10% by volume) indicate a submarine depth of extrusion of several hundred meters or less (Moore, 1965; Jones, 1969). Judging by bed thicknesses, the flows ranged from thin, ropy and fluid pahoehoe to (more commonly) thick, viscous and blocky aa with breccias. The breccias are glassy auto-breccias intimately related to the blocky flows.

The origin of the green volcanic arkoses is less clear, though they are found at the base of the member almost everywhere in the range. Marine reworking and proximal deposition of debris off of the encroaching volcanic flows is probable in this area. The cross-lamination to lenticular bedding and cross-bedding (formed by ripples and dunes), sorting and lack of fines are features that indicate a high-energy and, therefore, probably shallow marine environment. The biomicrite xenoliths in some of the lowermost flows indicate that at some point the lava either ascended through or flowed over this carbonate facies; probably the latter because of the stratigraphic restriction of these xenoliths to the very base of the member and the lack of baking, which would not be expected if the limestone was involved only in the autobreccia zone of a flow. The carbonate was deformed when entrained while still soft and probably originated in a quiet open shelf or fore-slope setting. This judgement is based on the presence of both unabraded low-energy and abraded high-energy morphologies and normal marine taxa in the skeletal grains, the fine lamination, and presence of a muddy matrix.

The middle and upper part of member A is characterized by less glassy matrix and auto-breccia, local oxidation and the presence of fluvial conglomerate lenses (channel cut-and-fill). This part of the section was subaerially deposited, unlike the shallow marine basal portion. The amygdules are locally very large and abundant volumetrically, supporting this conclusion. As elsewhere in the Jacksons, the member becomes less mafic upwards - hornblende replacing augite as the main mafic phase, and with augite and not hornblende likely to be in disequilibrium with the groundmass.

Member E is not as full of environmental indicators. It is composed of subaerial andesite flows and auto-breccias (as indicated by the flow textures), and rare fluvial conglomerate lenses.

The dacite member (F) is composed of thick blocky dacite flow rock and an explosive crystal tuff (identified as such by the glass shards, angular broken plagioclase grains, and glassy matrix). The shards in the tuff are long, thin and curved - a morphology that is most likely caused by a phreatomagmatic explosive event (Cas and Wright, 1987). The broken grains often have selvages of glass, another characteristic of a pyroclastic origin. The massive nature, thinness and patchy distribution of the dacite flow facies argue for a single subaerial flow and pyroclastic event.

The overlying andesitic succession belongs to member G, and has many features that help distinguish it from member E. Member G was extruded in a subaerial environment. This is indicated by the flow morphology (good bedding with glassy bottoms or basal breccias, massive flow interiors, and scoriaceous tops), porphyritic textures with fine grained groundmass, and the reddish weathering seen on top of many of the beds. The flows are blocky in nature, as flow contacts are irregular and are characterized by the presence of breccias - features ropy, phoehoe flows would not likely have (and these latter flows would also be thinner). The interbedded fluvial red beds (the clast-support, arenitic matrix conglomerate and arkose), debris flows (angular clasts floating in an arenitic matrix) and lahars (the diamictites with support by a muddy volcanic matrix) are also subaerial in origin, as interpreted elsewhere in the range. The diorite xenoliths have been reworked from the flows into the auto-breccias and as detrital clasts in the debris-flows, lahars and fluvial conglomerates.

Some of the hydrothermal alteration was quite early, as epidote clasts are present in the conglomerates. This fits the observation that the diorite sills in particular seem to be genetically related to the hydrothermal alteration (which increases in degree in direct ratio to the proximity to the intrusives), and that these diorites both intrude the andesites and are present as foreign fragments within them.

The basal King Lear (which is composed of coarse epiclastics with a source entirely in the Happy Creek) in some places overlies member G of the Happy Creek conformably, and without the appearance of a long hiatus or of pronounced weathering. Elsewhere, within the thrust domain, it lies on a highly weathered and uplifted part of the Happy Creek.

Depositional and Volcanic Framework for the Happy Creek Formation

The Happy Creek Formation is divisible into eight distinct informal members, designated A through H, and distinguishable on the basis of lithology and stratigraphic position. Some of the members are areally restricted (particularly B, D and H). Thickness changes are pronounced in all members (see the stratigraphic correlation, fence and block diagrams, fig. 29, 30, 31).

The depositional contact of member A with the underlying Bliss Canyon Formation is present almost everywhere in the range. This contact is conformable and gradational, with the uppermost strata of the Bliss Canyon overlain by a rather uniform thickness of 30 m of green, immature volcanic arkose, conglomerate, and breccia (coarsening-upwards). The epiclastics are immediately overlain by thick, blocky augite-phyric basaltic andesite. This basal interval of member A is remarkably constant over the range, despite the variability in depositional facies of the Bliss Canyon Formation. On the east side of the range the lower member of the Bliss Canyon (shallow basinal argillite and turbiditic facies) underlies the Happy Creek, while on the west side the upper member (littoral facies quartzose and cherty arenite and conglomerate) is the subjacent unit.

The evidence of abundant carbonate facies xenoliths (which were entrained while not completely lithified) in the basal flows and breccias of the Happy Creek also indicates the middle member of the Bliss Canyon was directly overlapped by member A. This relationship is, however, not actually exposed.

On the east side of the range, member A was deposited as blocky flows and flow-breccias in a shallow submarine environment, judging by the nature of the vesicularity and the interbedded shallow basinal sediments

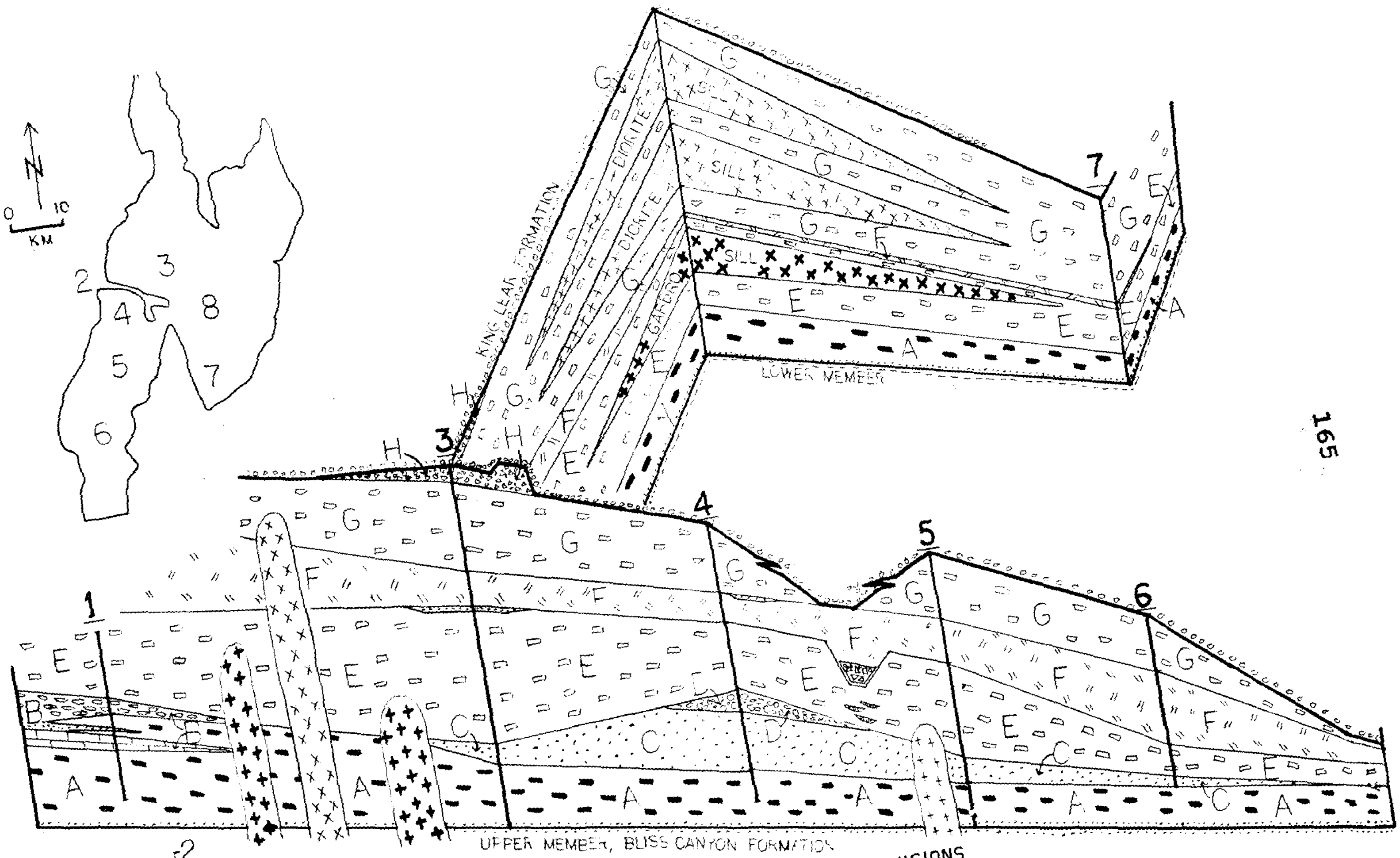


FIG. 29 FENCE DIAGRAM DISPLAYING THE STRATIGRAPHY AND DIVISIONS OF THE HAPPY CREEK FORMATION AND RELATED INTRUSIVE BODIES, JACKSON MOUNTAINS, NW NEVADA.

FIG. 30 Stratigraphic columns and correlation in the Happy Creek Formation, Jackson Mountains.

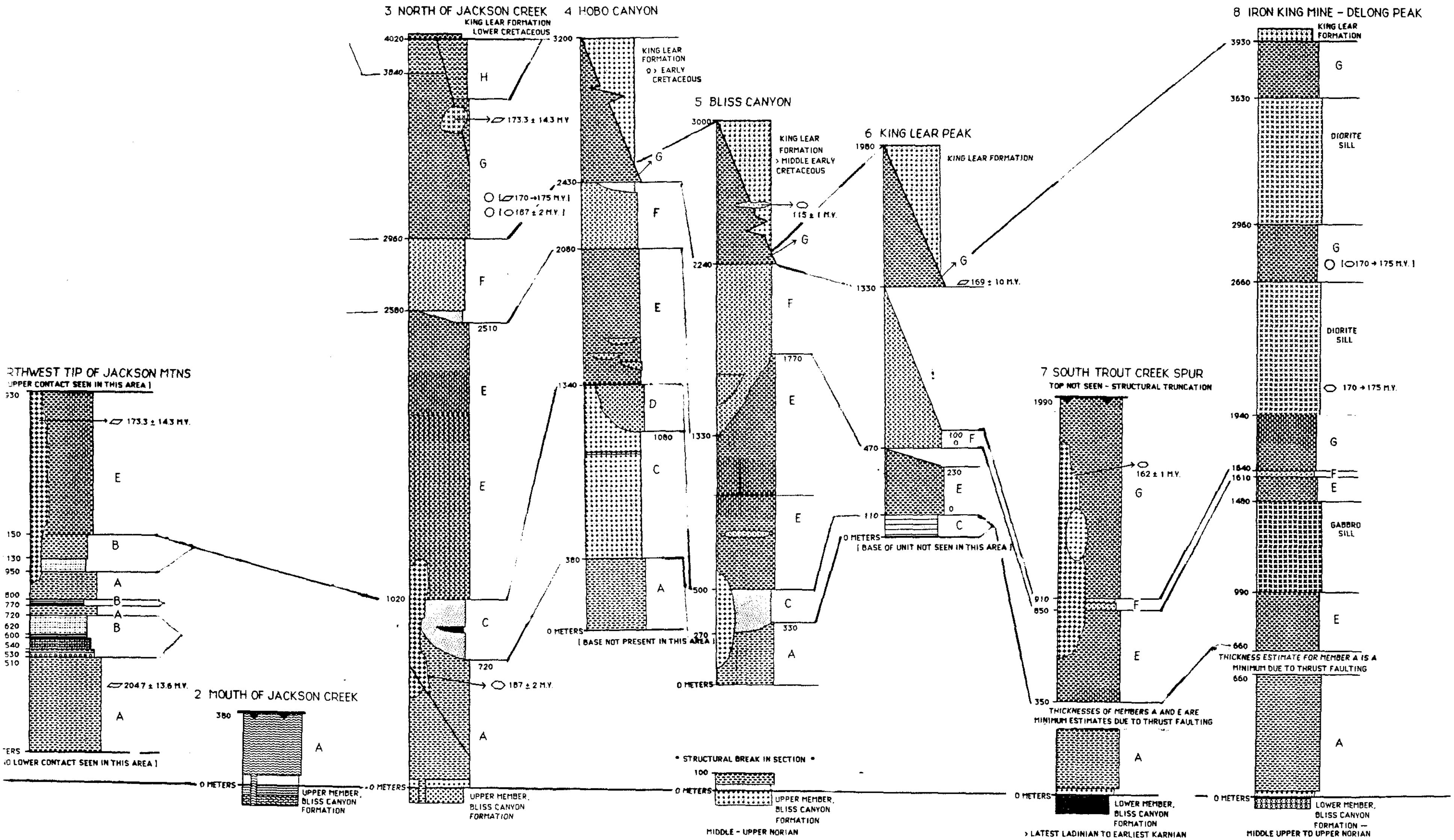
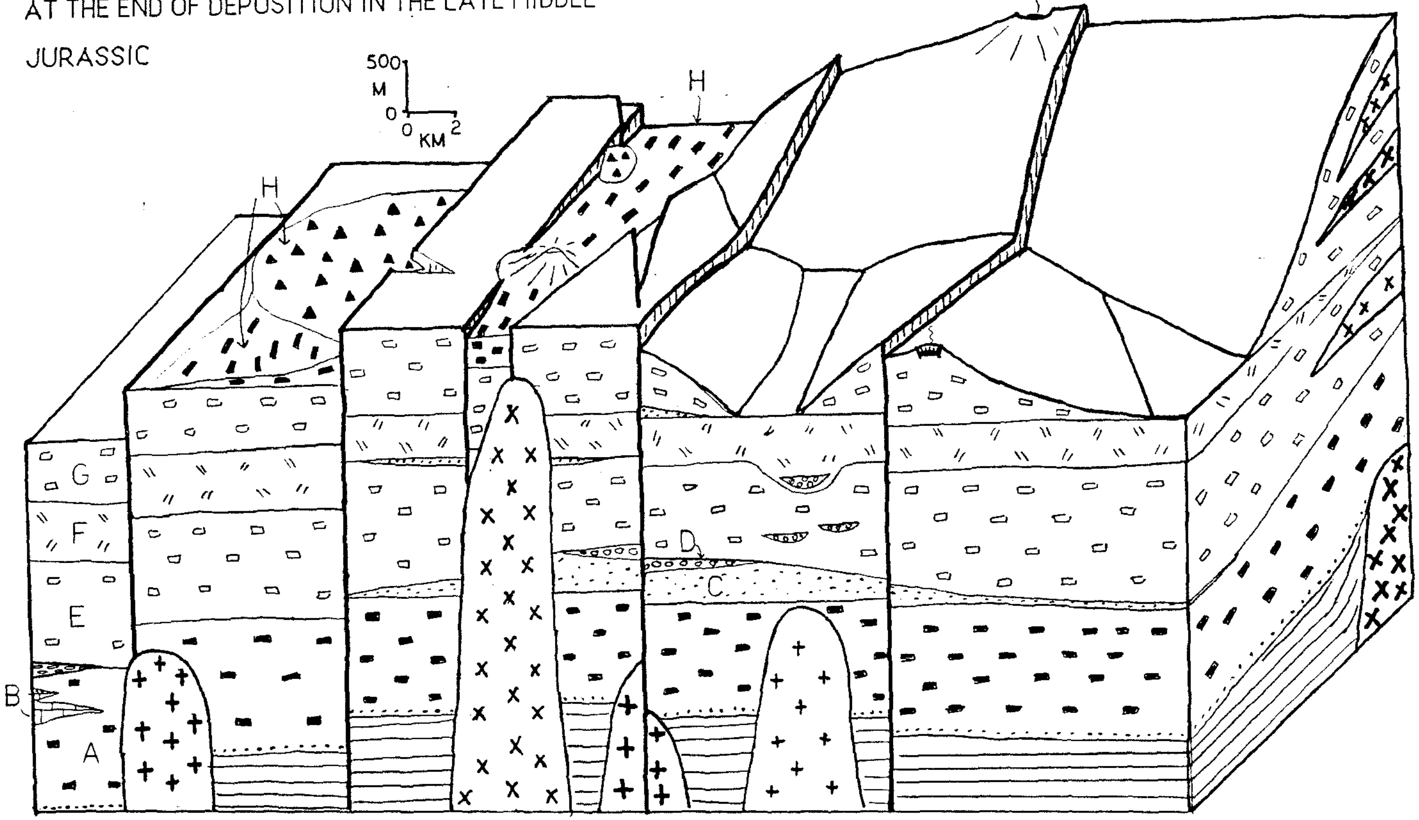


FIG. 31

DEPOSTIONAL AND VOLCANIC FRAMEWORK OF THE HAPPY CREEK FORMATION, JACKSON MOUNTAINS AT THE END OF DEPOSITION IN THE LATE MIDDLE JURASSIC



(see the section on the upper member of the Bliss Canyon Formation), and by the glassy groundmass and the breccia textures within the flows. In the northernmost exposure of the range, the environment for member A was subtidal to intertidal marine throughout, as the flows are interbedded with the sedimentary facies of member B (discussed below). On the west side of the range, it is not as clear what the depositional environment was because of metamorphic and tectonic overprinting, but a very shallow marine to subaerial origin seems probable - the basal epiclastic horizon is probably of a braided fluvial origin, and the flows have pillows and interstratified marine sediments (recrystallized calcarenite and laminated tuffaceous chert). The middle and upper portions of the member (particularly in Trout Creek Spur, where exposures of this interval are best) are subaerial in origin, with a less glassy matrix, local subaerial weathering and lacking both pillows and hyaloclastite debris. The member is fairly constant in petrology throughout the range, with no significant changes in character or thickness from area to area. The abundance of augite decreases and that of hornblende increases upwards in the section. At the base hornblende is in disequilibrium, but at the top it is the augite that reacts with the matrix.

Member B is seen only at the northern tip of the range, where it interfingers with member A and is overlain by member E. The unit is diagnostically shallow marine to intertidal. Facies include carbonate barrier bar (oolite, crinoidal biosparite), intertidal flat (bimodally cross-laminated siltstone, laminated volcanic arenite), and supratidal zone (flat-pebble conglomerate and stromatolitic mats). The detritus is entirely composed of glassy and immature volcanic debris, some of it of hyaloclastite origin. This and the common peperite textures in member B related to interfingers and dikes of member A indicate contemporaneity of depositon of the basaltic andesite flows and the very shallow marine facies. The epiclastics of member B coarsen and become higher-energy in facies upwards, due perhaps to the loss of the offshore bar (seen at the base of B only), to more effusive volcanism and concomitantly less reworking of the epiclastic facies, or to regional regression and offlap. The contact between members A and B seems to reflect the interaction of subaerial volcanic eruptive and very shallow marine environments at the paleo-coastline. This paleo-coastline probably ran roughly east-west and was stationary, with the volcanic pile to the south and the marine realm to the north, throughout the deposition of both members A and B.

Member C is laterally extensive through the range but does disappear in the extreme north, south and east. It reaches its thickest extent in the Hobo Canyon area. In this locale it consists of highly immature volcanic conglomerate, arkose and wacke laid down in a middle to upper alluvial fan setting by sheet flood, debris flow and braided fluvial processes. Away from this area in all directions, the unit thins and becomes much finer-grained and somewhat more mature, dominated by volcanic arenite and argillite with little or no conglomerate. These areas reflect distal to interfan alluvial plain and perhaps lacustrine settings, with deposition by braided fluvial and sheet flood processes.

Member D is found at the top of the thickest portion of member C, and is composed of fluvial volcanic conglomerates in a large lenticular body. Member D originated in a braided fluvial channel within the upper alluvial fan facies of member C. The entire system represented by C and

D is a very large and extensive alluvial fan with a source inferred to the east, and an entirely volcanic provenance. The subaerial nature is further documented by the ubiquitous hematitic cement. Volcanism was active during this epiclastic phase, as intraformational siliceous volcanic clasts and andesitic flows are present in the unit, and the clast population includes lithologies found in member E.

Member E, overlying members A to D, is composed of plagioclase ± hornblende andesite flows, flow breccias, lahars, and lenses of fluvial red conglomerates. It is distinguished from member A by lighter color, by the lack of augite and presence of hornblende, by containing fluvial epiclastic interbeds, by having a higher degree of vesicularity, and by more abundant oxidation of the flows. Hornblende becomes less abundant upwards in the section. The member is subaerial in origin, and displays some variation in thickness across the range. The thickest occurrences are in the area between Jackson and Mary Sloan Creeks, where as much as 1500 m of strata are attributable to the member; the unit thins somewhat to the south and east, but there is never less than about 500 m of section except in the extreme south of the range where it can be as little as several hundred meters thick. E is also thinner above the thickest portions of members C and D, due perhaps to positive topographic relief on this alluvial fan system. Within E, on the east side of the range only, are several local diorite sills of the Early Mesozoic Intrusive Suite (on the order of a hundred meters thick) and, more notably, a very thick and laterally extensive (up to about 500 m thick) gabbro sill. No clasts or xenoliths of these or other intrusive facies were found in the member, which implies the major intrusive phases came later.

Member F is characterized by lighter colored, bluish-grey siliceous flows, flow breccias, lahars, tuff and epiclastics. The flows contain quartz, plagioclase, alkali feldspar, biotite and hornblende. The nature of the very thick, blocky flows and the related epiclastics and common oxidation of the beds indicate, once again, a subaerial setting. The member is somewhat constant in thickness (400 to 500 m) and nature on the western side of the range, but pinches out altogether from a single flow unit within the eastern side, and also decreases in thickness greatly in the extreme south, near Navajo Peak. Another exception to the constant thickness is the Bliss Canyon area, where a deep channel (as much as 400 m deep) was cut into the underlying strata of member E and later filled by dacitic fluvial conglomerates and dacite flows. This channel is stratigraphically just above several small fluvial lenses within member E and is near the thickest and coarsest-grained areas of members C and D, and thus appears to represent part of a long-lived drainage system. F is also somewhat thicker in the King Lear Peak area where E is thinning southwards; F itself thins just a bit further south as well. To the east only a few tens of meter of section was present, and may well have been a single flow unit. In several places F is bounded on upper and lower contacts by laterally extensive lenses of volcanic arkose and conglomerate with an origin in a braided fluival setting. This may indicate that some interval of time existed between the various volcanic phases of members E, F and G. In addition, in the area of the channel mentioned earlier, member G is missing and F is directly overlain by the King Lear Formation with pronounced oxidation, subaerial weathering, and caliche formation along the unconformity. This indicates a long period of nondeposition.

Member G marks a reversal of the trend towards more highly evolved and siliceous volcanism witnessed in members A, E and F. The member contains plagioclase ± hornblende ± augite andesite flows, flow-breccias, lahars and conglomeratic lenses. The flows are blocky and subaerial, with basal breccias, glassy and scoriaceous tops and oxidation zones on the tops of most flow units. One conglomerate lens was fluvial; others are more likely to be debris flows and probably grade into the lahars. The abundance of the latter two facies and the thickness variations indicate substantial constructional relief existed during the extrusion of the member. G occupies several distinct extrusive centers - one around King Lear Peak and another north of Jackson Creek, both with a maximum 800 m of section preserved and thinning to zero in the area of Bliss Canyon and south of King Lear Peak. Depositional paleoslope was on the order of 10°to 15°. A third such center is found in Trout Creek Spur in the Iron King Mine - Delong Peak area, with a thickness of volcanic strata of about 900 m. This latter center is also characterized by the intrusion of several very thick (600 to 700 m) diorite sills into members E, F and G. These sills thin, interfinger with the volcanic strata and disappear to the west, north and south. On the western side of the range the more deeply exposed northern eruptive center is characterized by the intrusion of shallow-level stocks of diorite and monzonite (and by a north-trending dike swarm). Areas between such volcanic centers lack intrusives and have thinner volcanic sections. These intrusives are probably comagmatic with the volcanics of member G, as clasts and xenoliths of identical intrusive facies (as well as intruding G and the lower units) are found in the member (and also in H) but not in the lower units. G interfingers directly with the basal red volcanic conglomerates of the King Lear Formation in several places, and typically is not as highly weathered as F below the King Lear.

Member H is restricted in occurrence to the area north of Jackson Creek, where it is intimately related to the episode of high-angle east-west faulting. Field relations show this system of faults became active during the latest phase of Happy Creek volcanism. They cross-cut member G but have the structural basins filled to partially filled by ongoing andesitic volcanism similar to that of member G and by debris flows and other very proximal alluvial facies. H fills and overlaps several of the basins (in an area where, perhaps not coincidentally, jasperization and hydrothermal magnetite and hematite replacement is at its greatest), and occurs as the floor on another, pinching out to the north and east. H is probably laterally equivalent to the uppermost strata of G in areas where the faulting did not occur (or was initiated later). In some areas to the south G does contain Fe-mineralized debris, but a distinct set of facies assignable to H is not present. Some of the sedimentary facies in H are identical to those in the base of the overlying King Lear Formation, when the faults were still active during the deposition of the latter unit. H and the lowermost King Lear may also be lateral equivalents, with the latter distinguished by the lack of andesite flows and much greater lateral continuity.

The *timing* of deposition of the various members within the Happy Creek Formation can be fairly well constrained. Member A is upper Norian at the base, as it is interbedded with sedimentary rocks of that age in the uppermost Bliss Canyon Formation. A also has a Rb/Sr age placing it in the Norian to Lower Jurassic (204.7 $\pm$ 13.6 m.y.). The King Lear Formation overlying member F and interfingering with and overlying members G and H has fossils as old as middle Early Cretaceous fairly high up in the section; on those grounds the Happy Creek can be no younger than lower Early Cretaceous. A Rb/Sr age on flows from member G gave an age of late Middle to early Late Jurassic (169 $\pm$ 10 m.y.). Members A through E are intruded by a granodiorite body with a U/Pb age of 187 $\pm$ 2 m.y. Diorite sills intruding member G have a U/Pb age of 170 to 175 m.y.; the latter two intrusive facies are also present as xenoliths and clasts within G and H. Strata as high as member G and the faults active during the deposition of H are cut by intrusives with a Rb/Sr age within the Middle Jurassic (173.3 $\pm$ 14.3 m.y.).

Members A through E are upper Norian to as young as Toarcian. Member F is also probably Toarcian or older (based on the absence of intrusive xenoliths and clasts) but could be as young as Bathonian. Member G could be as old as Aalenian to Bathonian in the northern half of the range, and probably is within the younger end of that range due to the cross-cutting relationships with the diorite intrusives. Member H, and member G to the south, could be as young as Callovian, but can be no older than Aalenian, and are probably Bathonian.

The first phase of volcanic activity started in the latest Triassic and continued through the Early Jurassic, and was characterized by augite-phyric basaltic andesite. A large subaerial volcanic center was built up during this time period, and first-cycle epiclastic sedimentation during lulls in volcanism played a significant role during its construction. The nature of the volcanism became clearly less mafic and more siliceous with time, until a second phase of more mafic extrusive activity began fairly abruptly in the Middle Jurassic, with several distinct volcanic constructal piles. This later phase of activity was also accompanied by activity on a generation of high-angle faults with a substantial dip-slip component, forming structural basins that were partly filled by the latest stages of Happy Creek volcanism and epiclastic redistribution. A changeover to dominantly epiclastic and molasse sedimentation with subordinate volcanism took place at this point, as recorded by the King Lear Formation. The long period of time represented by the Happy Creek Formation (on the order of 40 m.y.) indicates an unusually stable setting within the arc. Presumably, the tectonic environment did not undergo major evolutionary changes during this period.

CHAPTER FIVE:

UPPER JURASSIC TO LOWER CRETACEOUS STRATIGRAPHY OF THE JACKSON MOUNTAINS: THE KING LEAR FORMATION

The King Lear Formation, as first defined by Willden (1958, 1963, 1964), included the sedimentary strata thought to unconformably overlie his Happy Creek volcanic series (generally equivalent to the formation of the same name defined here in chapter four). As redefined here, the King Lear Formation includes coeval silicic volcanic strata and related very shallow-level silicic intrusive facies (some of which were previously thought to be Tertiary, correlated to silicic volcanics of that age found in Humboldt County - Willden, 1963). It also includes the Pansy Lee Conglomerate on the northeast flank of King Lear Peak and on the west side of Deer Creek Peak (earlier thought to be Cretaceous or Tertiary, again based on regional correlation - Willden, 1963).

The formation occupies a topographically low strip down the middle portion of the range, within the central autochthonous structural buttress. In the northern and central parts of the range it is imbricated and folded by the west-vergent, eastern thrust system. In the southern part of the map area, it has been deformed by the east-vergent, western thrust belt. Throughout the range it has been affected by syndepositional high-angle faulting, and by the later reactivation of these faults as tears during thrusting.

The contact with the underlying Happy Creek Formation is well-preserved throughout the range, and varies in nature, exhibiting these relationships: (1) conformable and interfingering; (2) disconformable with a probable hiatus; (3) a buttress unconformity; (4) juxtaposition of the two units by the syndepositional high-angle faults; and (5) angular unconformity to nonconformity. This complexity reflects the various stages in the tectonic and stratigraphic evolution of the region during deposition of the uppermost Happy Creek and throughout the deposition of the King Lear Formation. In one area east of Buff Peak in the northern part of the range the King Lear Formation also overlies thrust-faulted strata of the lower member of the Bliss Canyon Formation with angular unconformity.

No one section of the King Lear Formation among the many examined represents the range of highly laterally and vertically variable facies found and the complex syndeformational sedimentation patterns. All sections have also experienced some postdepositional structural history. Stratigraphy is, however, quite well exposed overall and the facies patterns and history of the formation can be examined in considerable detail. The formation was named (Willden, 1958) for outcrops, east of King Lear Peak, of the basal portion of the unit. Though exposure is very good there, only a limited range of the facies found in the formation are present.

The King Lear is unmetamorphosed with the exception of exposures east of Navajo Peak in the southern part of the field area. Here, a gabbroic to granitic pluton of the Early Mesozoic Intrusive Suite intrudes and contact metamorphoses the basal King Lear. The only other intrusive bodies within the unit are related to the syndepositional silicic volcanic centers that can be distinguished within the formation, and which will be discussed with it.

The King Lear Formation, like other pre-Cenozoic units in the range, is overlain with angular unconformity by Tertiary basalt and rhyolite flows and sediments, and by Quaternary alluvium. The latter unit includes stream, alluvial, landslide and Lake Lahontan shoreline deposits.

A dozen stratigraphic columns (the majority measured in the field by means of Jacobs staff) will be presented, from north to south in the range. The localities include (see fig. 32): (1) Happy Creek; (2) Parrot Peak; (3) Mary Sloan Creek; (4) north Jackson Creek; (5) Iron King Mine; (6) DeLong Peak; (7) Trout Creek; (8) Jackson Creek; (9) south Jackson Creek; (10) Bliss Canyon; (11) Clover Creek; and (12) King Lear Peak. The rhyolitic volcanic dome complexes in areas (1), (6) and (8) and the dacitic laccolith system in areas (9), (10) and (11) will be included with the stratigraphic discussion of those localities.

The King Lear is divided into six informally defined members for purposes of mapping and analysis. From oldest to youngest (taken from first stratigraphic occurrence, as members interdigitate and are time-transgressive) these are: (I) basal volcaniclastic red beds, including conglomerate, arenite, breccia and siltstone, plus rare dacite and basaltic andesite flows; (J) green cherty/lithic conglomerate, arenite, siltstone, shale and fossiliferous micrite; (K) upper volcaniclastic red beds consisting of conglomerate, volcanic arenite, siltstone and shale; (L) rhyolite flows, protrusive domes, and related pyroclastic (crystal tuff to welded ignimbrite) and epiclastic (breccia, conglomerate and arenite) rocks; (M) dacitic flows, plug, sills and related pyroclastic

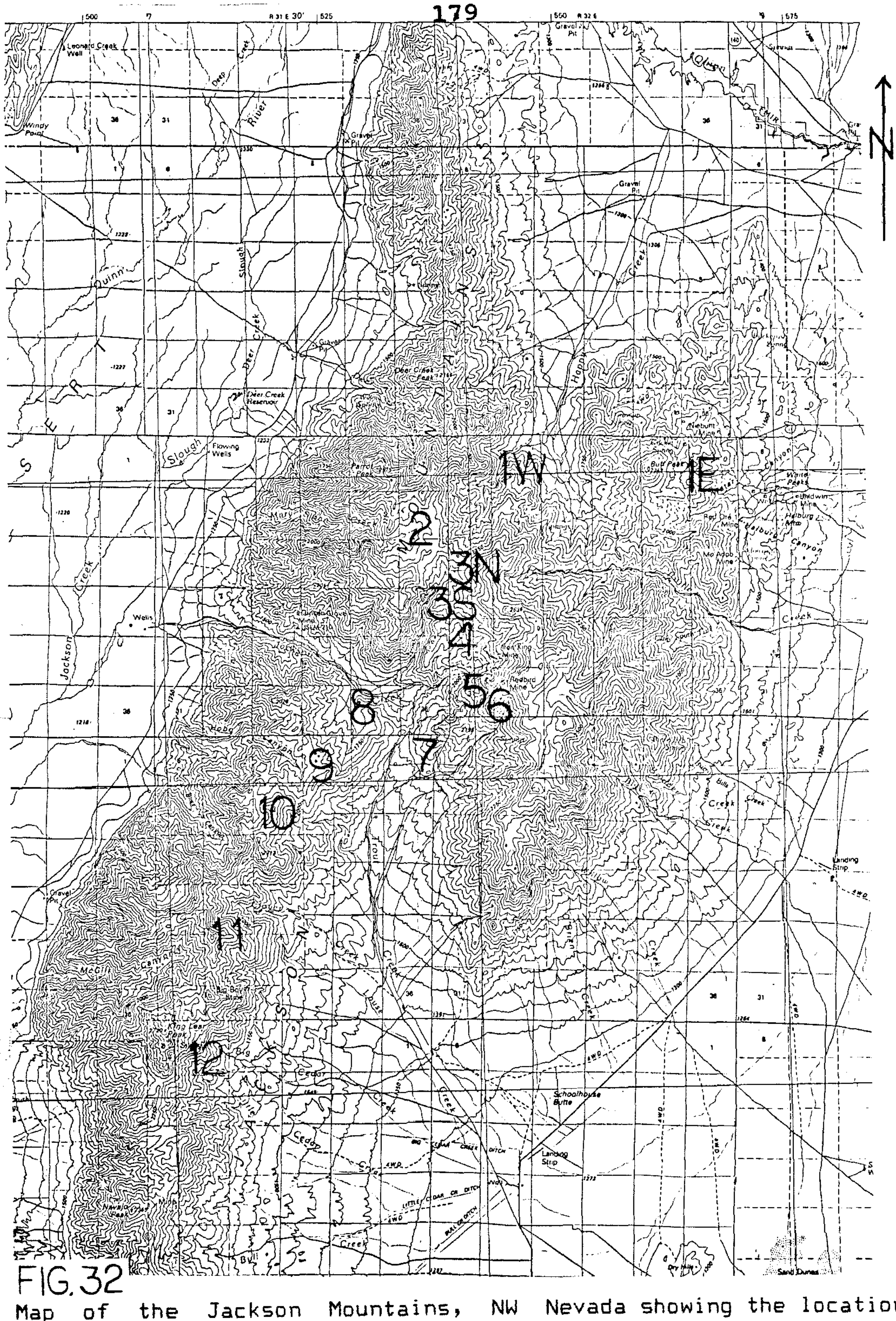


FIG. 32

Map of the Jackson Mountains, NW Nevada showing the location of the stratigraphic sections presented from the King Lear Formation. (1) Happy Creek; (2) Parrot Peak; (3) Mary Sloan Creek; (4) north Jackson Creek; (5) Iron King Mine; (6) DeLong Peak; (7) Trout Creek; (8) Jackson Creek; (9) south Jackson Creek; (10) Bliss Canyon; (11) Clover Creek; (12) King Lear Peak.

(crystal tuff) and epiclastic (lahar, arenite, conglomerate and breccia) rocks, plus rare fossiliferous micrite; and (N) red volcaniclastic breccia, conglomerate, arenite and siltstone intimately related to thrust faulting. There is, in addition, a rhyolitic tuff which is here proposed to have the status of a bed (IUGS, 1976) and designated the Jackson Tuff (as it is well exposed in that stream drainage). The Jackson Tuff provides an excellent time horizon across the entire range, and is found in almost every section. This bed occurs in members I and lower J. The base of the King Lear is then put at the bottom of the first epiclastic horizon above the massive andesitic flow rocks and related facies of the Happy Creek Formation (members F, G or H). The top of the King Lear is erosional, with Neogene sandstone, conglomerate and basalt flows unconformably situated above. The members in the King Lear were labelled in sequence after those of the Happy Creek Formation to avoid confusion with members of the same name in different formations.

Stratigraphic thicknesses, facies relationships and the syndepositional structural and volcanic relationships, though highly variable, are well-characterized in the Jackson Mountains and permit detailed reconstruction of a highly complex history for the King Lear Formation.

For each section (as with earlier chapters) the analysis will include: (1) description of the field and petrographic relationships, including lithology, volcanic and sedimentary features, the nature of syndepositional fault activity, and sedimentary provenance; (2) age data (biostratigraphic and geochronologic); and (3) the paleoenvironmental, paleogeographic, volcanic, temporal and structural framework within which that section of the King Lear was deposited. After this, paleocurrent data will be presented, and then the history of the King Lear Formation as a whole will be synthesized. The complex nature of evolution of the unit has many implications for regional correlation and tectonic evolution.

(1) Happy Creek

This section covers a large area, including the Happy Creek drainage and the extreme northeastern spur of the range. The strata of the King Lear Formation conformably overlie member G of the Happy Creek Formation within the Happy Creek drainage. On the eastern edge of the area the formation and related hypabyssal intrusive facies overlie and intrude a thrust sequence of the Happy Creek and Bliss Canyon Formations. The King Lear has itself been thrust-faulted and folded, and overlain with angular unconformity by channel-fill unconsolidated boulder conglomerate, and by laterally extensive basalt and rhyolite flows of Miocene age (Willden, 1963), as well as by Quaternary sediments. Just north of Deer Creek Peak, the Tertiary unconformity cuts down into the level of the Happy Creek and no more outcrop of King Lear is seen to the north. See fig. 33, 34 for stratigraphic columns (both western and eastern).

<u>Stratigraphy</u> The lowermost King Lear, member I in this area, overlies member G of the Happy Creek, except on the extreme east within the thrust belt (in that region higher stratigraphic levels of the King Lear-member J - overlap a thrusted sequence of Bliss Canyon and Happy Creek Formations). The contact between member I and the uppermost andesite

FIG. 33
Lithostratigraphic and other symbols used in the King Lear Formation stratigraphic sections.

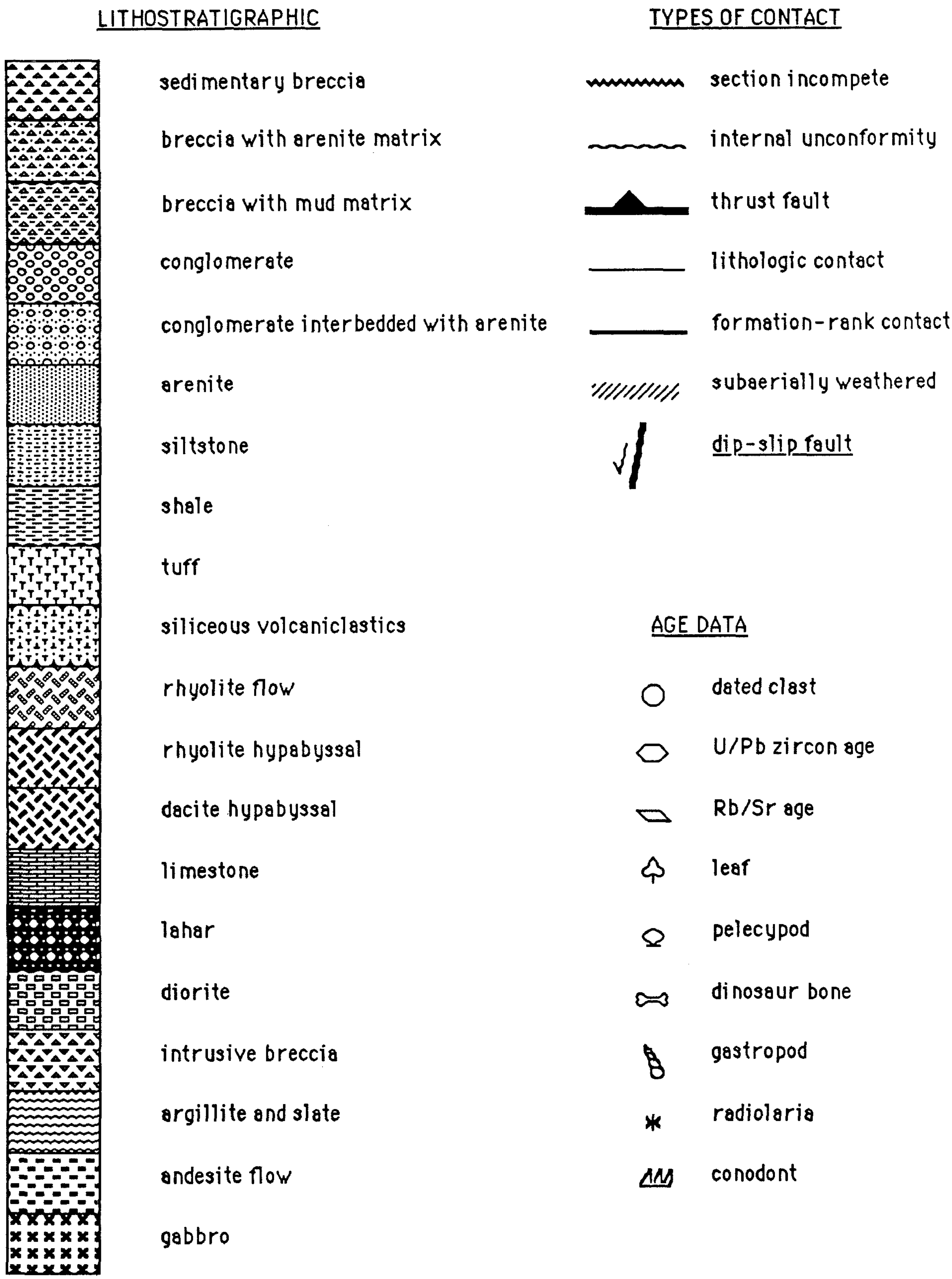
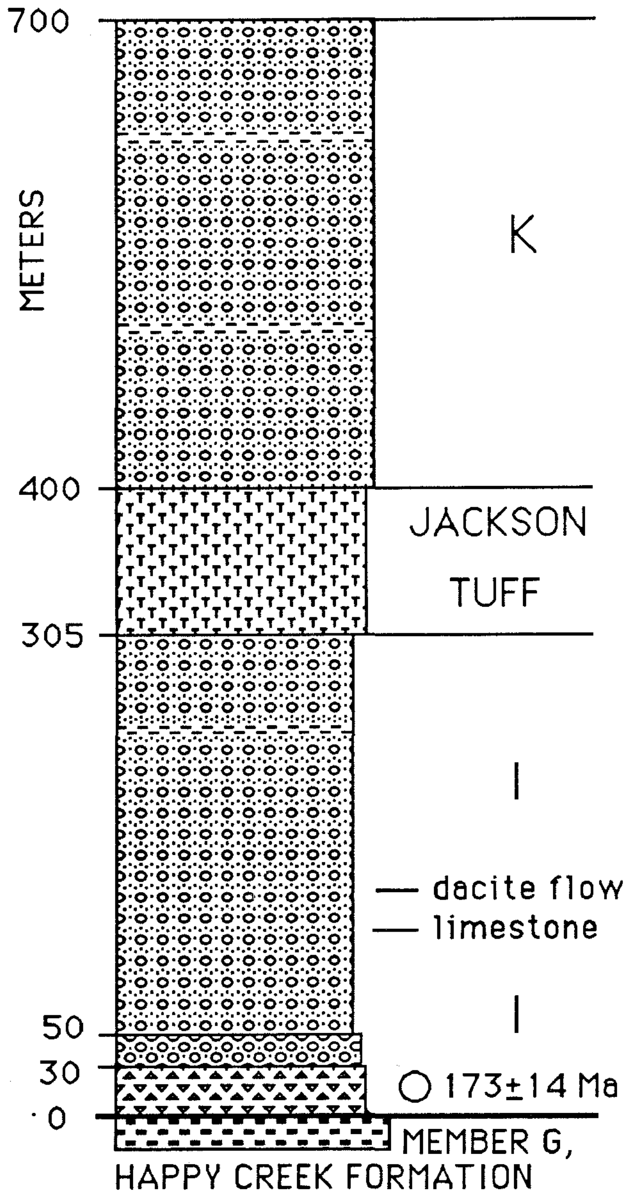


FIG. 34
1 HAPPY CREEK

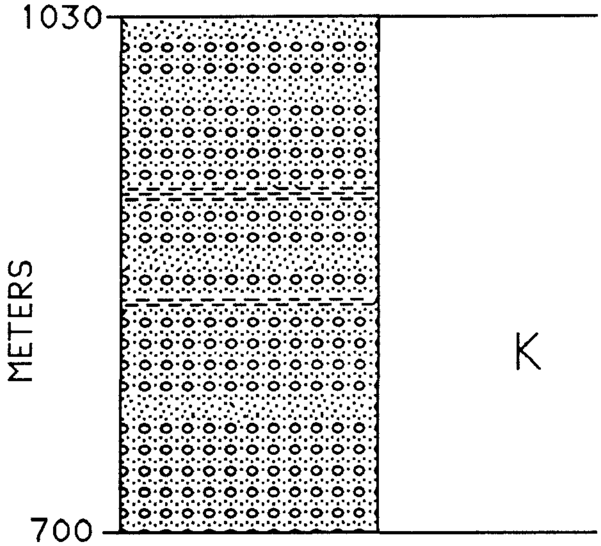
STRATIGRAPHY OF THE KING LEAR FORMATION
WESTERN SECTION



1 HAPPY CREEK (CONTINUED)

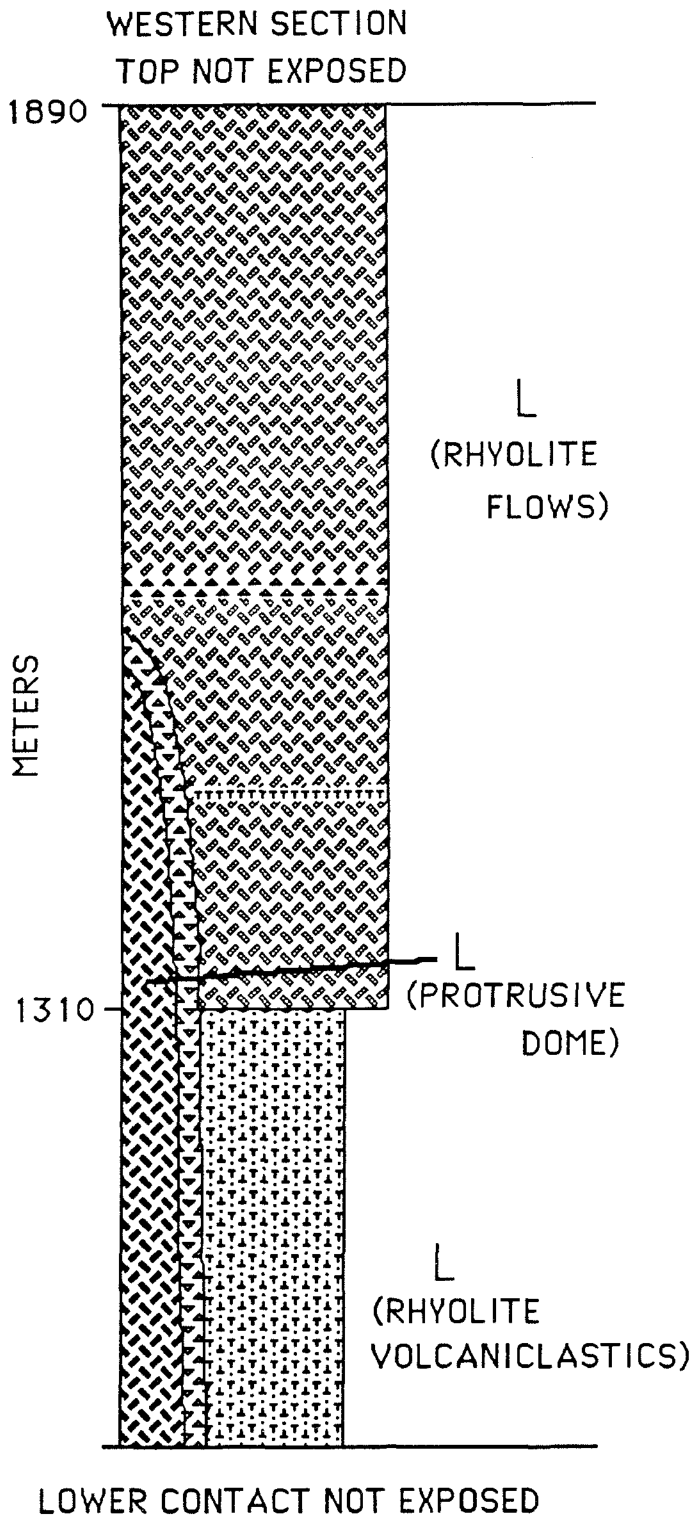
WESTERN SECTION

UPPER CONTACT NOT EXPOSED

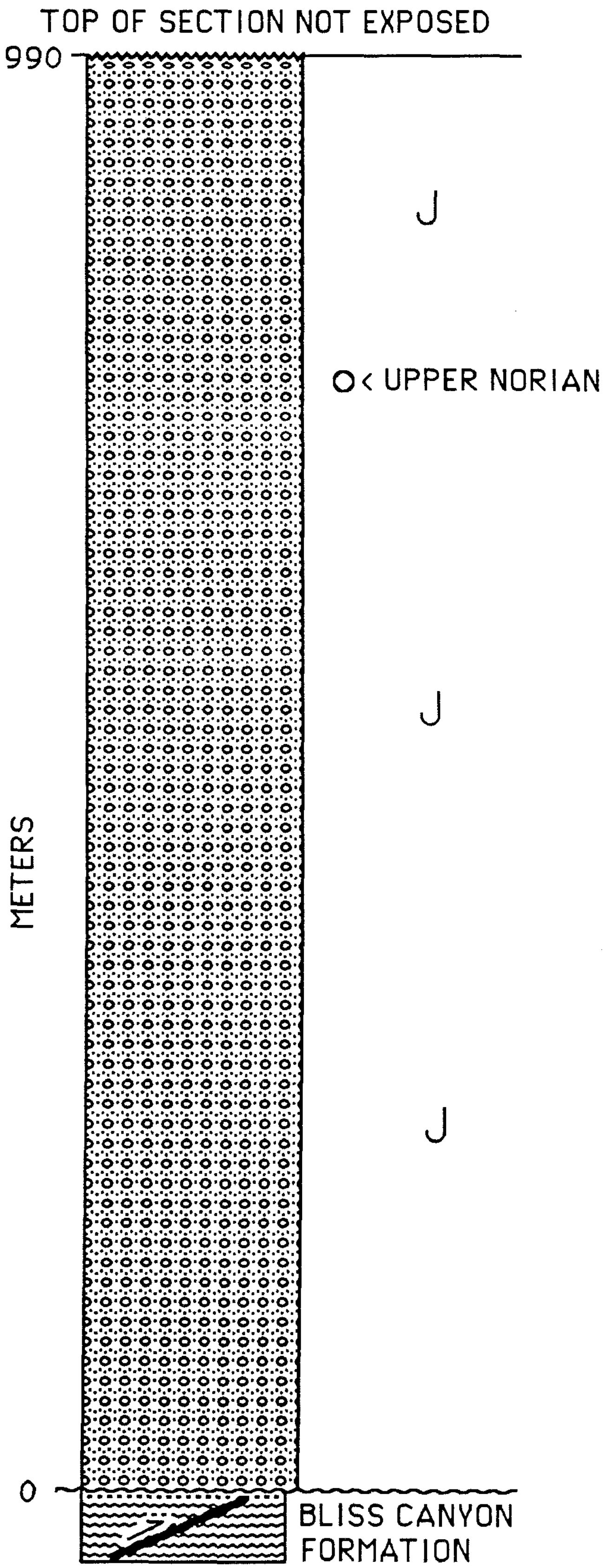


(CONFORMABLE ON THE JACKSON TUFF)

1 HAPPY CREEK (CONTINUED)



1 HAPPY CREEK (CONTINUED)
EASTERN SECTION



flow of Happy Creek member G is conformable with no evidence of a profound hiatus (such as the weathering and caliche development seen elsewhere). The bottom of member I, at the contact, is composed of massive, very thick-bedded red volcanogenic conglomerate, breccia and subordinate arenite. Some of the clasts are quite large, the deposits are very poorly sorted and there is matrix support. Upwards the member becomes somewhat thinner- and more distinctly bedded. There is clast support in the conglomerates, better sorting and rounding and a smaller maximum clast size, and cross-laminated arenite lenses. The provenance is entirely from within the Happy Creek. There is more volcanic arenite than conglomerate at this level, and interbedded thin red to green siltstone beds as well. The siltstone layers have lenticular cross-laminations, sand lenticules and large vertical burrows or rootlets. A thin dark grey micritic limestone, 2 m thick with stromatolitic mat laminations, is also found in this interval. A laterally continuous light grey aphyric dacite flow and breccia bed occurs at 128 m, and is 4 to 10 m thick.

At 305 m, after more red volcanic arenites and conglomerates, there is a 95 m thick massive white to very light pink or green rhyolitic tuff with abundant quartz and feldspar crystals - the Jackson Tuff. This same crystal tuff extends at least as far south as King Lear Peak (area 11).

Overlying the tuff and placed in member K is another 630 m of well-bedded red volcanic conglomerate, volcanic arenite and siltstone. The conglomerates are entirely volcanic in provenance and include only Happy Creek lithologies - no silicic volcanics such as those in member L of the King Lear are present. Also, little cherty or metasedimentary lithic debris is present. The beds exhibit clast-support with an arenitic matrix, channel scour and fill, cross-stratification, and normal grading. Some matrix support conglomerates are also present in the section. The arenites and siltstones occur as thin cross-laminated lenses within the conglomerates or alternatively as distinct several meter thick horizons, red or green in color. In the thicker occurrences, these finer-grained lithologies can have silicic paleosols (silcrete) developed upon them, with pronounced weathering, silicification and oxidation, rootlet casts and thin mottled to banded silcrete horizons. This whole sequence becomes finer-grained, thinner bedded and greener in color upwards, with more paleosol development exhibited. The conglomerates become entirely clast-support in texture with abundant channelling and cross-stratification, and are still entirely volcanic in provenance.

J does not crop out in the western part of this area. The contact of member K with the overlying member L is obscured by talus, so the conformity of the contact (or lack thereof) cannot be assessed. In the eastern part of the area, Member J does exist as a sequence of green chert-pebble lithic conglomerates and cherty lithic arenites and green siltstone, about 990 m thick. It overlies a sequence of Happy Creek beds (member A) and Bliss Canyon (lower member) beds, and unconformably overlaps a thrust fault, which cut and assembled the package. The provenance in this unit is entirely lithic (cherty and metasedimentary facies of the Bliss Canyon and McGill Canyon Formations); no volcanic lithologies are found in the clast population, either from the Happy Creek or from member L of the King Lear. Large boulders of light grey fossiliferous limestone (biomicrudite or floatstone, with scleractinian coral colonies, bryozoa, bivalves and fish teeth) are found and are not inferred to be very far from their source. These lithic conglomerates and arenites are chanelled and cross-bedded, with clast-support and thin interbeds of green siltstone and shale. The lithic and cherty clasts tend to be very rounded, and were reworked directly out of the older formations (in fact, care should be taken in interpreting the maturity and origin of this facies because of this factor; the coarse component of detritus in these lithologies is multicyclic). The sedimentary structures are identical to those of the upper part of member I and in K, despite the distinct color and provenance of the rocks. The member is cut by rhyolitic dikes, which are thought to be part of member L, but could also be Tertiary (though no demonstrably Tertiary silicic intrusives were found in this study, rhyolite flows of that age are present).

In the central part of the area, superjacent to the reworked andesitic epiclastics of member K, is a thick pile of rhyolitic volcanics assigned to member L. The lower portion of the unit, exposed in the core of an anticline, is composed of white to yellow rhyolitic breccia to angular arenite. These rocks are poorly consolidated with very well-developed low-angle wavy layering and trough cross-stratification and channelling (up to 1 m deep and 10 m across), or have planar stratification. In the areas of more planar stratification are thin (4-6 cm) red, oxidized zones. The detritus is pebble- to sand-sized, very angular and purely rhyolitic in provenance, with excellent sorting within individual strata. This volcaniclastic interval is at least 280 m thick.

The rhyolitic volcaniclastic succession is directly overlain by and interbedded at the very top with rhyolite flows having a total thickness of at least 580 m (the upper contact is not present). The individual flows are extremely thick and are characterized by pink, red, purple and gray coloration and the presence of large, clear quartz and feldspar phenocrysts. The bases, and sometimes the entire flows, are obsidian, black to red and pink in color. Extremely convolute and ductilely deformed siliceous flow-banding is also characteristic of the flows. Some associated glassy rhyolitic breccias are present. Spherulitic devitrification along discrete laminae (mm to cm scale) helps to delineate the flow-folds. Yellow epiclastic rhyolitic breccias and crumbly sand-sized volcaniclastic beds, though not abundant, are found in between some of the flows. These breccias have clayey matrix, thin bedding, poor sorting and, in some cases, are baked by the superjacent flow. Some thin interbedded tuffs, white with very fine-grained texture and very thin lamination, are also present in the sequence.

The rhyolite in thin section is composed of clear grey to tan glass with plastically deformed interlaminae of brown, dirty glass. The glass is in some instances full of very fine dark acicular microlites, well aligned parallel to the flow laminae in general, but overprinting it in detail. Also in some places the silicic glasses have perlitic cracking or have been overgrown by patches of tridymite with or without very fine-grained white mica (this is particularly abundant in the browner, dirtier glassy laminae). Andesine and sanidine are both present and possess perthitic textures, in addition to large euhedral but corroded and embayed quartz crystals, rare plates of biotite and hornblende granules. In some of the glassier units the phenocrysts are brittly fragmented and silicic volcanic rock fragments are also present; both tend to be confined to the dirtier, browner glass laminae. The top of the member is not preserved due to overthrusting and erosion.

The thrust plate structurally overlying member L contains a large intrusive rhyolite plug and related dikes. The plug has intruded a tightly folded interval of upper Bliss Canyon and lower Happy Creek strata, and injected dikes preferentially along the adjacent anticlinal axial planes. Other smaller intrusives cut the deformed pre-King Lear strata to the east. These irregularly shaped dikes are greenish-white and sugary-textured with quartz phenocrysts, and fluidal banding isoclinally folded or parallel to the margins. The core of the plug (making up Buff Peak) is pink, red and light grey rhyolite with fluid-banding, intensely contorted and very similar to the rhyolite flows of member L. The plug is about 900 by 1200 m, elliptical and elongate in a northeast-southwest direction. The fluidal banding is generally concentrically oriented, but is highly variable on an outcrop scale. This massive rhyolite has an outer shell of bleached and devitrified breccia composed of large fragments of the same rhyolite, with black and brown obsidian veins (with quartz and felspar phenocrysts) cutting the breccia envelope, which is 200 to 300 m thick. This whole sequence, though cutting the more easterly thrust planes, was itself cut by the westernmost and structurally lowest thrust fault, indicating perhaps some gap in time between phases of thrust-fault activity. The plug is probably the eruptive center for the rhyolite flows, and is also placed in member L.

Age The relative chronology is clearly displayed by field relations, but the absolute timing of the various members is not well established in this area. There does not appear to be a pronounced disconformity between member I of the King Lear and member G of the Happy Creek (though member H of the King Lear is missing in this area, having pinched out to the south). The base of member I is therefore younger (but probably not much) than Bathonian. Member I also contains clasts identical to dated intrusive rocks from this area with an age of 173.3 $\pm$ 14.3 m.y. (Rb/Sr, from R. Kistler, pers. comm., 1988), and so again must be Middle Jurassic or younger. Zircons have been separated from the rhyolite flows of member L, but have not yet been dated. The reworked limestone boulders (clearly originating in the Bliss Canyon Formation) in member J also contain upper middle to upper Norian fossils, and so the member is younger (considerably) than this (see chapter 3). Miocene basalt flows (Willden, 1963) overlie the formation with angular unconformity, so the unit is also Miocene or older.

Depositional Environment The lower, massive and matrix-supported red conglomerates and breccias of the lower part of member I are characterized by very large clasts, very poor sorting and red wacke to arenite matrix. This characteristic facies was clearly deposited by debris flows. The clast-support conglomerates and volcanic arenites, with better sorting, cross-stratification, pebble imbrication, normal grading and channel scour and fill are distinctly braided fluvial channels. The micritic limestone was deposited in an areally limited lacustrine environment, perhaps dammed by debris flows. The lenticularly laminated siltstone with root impressions was laid down in a levee or overbank setting during a high-energy flood event, and then overgrown. This collection of facies occurs in a semi-arid alluvial fan environment, probably in a mid-fan setting. The red coloration of the sequence, caused by hematitic staining, supports the conclusion of a semi-arid origin. The upwards disappearance in members I and K of debris flows, decreasing clast size in the conglomerates, better sorting and decreasing overall grain size, thinner bedding, greater development of paleosol silcrete horizons and increasing trough cross-stratification and channel fill and scour all indicate a more distal alluvial fan environment for beds higher in the section. The source area was either more remote or had lower relief.

The white rhyolitic crystal-rich ash-fall Jackson Tuff in this area was laid down during a plinian pyroclastic event in a distal setting (as evidenced by the homogenous texture, lateral extent and thickness distribution and the fine grain size - Cas and Wright, 1987). The vent location is known (possibly, it is related to the silicic Late Jurassic volcanism in the nearby Pueblo Range - Roback, 1988). The tuff is very similar to the silicic volcanics of member L, and may indicate activity in that phase of volcanism began quite early (though the stratigraphic levels are quite disparate). As the thicknesses of such plinian deposits are generally not more than 25 m, this was either an unusually large eruption or the tuff is a composite of several close-spaced events; the former conclusion is favored as there is no internal evidence for a hiatus. The 700 Ka Bishop Tuff, for example, is a similar crystal-rich rhyolite tuff (with the same assemblage of quartz, sanidine, plagioclase, biotite and Fe-Ti oxides) covering an area well over a thousand km^2 in area to a depth of hundreds of meters (Bailey et al, 1976), and is very similar in nature to the Jackson Tuff of the King Lear Formation. A number of similar thinner and less extensive tuff beds occur within members I and J.

The cherty and lithic conglomerates and arenites of member J have a very similar facies interpretation to the volcanic conglomerates and arenites of member I and K. Dominantly braided fluvial channel and overbank sheetflood deposition in a middle to distal alluvial fan is inferred for the unit. Due to the unique setting, however (overlying a thrusted package containing Bliss Canyon Formation rocks well back from the forward edge of the thrust belt) the member may have been deposited as an allochthonous or para-autonochthonous piggyback successor basin on top of and within the thrust belt.

Member L is, as were both the other members, quite distinctive in provenance. The lower volcaniclastic interval is entirely composed of angular white rhyolitic lapilli, with no cherty/lithic or andesitic volcanic detritus such as dominated in members J, K and I, respectively. The interval is characterized by low-angle cross-stratified dune forms and planar beds with scour and fill features. Large grain size variations on a mm to cm scale establish clear stratification, with good sorting in the individual laminae. These deposits are best interpreted as proximal pyroclastic surge deposits - volcanic turbidity currents related to the collapse of a phreatomagmatic eruption column. Whether one such event or more is represented, and the extent of later epiclastic reworking, is not clear. It does not seem possible stratigraphically that this interval is the lateral (and more proximal) equivalent of the Jackson Tuff, but instead a completely separate and later phase.

The overlying rhyolite flow sequence of member L was deposited as very thick, viscous flows close to a vent (such flows are not capable of travelling long distances). This origin is indicated by the intense internal flow-shearing, the related flow-breccias, glassy nature and obsidian bases, and areally restricted nature of occurrence of the member. The rare interbedded rhyolitic breccias are probably debris flows (and perhaps volcanic lahars), while the sand-sized volcaniclastic layers may be more surge-related lapilli tuff beds. Certainly there are ash-fall tuffs as well, laid down by deposition from small sub-plinian eruptive events (judging by their thinness, fine lamination and fine grain size). Some of the glassier rhyolite beds (lacking flow-breccias but with brittly fragmented crystals) may also be welded or ignimbritic tuffs.

The rhyolite plug is an upheaved protrusive dome. Whether the level exposed is from the emergent portion or was buried is not clear. In some cases shallow protrusive domes cannot be distinguished from emergent ones (Williams and MacBirney, 1979). The marginal breccia carapace is characteristic of such silicic domes, and is related to the protrusive growth of the dome and the repeated distension and fracturing of the outer crust. The dome does contain epiclastic deposits, supporting emergence. The family of hypabyssal rhyolitic dikes is associated with the dome spatially and (by field relations and similar lithology) temporally. Though similar in composition to the rhyolites in the dome and flows, these intrusives are slightly coarser- grained and are white, light green or pink instead of shades of red, purple and grey (the color is probably related to the oxidation state of iron and the subaerial vs. intrusive location, and is characteristic of the hypabyssal phases of the rhyolite volcanic centers).

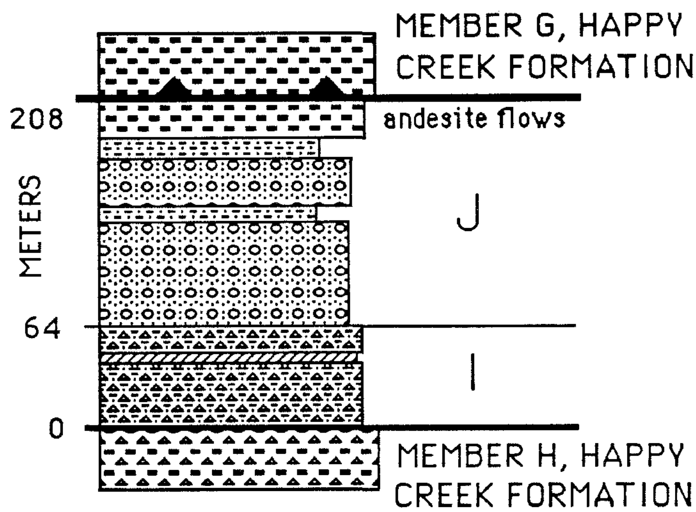
(2) Parrot Peak

This area covers the headwaters of Mary Sloan Creek, with the section measured along the saddle just southeast of Parrot Peak. The underlying member H of the Happy Creek Formation is distinguished from the basal King Lear by the more massive nature and interbedded red andesite flows of the former; the contact is conformable and the sedimentary facies similar. The top of the King Lear section is not preserved as a result of thrusting in the headwaters of Mary Sloan Creek. See the stratigraphic section in fig. 35 for details.

Stratigraphy The uppermost Happy Creek Formation in this area consists of thick and homogenous red andesite flows and volcanic epiclastic breccia beds. The lowermost unit of the King Lear - member I - is made up of 64 m of similar, thick but well-bedded, dark red sedimentary breccias. The clasts are of both volcanic Happy Creek and Early Mesozoic Intrusive Suite plutonic provenance. They are highly angular, poorly sorted and are in clast support with a red arenitic to muddy matrix. Locally, clasts range up to a meter or so. The interval lies above the last andesite flow of the Happy Creek. In the middle of the interval is a pronounced silcrete horizon, with black and grey banding and mottling and extensive silicification overprinting the breccia textures. The

FIG. 35
2 PARROT PEAK

STRATIGRAPHY OF THE KING LEAR FORMATION



portion of the member above this horizon contains dark red thin-bedded volcanic arenite and siltstone layers as well as the breccias, unlike the lower portion. The breccias can be reversely graded at the bases, and normally graded and cross-laminated at the tops.

Overlying the red volcaniclastic lithologies of member I are green to grey chert-pebble conglomerate and chertarenite beds of member J. The member is a minimum of 142 m thick, and is structurally overlain by a thrust-slice of Happy Creek. The contact here is quite sharp. Clasts in the conglomerate are well-rounded and up to 40 cm in diameter. Volcanic debris in the member is present but subordinate to rare. The debris consists of grey to green chert and greenish quartzite, very similar to the siliceous clasts seen in the McGill Canyon and Bliss Canyon Formations. The clasts in the conglomerate are in clast-support in a well-sorted arenitic matrix, and are commonly imbricated and moderately sorted in thick beds. Some normal grading, low-angle cross-sets and lenses of cross-stratified arenite are also present. Interbeds several meters thick of red homogenous siltstone and volcanic wacke occur at several points in the section. Higher in the section some of the chert-pebble conglomerates can also be red in color.

At the top of the interval there are 8 meters of green, tan and grey siltstone and fine-grained arenite, with abundant plant-fossils. The strata are very thin-bedded and well-sorted with planar laminations. The top of this horizon was convolutely and ductilely deformed before lithification of the sediment. Directly overlying this convolute zone (and presumably causing it) are two basaltic andesite lava flow units, 10 and 18 m thick. The flows are rather glassy (particularly at the base) and dark green in color, with spheroidal devitrification textures. In thin section the texture is fine-grained intergranular, with augite granules, andesine laths, plates of chlorite (inferred to be after hornblende), and interstitial chlorite and Fe-Ti oxides. Rare small intergrain vesicles are also filled by chlorite. The flows have undergone extensive replacement by chlorite, and the plagioclase is extremely turbid and altered. The flows thin and disappear rapidly to the south in the section, and are truncated structurally to the north; this is the only outcrop of this facies in the entire range. Because of their lateral restriction and thinness they will be retained in member J.

To the south within this area, member I thins (to 10 meters or less on a syndepositional horst) and consists of clast-support imbricated conglomerates instead of breccias. There is also, at the contact of the volcanic- and lithic-dominated strata, a zone of mixed chert and volcanic pebbles in the conglomerates which is placed within member J. At the contact with the Happy Creek, that unit is weathered and oxidized, with grey seams of calcitic caliche.

Age No age constraints exist for this section, other than that it is younger than the underlying member H of the Happy Creek (which is constrained to the Middle Jurassic). As the contact between members H (Happy Creek) and I (King Lear) is gradational, member I probably dates from a period just younger, also in the Middle to Upper Jurassic. The attempt to extract zircon from the basaltic andesite flows was not made, but the age information from these flows would be extremely useful in constraining the age of the lower King Lear Formation. The plant fossils in the siltstone, if sampled, might prove similarly useful.

Depositional Environment The sedimentary red volcanic breccias of member I are certainly proximal subaerial debris flows, being shed off some structural relief exposing Happy Creek (and no stratigraphically deeper) lithologies. Due to the size, the sorting and angularity of the blocks this sourceland was very close by. The facies can be traced southwards, thickening and coarsening up to the high-angle faults bounding the margin of the basin containing the section (on top of the horst in area 3, the member is much thinner, better sorted and fluvially worked). The basal inverse grading is characteristic of such debris flows, and the cross-lamination at the top of some of the beds was caused by late-stage sheet-flood reworking at the top of the mudflow during the emplacement event. The cause of the relief will be examined in later sections. The silcrete horizon is an ancient paleosol; such siliceous soil horizons form in semi-arid environments over a period of thousands of years (Collinson, 1986). A semi-arid to arid environment is also indicated by the red hematitic staining evident in the breccias, and the color of the finer-grained clastics from the top of the member. These latter siltstones and arenites are more sheet-flood deposits. Member I was laid down in fault-scarp-related alluvial fans by mass flow processes; the thinner southern exposures within the basin may have occupied an inter-fan position.

Member J is quite distinct in color and other aspects. Volcanic debris is subordinate at the base and absent at the top. The clasts are better rounded and are better sorted. The new sourceland was probably considerably more distant, though some of this apparent maturity could be inherited. The internal structures indicate sedimentation from traction (normal grading, imbrication and cross-stratification). This facies has a distinctly fluvial channel origin. The red siltstones and wackes are interchannel overbank to floodplain deposits, exposed for longer periods of time to subaerial oxidation and weathering. The copious plant fossils support a subaerial origin as well. The planar laminated siltstone and sandstone at the top of the sedimentary portion of member J are also flood-plain facies, perhaps with individual flood events recorded. The abrupt dominance of chert and quartzitic clasts in the section implies that in the sourceland, the McGill Canyon and Bliss Canyon Formations were tectonically exposed for erosion (possibly by the west-vergent thrust system lying to the east). Member J was deposited in an alluvial flood-plain by high-sinuosity fluvial and overbank processes. The thin interbeds of volcanic wacke represent interfingers (though too thin to break out) of member K.

The basaltic andesite flows overrode these sediments while they were still unconsolidated, judging by the extent of ductile deformation in the meter or so just below the glassy base of the flow. The flows themselves are subaerial by association, and appear to have been rather fluid - they can be traced for several hundred meters along strike before thinning and disappearing, are glassy, and lack associated flow-breccia facies. These flows are, for the King Lear Formation, highly mafic and unusual, and may represent the last gasp of the more mafic Happy Creek phase of volcanism.

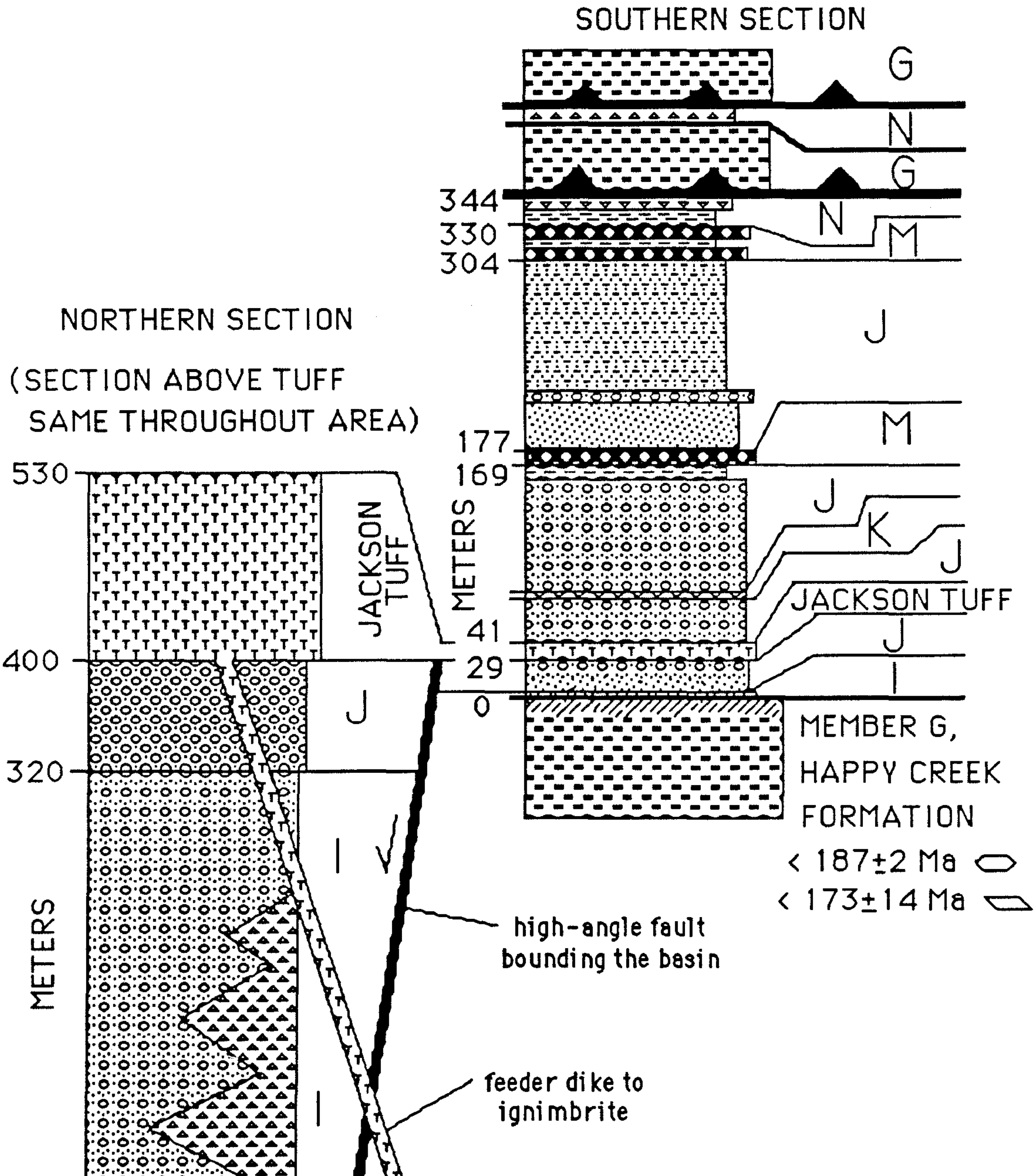
(3) Mary Sloan Creek

The area covered here is the triple divide between Mary Sloan, Happy and Jackson Creeks. The measured section extends east-west from the headwaters of Mary Sloan Creek to the Jackson-Happy Creek divide. The section has been cut and carried by a west-vergent thrust (so the base is not seen) and has been overthrust on the eastern end (and stratigraphic top) by another such sheet. The area of the section straddles a high-angle fault system that was active during deposition of the lower part of the formation, and stratigraphic columns from the horst and graben areas are both presented. See the two stratigraphic sections in fig. 36 (northern graben and southern horst, respectively).

<u>Stratigraphy</u> Within the down-dropped fault-bounded basin in the northern half of the area, the lowest strata of the King Lear above the thrust fault belong to member I. The basal lithologies of member I (as well as those next to the high-angle fault higher in the section) are massive red breccias. They exhibit matrix to framework support, and contain very angular and poorly sorted clasts ranging up to meters in size (including some mega-blocks tens of meters across) with a source in the Happy Creek. The unit thins away from the graben margin. Where at its thickest and coarsest against the margin it is also slightly sheared by the fault system. It overlaps the edge of the horst, as well. Just to the east, the shoulder of the horst structure is incised by a large canyon feature filled with highly immature red sedimentary breccias (and lacking interbedded andesite flows), which is also interpreted as belonging to member I of the King Lear. The breccias pass laterally

FIG.36
3 MARY SLOAN CREEK

STRATIGRAPHY OF THE KING LEAR FORMATION



MEMBER I, KING LEAR FORMATION (AREA 2)

(northwards) and upwards into well-bedded red volcanic conglomerate, breccia and arenite. This interval is characterized by clast-support and pebble imbrication, low-angle cross-bedding and small channel features. Member I is 320 m thick within the basin. On the horst to the south it is missing to only several meters thick, and is characterized by a thin red clast-support conglomerate with an arenitic matrix.

At its base, member J has intermixed cherty/lithic and volcanic debris, but the volcanic component disappears rapidly upwards. J is characterized by green chert-pebble conglomerate, chertarenite, and shale. On the shoulder of the horst, only several m of this member exist underneath the Jackson Tuff bed, and there are volcanic clasts within the conglomerates. Across the fault system bounding the horst and graben, some very interesting relationships exist. Several growth faults with north side down increase the thickness of member J underneath the isochronous tuff in several discrete steps to 80 m. The tuff overlaps some of these high-angle faults or is very slightly offset by others, thus establishing the major period of activity on the faults as being pre-eruption.

The Jackson Tuff here is thick - between 20 and 130 m. Rhyolitic dikes very similar to the tuff cut the Happy Creek, the high angle faults, member I and lower Member J of the King Lear. These very light green or gray dikes come up to the tuff and join it without cutting the tuff or strata above it. There are four of them, very planar and parallel, and several m thick and tabular with fluidal banding and a trend of N20W. Where the dikes are closest to the outcrop of the tuff the tuff bed is thickest. The tuff consists of white to very light green hard and siliceous rhyolitic crystal-rich ashy (vitric) tuff, very fine-grained and grading at the base into a foliated light green glassy welded ignimbrite. The ignimbrite has flattened and plastically stretched and deformed pumice fiamme plus volcanic rock fragments, feldspar and quartz grains in a glassy matrix (a eutaxitic texture). The base of the ignimbrite can be highly glassy and verge on obsidian in texture. Rheomorphic deformation of the ignimbrite has resulted in both a flattening foliation and a shear-induced lineation due to secondary mass flowage.

In thin section the lower, ignimbritic portion of the tuff bed is composed of welded, compacted and flattened angular glass shards with a few percent by volume of crystals of quartz, sanidine and a mafic phase (altered to chlorite). The dikes in thin section are very similar to the tuff and consist of locally devitrified and oxidized glass with microlites of hornblende, muscovite, topaz, quartz, sanidine and Fe-Ti oxide. The quartz, as in the tuff, is made up of large euhedral but corroded and embayed ã-quartz grains. Siliceous fluidal banding is present in some of the dike samples, and significant amounts of volcanic xenoliths can be entrained (these latter appear to have originated in the Happy Creek Formation and to have undergone substantial alteration).

The stratigraphy above the tuff in this area is not affected by the high-angle faults, the stratigraphy in the two sections is the same. Overlying the tuff bed is more member J, with the addition of several thin intertongues of red volcanic conglomerates assigned to member K (member K interdigitates from the north in this area). These latter beds are cross-bedded and have volcanic arenite lenses and tops (normal grading is present), with subrounded pebbles. The green chert-pebble conglomerates of member J have similar sedimentary structures and well-rounded and imbricated clasts, but are interbedded with green chertarenite, siltstone and shale plus a thin lens of grey micritic limestone. The arenites are well-sorted and lack fines, are subrounded to rounded, cross-laminated and have abundant silicified logs and plant imprints.

One hundred and twenty-eight m above the tuff, a 7 m thick bed of diamictite is found interbedded with member J. The diamictite has a dark green to grey feldspathic siliceous and very fine-grained wacke matrix with angular clasts of coarse-grained green plagioclase-phyric dacite and less common, more rounded chert, quartz and diorite clasts. The clasts are very poorly sorted, ranging from sand-sized to cobbles, and are matrix-supported. The clasts total only 10-30% of the rock by volume. Log imprints are found at the top of the bed. This lahar bed and several others like it in the overlying interval are assigned to member M, and interfinger from the south.

Overlying the diamicite of member M is 128 more meters of green cherty siltstone, arenite and conglomerate of member J. Overall the member is much finer-grained and homogenous, and conglomerates are rare. At the top are two more diamictites of member M, 4 and 9 m thick, with 12 m of green siltstone in between them.

Overlying this upper occurrence of member M unconformably is a coarsening-upwards sequence, 14 m thick, of member N. Red siltstone passes upwards to red volcanic arenite and then to a red volcanic framework breccia with an arenitic matrix or open voids. Immediately over this is emplaced a thrust plate of Happy Creek lithologies, identical to the clasts in the breccia (with various flavors of andesite). This thrust plate is thin and is unconformably topped by 7 m more of the red volcanic breccia, and then by another thrust sheet of Happy Creek rocks.

Age: No geochronologic data was determined in this area; dinosaur bones and plant leaf and stem imprints were collected, but were not diagnostic.

Depositional Environment It is quite clear that the east-west trending high-angle faults were active with a vertical offset of at least 400 m before and during deposition of members I and J of the King Lear (starting during deposition of member H of the Happy Creek). The vertical offset is found from the stratigraphic distance between the strata overlapping the fault and the lowermost strata of the King Lear. The magnitude of horizontal displacement is unknown but could be significant. Offset could have been substantially more, as the base of the King Lear is not present here and the shoulder of the horst might have been eroded. These growth faults were active during sedimentation, but the first phase of their activity had almost entirely ceased by the time that the Jackson Tuff was deposited.

The basal beds of member I are of proximal debris flow (with matrix-support) to fault-scarp talus (with open framework) origin, as evidenced by the lack of sorting and bedding, the texture and the presence of mega-boulders. This facies is intimately associated with and temporally overlaps the high-angle faults, which must have been topographically evident as active scarps. The facies was deposited in the upper alluvial fan or slope apron setting against the fault scarps, and was shed from them. The scarps were at least partially incised and degraded, as a canyon feature cut into the shoulder of the horst is evident. This feature is also filled by very proximal epiclastic debris. The fan may indeed have been fed through this north-south paleocanyon, incised into the basin margin.

Both upwards and away from the fault scarps the lithologies become more mature and distal (good bedding, finer grain size lacking mega-blocks, better sorting, and the sedimentary textures of clast-support, imbrication, and cross-stratification and channel cut-and-fill). This transition may also exist along strike on the fault scarp, as the more proximal and coarser facies are missing in the lower thrust sheet in area 2 just to the west. Member I is, however, almost entirely missing just to the south on the actual shoulder of the horst. In this area it is characterized by thin and patchy occurrences of coarse-grained red andesitic epiclastic conglomerate of a fluvial origin (with clast-support, an arenitic matrix and significant rounding of the clasts).

In these same areas of relative uplift, member J below the Jackson Tuff is also quite thin and consists of green lithic and cherty conglomerates with textures similar to the underlying red beds of member I. The interpretation for these beds is the same - fluvial (and probably braided, though diagnostic features are lacking). Within the graben, the lithologies and facies of deposition are the same, but the interval below the Jackson Tuff is much thicker (due to the growth faults). Above that bed, member J shows no evidence of synsedimentary faulting. Member J in this interval is composed of distinctly fluvial conglomerates and lithic arenites (with some overbank siltstone and shale and lacustrine limestone facies) and was subaerial (judging by the abundant terrestrial plant fossils).

The overall setting was a braided fluvial flood plain, which became more distal (or relief in the sourceland decreased) upwards in the section. The interfingers of volcanic epiclastic red beds of member K from the north (member J disappears towards the northwest within area 1) are of similar origin, and the interdigitation must mark the confluence of two drainages with distinct provenance but similar fluvial features.

The Jackson Tuff itself is clearly the result of phreatomagmatic explosive events. A welded ignimbrite formed as the plinian column collapsed, and was covered in the later stages of the event by fine vitric and crystal ash-fall tuff. The ignimbrite was hot enough to weld to a fiamme obsidian at the base, and thick enough that ductile flow after deposition and compaction is recorded by the internal textures. A proximal origin is indicated by the thickness and presence of flattened and deformed pumice fragments. The tuff may be at least in part the result of a fissure eruption, as four planar and parallel dikes of similar rhyolitic composition up to several meters thick are observed to come up to the base of and feed the tuff in this area, where it is unusually thick (on the order of 100 meters). Such fissure vents are often associated with unusually thick ignimbrites (Cas and Wright, 1987). The feeder dikes crosscut the high-angle faults and lower members I and J below the tuff, but are not seen to occur stratigraphically above it. This relationship and the petrography support the identification of the dikes as feeders to the ignimbrite.

Member M is present in this area as several diamictite beds high in the section. These beds are clearly volcanic debris flows - lahars - as evidenced by the matrix support, nature of the clasts and matrix, and texture. The source was distinctly the plagioclase-phyric dacitic volcanic province. This dacite volcanic center is well exposed further south, and these beds represent its northernmost extent. Member M interfingers with member J and K in the upper part of the stratigraphic section as several thin, distinct beds. Member J in this interval is finer grained and more distal than in lower occurrences, dominated by overbank and floodplain facies shale and siltstone at the top; the fluvial conglomerates and later the lithic arenites become less abundant upwards.

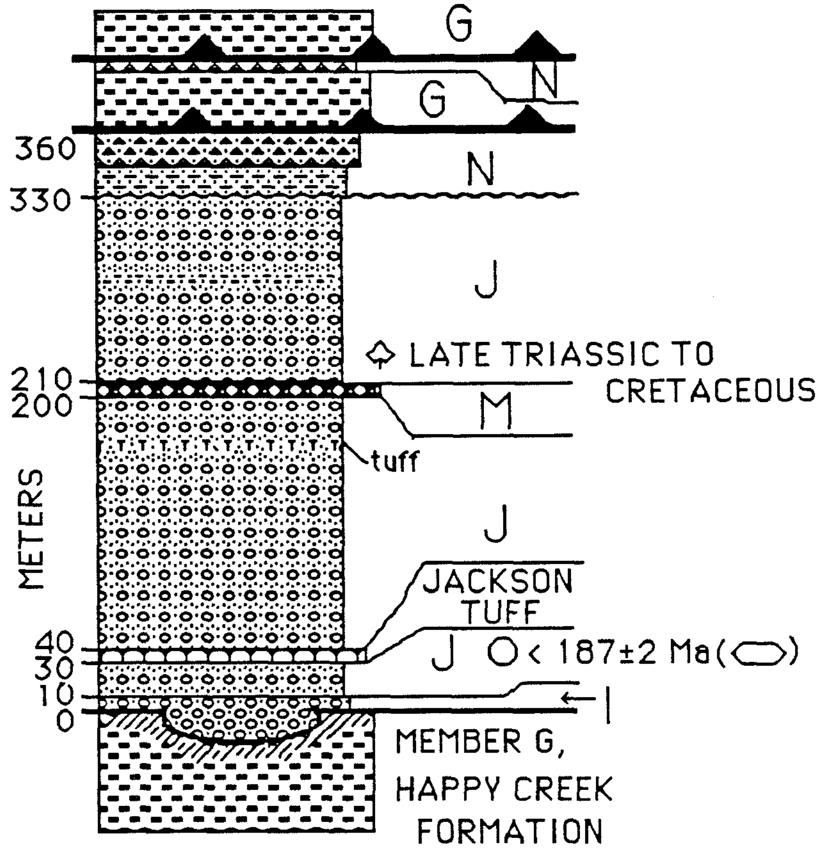
At the very top of the stratigraphic section (and overlying member M) is member N, a coarsening-upwards sequence of volcanic epiclastic red beds passing from siltstone to arenite to an open framework breccia. The latter facies especially is characteristic of extremely proximal deposits in an area of significant relief. This unit represents debris shed from an advancing thrust sheet (composed of Happy Creek rocks) and then, in turn, overridden and folded (into a tight overturned syncline) by that thrust sheet. The nappe is itself quite thin (the eroded toe) and is itself overlain by more such fault scarp talus breccia shed from another immediately overlying thrust. Thrusting is thus definitely established to have been synchronous with sedimentation in member N, and possibly to have been a factor in the sedimentation patterns in member J and upper member I (providing the sourceland for the post-graben volcanic, cherty and lithic debris). The paleogeography will be discussed in detail later.

(4) North Jackson Creek

This area lies to the northwest of the Iron King Mine, within the upper drainage basin of Jackson Creek. It is bounded on the east by thrust faults, on the south by a major high-angle fault, and on the west by the basal unconformity; to the north lies area 3. The section is situated entirely on one of the horsts. This section was not measured in the field, so stratigraphic thicknesses are approximate. See fig. 37.

Stratigraphy At the basal unconformity, member G of the Happy Creek is highly weathered and oxidized to a depth of at least several meters. The lowermost King Lear, member I, occurs as a thin gravelly veneer on the unconformity and as an incised channel. The latter feature is as much as 20 m deep, and is filled by boulder conglomerate. The boulders are up to 1.5 m in diameter, and are all a pink medium-grained granitic intrusive with epidote seams. These boulders are rounded and in clast support with a matrix of red coarse-grained, poorly sorted arkosic arenite (possibly a sieve deposit, filtered down into the interstices after the boulders were emplaced, in such a strongly bimodal deposit). The granite in the boulders has large pink zoned alkali feldspar with myrmekitic textures, plagioclase, mafic phases and quartz. Middle Jurassic plutons of the Early Mesozoic Intrusive Suite very similar to this intrude the Happy Creek and occur as xenoliths in members F and G; these intrusive bodies must have been exposed nearby by the time of iniation of King Lear sedimentation. Where the boulder conglomerate channel deposit is not

FIG.37
4 NORTH JACKSON CREEK
STRATIGRAPHY OF THE KING LEAR FORMATION



present, only about 10 m or less of red volcanic conglomerate and volcanic arenite is present.

Above member I with a fairly abrupt contact is green cherty and lithic conglomerate, green siltstone and green and grey shale of member J. At 30 m there is a 2 to 10 m thick tuff, the Jackson Tuff, which thins to the south. Though much thinner here, it is identical in lithology to other occurrences, consisting of light green welded ignimbrite with xenolithic lapilli at the base grading to a white very fine-grained vitric and crystal tuff at the top. In thin section phenocrysts include euhedral and corroded quartz grains and sanidine in a semi-devitrified glassy matrix with siliceous convolute bands and deformed fiamme. Member J continues above the tuff with interstratified thick beds of chert-pebble conglomerate, green siltstone and shale, fining upwards. The conglomerates are imbricated and in clast support with a lithic chertarenite matrix and have sandy lenses and a high degree of grain sorting and rounding. The clasts are green, grey, and black chert, quartzite and white vein quartz. These clasts, as elsewhere in member J, are identical to those in the McGill Canyon (and to a lesser extent) the Bliss Canyon Formations.

There is a very thin grey medium-grained crystal tuff (full of brittly broken plagioclase grains) in the upper part of the interval (and also characteristically full of plant fossils), and several distinct thin beds of arkosic coarse-grained sandstone as well. This tuff bed is found at the same stratigraphic position in areas 3 and 9 as well. Overlying these occurrences at about 200 m and probably part of the same volcanic episode is a thick bed of diamictite. This bed has rounded clasts of chert, argillite (apparently intraformational) and coarse-grained plagioclase phyric dacite up to 20 cm in size floating in a dark grey-green siliceous and very fine-grained matrix. The bed is part of member M and correlates to the lower such occurrence to the north in area 3 (the upper finger has been cut out by erosion beneath member N). In thin section the diamictite is very poorly sorted, and the matrix is composed of volcanic rock fragments (plagioclase phyric dacite), plagioclase and minor quartz grains floating in mud with interstitial voids filled by chlorite.

Member J continues above this pinching-out finger of M with rocks similar to those below it. Fingers of member K (present just to the north) are, however, missing. At the very top of the vertical extent of the member, there is an admixture of volcanic clasts in the conglomerate, and member N overlies member J with significant angular unconformity at about 330 meters. J had been folded and warped before N was laid down.

Member N at the base consists of 10 m of red siltstone and volcanic arenite, grading upwards to 20 m of volcanic sedimentary breccia. The breccia is very coarse and is well-bedded, with weathered and oxidized clasts of Happy Creek lithologies (including epidotized rocks). The breccias have very poor sorting, very angular clasts, clast-support and either a volcanic arenite infilling or grey sparry limestone cement. Sedimentary features (not always present) include clast imbrication, cross-bedding, scoured channels and cross-laminated sand lenses. The member is highly laterally variable. An additional type of deposit was thick and muddy beds with large breccia clasts floating in the center of the bed (reverse grading at the base, normal grading at the top). The upper, breccia portion of the member also includes cross-laminated magnetite placer deposits. These placers are made up of very angular but well sorted and graded magnetite grains in a very coarse-grained, monocrystalline calcite cement (no matrix). Red shale beds are also present, as well as a bed of grey micritic limestone. The limestone is burrowed, with sandy patches, calcispheres and some unidentified fossils. Juxtaposed structurally above this member directly is a thin thrust plate of diorite and Happy Creek andesite (identical to the clast population in member N). This thrust slice is in turn overlain by several meters more of breccia of member N, with another thrust sheet of Happy Creek above that.

<u>Age</u> Zircons have been separated from the monzonitic boulders at the base of member I. When dated they will give a maximum age for the unconformity and for that member. The boulders are very similar to the Harrison Grove pluton, which has a preliminary U/Pb zircon age of 187 $\pm$ 2 m.y. A leafy branch was collected from J, and identified by S. Ash (written communication, 1986) as belonging to the conifer <u>Pagiophyllum</u>, ranging from the Triassic into the Lower Cretaceous. The formation thus lies between the Middle Jurassic and the Early Cretaceous in age. Additional dinosaur bone fragments were found in member J, but are non-diagnostic (W. Clemens, written communication, 1986). No minimum age constraints have come from this area, though the thrust fault far to the south has been cut by a monzonitic pluton dated at 162 $\pm$ 1 m.y. by U/Pb geochronology on zircons.

Depositional Environment Upon the horst, member G of the Happy Creek is deeply weathered, and a hiatus may have taken place before the deposition of member I of the King Lear. The boulder conglomerate-filled channel at the base of member I is definitely fluvial, incised and then filled by a stream system. The sandy matrix was deposited, after whatever catastrophic flood event emplaced the boulders, by sieve sedimentation. The thin veneer of similar facies seen in other sections also seems to be fluvial, and could have been deposited by braided stream systems across a pediment floored by member G. This member could cover a substantial amount of time if net deposition had been near zero, with debris passing through the system to be laid down elsewhere. By the size of the boulders, the monzonitic source must have been fairly close.

The appearance of member J is abrupt, marking a sudden change in provenance and local paleogeography. The detritus is entirely lithic (chert, quartzite and vein quartz) and lacks a volcanic component. The rocks, imbricated clast-support chert-pebble conglomerates with lithic chertarenite lenses, green siltstone and shale, were deposited on an alluvial flood-plain or the distal part of an alluvial fan in stream and overbank settings. Several large fining-upwards cycles may exist, but are not well defined. Much of the apparent maturity of the sediment is inherited, as the clasts come from the McGill Canyon Formation and were already sorted and rounded. No fingers of member K are present in this area; the paleogeographical inference is that the sourceland to and drainage system containing that member lie too far north and east.

The Jackson Tuff is substantially thinner here. This may be because it was more distal, or because the thicker localities are in the pre-existing basins formed by the high-angle faulting and the ignimbrite ponded in these topographic lows. Nevertheless, the tuff is a welded ignimbrite grading upwards to an ash-fall tuff and represents a single event.

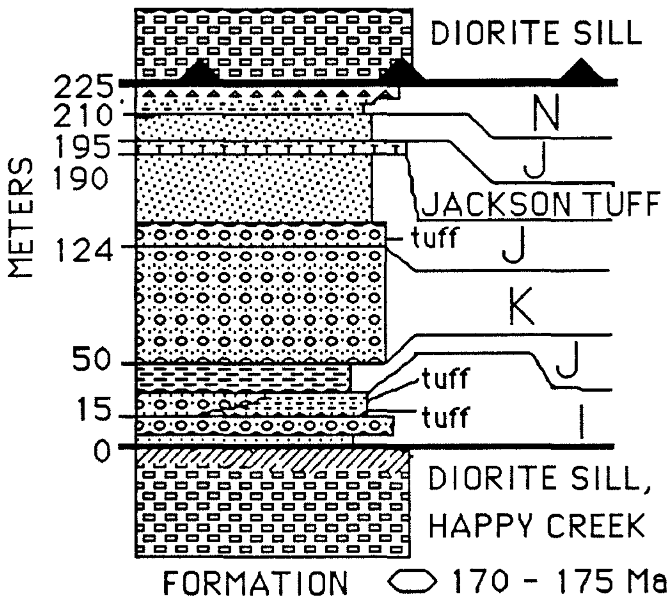
The diamictite of member M is a volcanic lahar or mud flow that entrained some surficial sedimentary debris. The volcanic province it originated from appears to have been dominated by plagioclase phyric dacite, distinct from the andesitic flows in the Happy Creek. The implication is that such a dacitic volcanic system was active during deposition of the King Lear. Other evidence of it is seen in member J (though not extensive enough to break out) in the crystal tuff and certain arkosic arenites, which are inferred to be reworked tuffs. The upper lahars seen in area 3 are missing, and were probably eroded before the deposition of member N.

The existence of such an intraformational discontinuity is also supported by the angular unconformity between N and J of as much as 30°. The King Lear section was deformed by the approaching thrusts before the last member was deposited and, in turn, overridden. Member N is clearly intimately related to the nearby thrusts: the abrupt change in provenance from J and M at the base, bedding dipping away from the fault, the very pronounced coarsening upwards sequence (as the thrust frontal scarp advance closer), and the nature of the debris all lead to this conclusion as well. The detritus comes uniquely from the overlying thrust plates. The magnetite, in particular, is a good fingerprint as magnetite mineralization is extremely well-developed in the immediately superposed nappes. Indeed, the presence of this mineral and of epidote in the debris indicates that the alteration and mineralization in the Happy Creek took place before deposition of member N. N was laid down by fluvial and debris flow processes in an upper alluvial fan and talus apron environment in front of the fault scarps. Small depressions filled with ponds or lakes must also have existed (the limestone and shale). Fluvial processes are indicated by the bedded, clast-support and cross-bedded nature of most of the breccias and volcanic arenites, while a debris flow deposited the thick diamictite bed with floating clasts concentrated at its center. The second sliver of member N, unconformably on the overlying thrust plate, may indicate out of sequence thrusting (the breccia apparently overlapped the toe of the lower thrust, and was then imbricated). The alternative is that this upper breccia belongs to member I, and that thrusting was in sequence. This seems unlikely as in several occurrences of this type only the volcaniclastic breccia is present and never the overlying chert-pebble conglomerates of member J.

(5) Iron King Mine

The stratigraphic section to be discussed here was measured with a Jacobs staff on the slopes west of the Iron King and DeLong mines, and is contained within a single thrust plate in the eastern, west vergent orogenic zone. It is bounded on the north by a major high-angle fault (reactivated as a tear fault during thrusting), and is cut out to the south by the upper thrust fault. Some of the best exposures of the base of the King Lear Formation are included in this section, along the road cuts up to the Iron King Mine. See fig. 38.

FIG. 38
5 IRON KING MINE
STRATIGRAPHY OF THE KING LEAR FORMATION



<u>Stratigraphy</u> The lithology underlying the unconformity is actually a fine-grained diorite of the Early Mesozoic Intrusive Suite in this area, implying some unroofing and erosion before deposition of the basal King Lear. The diorite is grey-green except within 10 m of the nonconformity, where extensive weathering (particularly of the plagioclase, which is turbid) and reddening due to oxidation takes place. Grey seams of dolomitic and calcitic micrite penetrate down to 10 m from the contact. At the contact at the base of the King Lear there is an irregular layer 50 cm thick of similar micrite, mottled in with greenish chert; in places these lithologies can be seen to be overprinting the basal clastic lithologies. Red on grey lisagon staining is also abundant near the contact within the diorite.

The basal King Lear is composed of green (at the base) and red volcanic cobble conglomerate, coarse-grained volcanic arenite, siltstone and shale. The latter two lithologies have septarian nodules developed. The conglomerates have rounded to subangular clasts, up to 50 cm in size, imbricated and in clast-support and composed of monzonitic, andesitic and dioritic rocks from the Happy Creek. Some of the intrusive clasts were already epidotized. The clasts have rinds, are oxidized and contain hematite. Normal grading at the top and large-scale trough cross-stratification (often in sandy lenses) are abundant. Sorting of the cobbles is moderate to poor, and the matrix is an immature arenite to wacke. Where the matrix is missing there is a green siliceous cement. The bases of the beds are scoured and uneven. The conglomerates form lenses up to 12 m thick and several kilometers in width in a sequence of predominantly thinly-bedded shale, siltstone and arenite; two such lenses exist in this area. The lowermost 6 m of the formation are made up of shale and siltstone. The lenses of conglomerate are thicker to the north, but the member as a whole thickens somewhat to the south. These facies, member I, are a total of 20 m thick along the section line. In thin section, a pebbly volcanic arenite exhibited volcanic rock fragments, sericitized plagioclase, monocrystalline fresh quartz, and detrital Fe-oxide with a chlorite and chert cement. The texture was sorted and subangular, and interstitial fines were lacking. The source, by the similarity in clast lithologies, was the Happy Creek.

Overlying these volcanic epiclastic sediments abruptly is member J- a cherty and lithic clastic succession. The bottom bed is a cobble conglomerate, resting with a scoured base on I (and locally cutting out the upper part of I). The sedimentary structures are identical to those in the channel deposits in I. The provenance is distinct (the clasts are 99% chert, quartzite and vein quartz). This bed also is laterally extensive and thickens to the north from 5 to 15 m. In detail it is a composite of many smaller cobble conglomerate lenses contained in green shale, siltstone and lithic arenite. Similar fine clastic facies overlie the chert-cobble conglomerate lens up to 48 m, where a second abrupt change in provenance takes place.

The provenance changes back to a volcanic source in member K. Red and green volcanic arenite and pebble conglomerates with paleosols overlie J. At 76 m, a very large channel feature (extending up to 124 m in the section) is present, and thins fairly rapidly to the south. This feature is filled by rather uniform .5 to 2 m thick beds of tan to green pebbly sandstone to siltstone, normally graded. The beds are generally conformable but locally can have scoured bases, and are cross-laminated with ripple bedding. From a distance the bedding within this channel feature can be seen to be at a very low angle to overall bedding.

Above the channel feature there is a change in provenance back to cherty and lithic (member J). The basal portion is a grey organic-bearing reworked tuffaceous sandstone with abundant black carbonized plant imprints and silicified logs up to 30 cm across and a meter long. The tuffaceous sandstone is composed of quartz and plagioclase grains and less abundant fragments of both volcanic and metamorphic origin in a matrix of long, thin angular glass shards with carbonized plant remains. This volcaniclastic is a distinctive horizon found in other sections (3, 4, 9 and 10). The logs are coniferous, and have rings indicating a temperate region with strong seasons (S. Ash, written communication, 1985).

Above that reworked tuff, lithic and cherty arenites continue up to 210 m. They are generally green in color with interbedded green siltstone and grey shale. Several more thin beds of grey tuffaceous sandstone occur as well. Sedimentary textures are rare and the lithologies are homogenous. At 191 m, a 1 to 5 m thick bed of the Jackson Tuff is present, composed of a white rhyolitic very fine-grained ash. Above this the cherty arenites begin to display a detrital plagioclase component as well. This is possibly an admixture related to the dacitic volcanic component of member M, or to encroaching thrusts bearing andesitic basement of the Happy Creek, as in member N.

An abrupt transition to green and red siltstone takes place above these strata, overlain by red volcanic conglomerate passing upwards to red sedimentary breccia. This is member N, which is topped at 225 m by a thrust plate of diorite and Happy Creek greenstone. The conglomerate has clast-support and good rounding. The breccia is composed of angular pebbles compatible with the overlying Happy Creek, and contains lenses of red siltstone as well.

The section is intruded by grey to brown amygdular flow-banded andesitic dikes, up to 2 m thick and thinning upwards. The dikes trend N05E and are affected by the high-angle faults. They have chilled margins, the amygdules are filled by calcite, and amphibole is a phenocryst phase. In thin section, hornblende, plagioclase, and Fe-Ti oxides occur in an intergranular groundmass of plagioclase, clinopyroxene granules, and chlorite with vesicles filled by tridymite. These andesitic dikes are inferred to be syn-King Lear as they are cut by the high-angle faults (themselves cut and carried by the thrusts). They are inferred to be part of the same phase as the mafic flows in member J in area 2, near Parrot Peak.

Some interesting lateral stratigraphic variation is found within this area. A few hundred meters to the south of the measured section, the large conglomerate lens (variable in thickness, 9 - 12 m here) rests directly on the unconformity and is overlain by red siltstone and volcanic arenite, generally much finer grained than the facies just to the north. The chert-cobble conglomerate (only 4 m thick here) at the base of member J is not seen until 41 m, where it rests on the red siltstone, shale and volcanic arenite with a scoured base and flame structures. Overlying the chert-cobble conglomerate is a brownish grey arenite with thin mottled red shaly horizons and root casts. Thin (a few

10's of cm) crystal tuffs are found at the 21 and 34 meter levels.

Further south across a large high-angle fault, the basal unconformity is exposed with a reddish weathered zone 8 m thick developed on the Happy Creek Formation. The unconformity surface is irregular, with the paleosol preserved - angular clasts coming directly from the substrate and floating in a red clayey matrix, 2 m thick. Above this is 12 m of volcanic sub-breccia, red and poorly sorted with a grey sandy matrix. At 14 m there is a 2 m thick bed of very light green rhyolitic tuff (not the Jackson Tuff, which is present stratigraphically higher). This is overlain by at least 12 m of reddish-brown siltstone, shale and volcanic arenite beds 70 to 100 cm thick, and with paleosols developed on almost every bed. All of this, including the paleosol, is member I.

At the southernmost part of this exposure area (north of the headwaters of Trout Creek), the lowermost member I is 2 m of red siltstone overlain by a thick bed of red volcanic conglomerate - the same lenticular body seen throughout the area. This conglomerate has wedge cross-sets 15 to 20 cm thick and pebbly foresets. There are also several diamictite beds, 1.5 m thick, with matrix-support and exhibiting concentration of the largest clasts in the center. This interval of member I is overlain by a few tens of meters of green siltstones (inferred to be member J - much thinner and finer than to the north), and then by the Jackson Tuff. The latter unit is 5 m thick here and is a very fine-grained rhyolite tuff with quartz phenocrysts. The sedimentary fault breccia of member N is not seen here, and Happy Creek volcanic facies are thrust over the section just above the tuff. The Happy Creek at the unconformity is still a weathered and oxidized diorite.

The high-angle fault, running east-west with north-side down, appears to have been a factor during sedimentation. Member I and the lower part of J are deformed and tilted by it, but are similar in thickness and lithology on either side. The beds at the top of the section (the Jackson Tuff and member N) are also unaffected; the fault does not cut either of the two thrust faults bounding the section. The middle part of the section (member J), however, thins greatly to the south by a factor of at least 3 across the high-angle fault zone. The fault perhaps had a surface expression as a monocline; member I is warped by 90° but not actually cut, implying a growth-fault relationship during deposition of that portion. The activity on the fault is thus constrained to be post-member I and pre-Jackson Tuff, similar to most other faults in this generation.

Age The diorite sills in the Happy Creek, upon which the King Lear rests nonconformably in this area, have a preliminary U/Pb zircon of 170 - 175 m.y. The diorite is also present as clasts in I. Member I must therefore be younger than this.

Depositional Environment The basal unconformity shows evidence for both a substantial hiatus and some erosion, judging by the exposure of a hypabyssall sill and by the depth and degree of weathering. The dolomitic seams and carbonate-chert replacement zones below and at the unconformity (calcrete and silcrete) are characteristic of a semi-arid caliche soil zone (Collinson, 1986); the abundance of hematite formation and staining (the lisagon features) supports this.

The facies of member I in the measured section represent alluvial flood plain environments. A fluvial origin for the conglomerates is indicated by the clast-support, imbrication, large-scale cross-bedding and moderate sorting and rounding. The interbedded shale and siltstone are overbank flood plain facies. The normal grading within the conglomerates, their thickness and lateral continuity and the abundance of finer grained clastic facies in the section indicate a meandering and not braided fluvial system dominated the flood plain (Selley, 1985), which was established on a pre-existing pediment (the unconformity is quite planar). Because of the general coarseness of the sediment, basal channel floor and point bar facies within the fluvial deposits are not easily distinguished. To the south, the member becomes more immature and is dominated by massive breccias - probably debris flows on an alluvial fan, and more proximal in facies than to the north. This may indicate that higher relief, such as one of the high-angle fault scarps, lay further south, but this inferred topographic feature is not preserved. The high-angle fault observed in the area is too young and in the wrong place to be such a source. Possibly member I was deposited during a period of local graben activity and pediment formation on the horsts, as well as during the first stages of compressive orogeny to the east when only Happy Creek lithologies were exposed (this latter geographic assertion will be addressed later in this chapter).

An abrupt change in provenance is indicated by member J. The member, unlike I, is lacking in volcanic debris, and is made up of lithic debris from the unroofing of the McGill Canyon and Bliss Canyon Formations after the Happy Creek was stripped off. The sedimentary structures and interpretation are very similar to member I. However, a sudden tectonic event changed the character of the sourceland. Member J indicates the first definite activity on the thrust system, with strata as deep as the McGill Canyon Formation exposed and eroded.

Member K indicates another change in sourceland nature, back to andesitic volcanics. Member K may record the propagation of the thrust system nearer with the exposure of Happy Creek rocks in the newer frontal thrusts. The interfingering of J and K may record the tectonic emergence, erosion and mantling of various units dominated by andesitic (Happy Creek) and cherty/lithic (Bliss Canyon and McGill Canyon) rocks. The lack of component mixing is somewhat puzzling, and may reflect proximity to the source and a distinct nature of the drainages.

The basal conglomeratic horizons of member K, in conjunction with the overlying well-bedded arenites, make up a very large meandering channel feature. The conglomerates are the channel floor deposits, overlain by point bar pebbly arenite to siltstone beds. The uniformity, low angle lateral accretion bedding style, and internal ripple-bedding support the point bar origin of these beds. If this is so, relief in the channel could have been great.

The depositional setting of the portion of member J overlying K is not so clear. A subaerial setting is indicated by the common plant fossils, including whole logs. The section is better sorted, finer grained than those below, and is deficient in conglomerates, but also generally lacks finer clastic facies. The best interpretation to be made in light of the lack of distinguishing features, is of a braided alluvial flood plain, indicating perhaps an increased influx of sediment as the orogenic zone to the east approached and then choked the drainage with debris.

Member N records the immediate impingement of the thrust front on the area. The basal siltstone may have been laid down in depressions in front of and caused by loading due to the advancing thrust sheet. The conglomerate and then breccia were shed directly off of the thrust scarp, and their increasing immaturity and coarseness record its approach. The depositional environment was the alluvial slope (conglomerate) and talus apron (breccia) in front of the fault scarp. These deposits were in turn overridden by the thrust that shed them. Either thrust faulting took place earlier here than in sections further west (from the thrust front) or large portions of the King Lear were stripped off, as the upper portion of the formation seen elsewhere is missing.

Syn-sedimentary volcanic activity in the section is indicated by several white rhyolitic ash tuffs. These are two lower and not laterally extensive beds (they are also seen in area 9 and 10 in the southern Jackson Creek and Bliss Canyon drainages). There is also the Jackson Tuff), the grey reworked vitric tuffs, and the andesitic dikes.

Syn-depositional movement on the high-angle faults (expressed at the surface as monoclinal warping with an interlimb angle of 90°) is also documented in this area, and is constrained to have taken place before the Jackson Tuff, member N and the emplacement of the overlying thrust. This phase also probably took place after the deposition of member I.

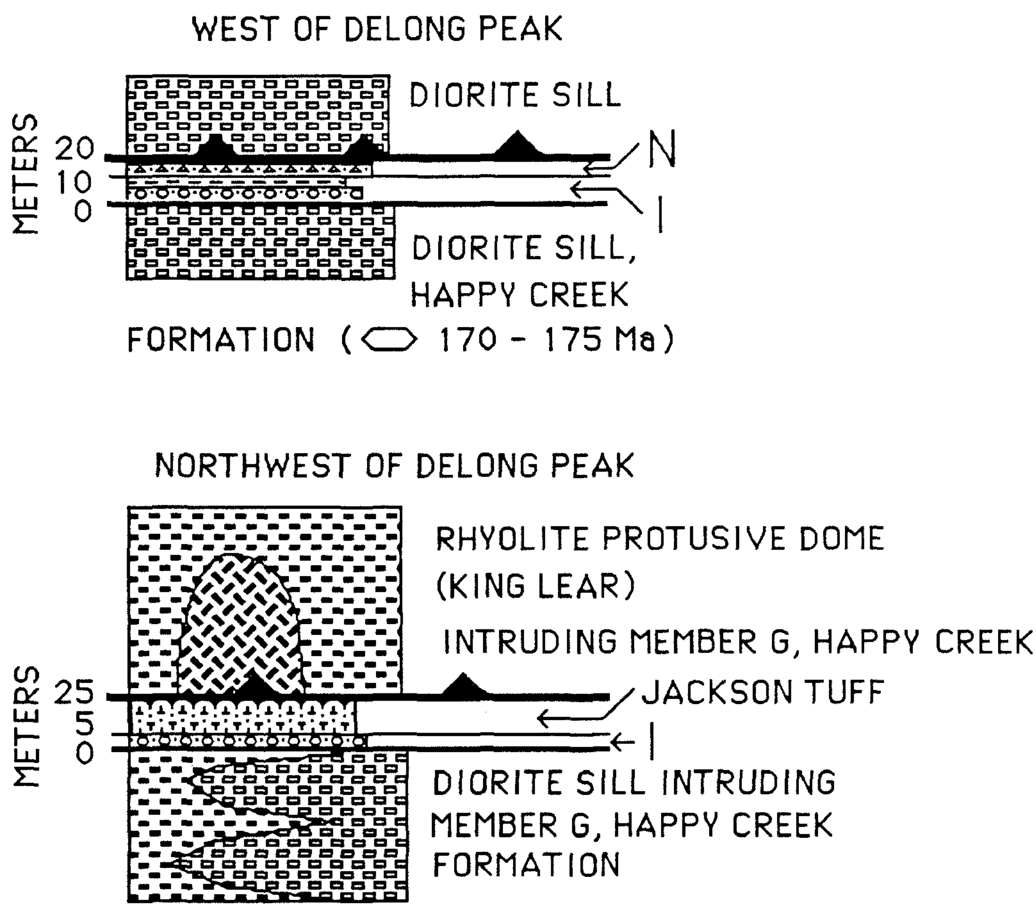
(6) DeLong Peak

This area contains several different and small exposures of King Lear sediments, volcanic facies and related hypabyssals within the eastern thrust belt, and exposed on the western slopes of DeLong Peak (both above and south of area 5). The two sections shown are tectonically sandwiched beneath the thrust faults, and rest unconformably on top of Happy Creek in different structural positions. One section is due west of DeLong Peak, the other on a saddle northwest of the peak. Additional small exposures are also discussed. See fig. 39.

<u>Stratigraphy</u> In at least five areas, two of them well-exposed on road cuts in the mining areas, sedimentary fault scarp breccias of member N are present. These rocks are characteristic of member N; other members also crop out in limited fashion.

In the best exposed section (right below the Iron King Mine, and west of DeLong Peak), the King Lear rests nonconformably on Happy Creek diorite with limestone nodules developed at the unconformity (calcrete related to soil development). Volcanic conglomerates form the bottom 5 m of the section, with large rounded clasts in clast-support a red muddy matrix. The matrix is almost certainly a later sieve deposit or soil feature because of the wide difference in energy indicated by the two fractions, and implies channel abandonment. This bed is overlain by 5 m of grey and green shale and siltstone, which also include limestone nodules. All of this is placed in member I. Above is 10 m of angular breccia, in beds 30 to 50 cm thick, making up the rest of the King Lear here - member N. The breccias have inverse-grading at the base and normal grading at the top, are in matrix- to clast-support, have a muddy matrix to an open-framework (the latter particularly in the center) and

FIG.39
6 DELONG PEAK
STRATIGRAPHY OF THE KING LEAR FORMATION



are topped with cross-laminated red arenite and siltstone lenses. In thin section a sample of arenite was composed of angular volcanic rock fragments in clast-support, in a replacive hematitic cement with floating quartz and feldspar grains.

Other outcrops are not so well exposed, and often only the upper sedimentary breccia facies is present. The clasts always represent lithologies identifiable in the overlying thrust plate of Happy Creek Formation, with no other components. As they contain detrital magnetite and epidote, they post-date these replacement and mineralization phases in the Happy Creek.

Intruding the Happy Creek in several thrust slices in this area are white to very light green or pink hypabyssal rhyolite bodies - several dikes in the Iron King Mine, and a small plug in the headwaters of Jackson Creek. The rhyolite is fine-grained but not glassy, often flow-banded at the edges, and has quartz and rare feldspar phenocrysts (the latter are highly sericitized, and the quartz corroded). In thin section the ã-quartz is hexagonal and embayed and the feldspar is sanidine $\pm$ andesine, and there is minor biotite and FeTi-oxide. The groundmass is locally glassy and has highly sheared siliceous fluidal banding. The rhyolite hypabyssal intrusives are not affected by the hydrothermal metamorphism and mineralization prevalent in the surrounding Happy Creek. The plug has columnar jointing and certainly does not cut any thrusts, and the dikes have been brecciated tectonically during thrusting. These rhyolitic bodies are placed in the Jackson Tuff.

In addition, beneath the thrust plate containing the plug is a small exposure of King Lear sediments. This section (situated northwest of

DeLong Peak) has a basal section of 5 m of red sedimentary breccia, which includes magnetite chunks and altered clasts of Happy Creek lithologies (and is assigned to member I). Overlying this is as much as 20 m of rhyolitic epiclastics and volcaniclastics, inferred to be a lateral equivalent of (and placed in) the Jackson Tuff. A conglomerate with rounded pebbles in a grey silicified arenite matrix (all rhyolitic) is overlain by sedimentary rhyolitic breccia and an airfall tuff with a black obsidian base, fiamme and lapilli (possibly the Jackson Tuff itself). The section is then structurally overlain by a thrust slice of Happy Creek. This occurrence of these rhyolitic volcanics so close to the rhyolitic hypabyssal plug (even though some telescoping by thrusting took place) is probably not coincidence; the two are inferred to be genetically related. These lithologies are similar to rhyolitic volcanic centers seen to the southwest in the lowlands of Jackson Creek in area 8 and around Buff Peak to the north in area 1, but are older than the Buff Peak center and probably older than the Jackson Creek complex. The rhyolitic volcaniclastic here are especially similar to the Jackson Tuff and its feeder dikes and are of about the same age, and are therefore correlated with them.

Age As with area 5, the fact that several of the sections with the sedimentary fault scarp breccia facies overlie the diorite sill means that they are younger than that sill, which has a preliminary date of 170 - 175 m.y. As some of the thrust faults (though not the frontal-most and youngest ones) are cut by a 162 $\pm$ 1 m.y. (U/Pb on zircons) pluton to the south, some of these structurally higher fault scarp breccias are constrained to the Callovian. Others (strucutrally lower) are related to the younger thrusts, which are Aptian (and perhaps slightly older).

Depositional Environment The thin basal occurrences of member I - red volcanic conglomerates - are probably pediment cover, deposited by fluvial processes across a planated rock surface. Member N, on the other hand, was deposited in a setting in front of an active fault scarp, with the alluvial apron shed from that topographic feature and preserved immediately beneath the overriding parent thrust fault. The mode of deposition was by thin, matrix-poor and therefore proximal debris-flows, with waning stage sheet-floods recorded by the cross-laminated finer clastics at the tops. Hematitic cement indicates a semi-arid environment. The shale, siltstone and grey micritic limestone underlying the breccia in several places may have been laid down in ponded depressions in front of the scarp. The section west of Delong Peak has a structurally higher position than others, and either thrusting took place quite early (so that much of the stratigraphy seen elsewhere was not deposited) or those upper parts of the succession were stripped off before the thrust package actually overrode the area, or both.

The tuffaceous rhyolitic interval (overlying a thin succession of member I) records epiclastic sedimentation proximal to rhyolitic volcanism (the conglomerate and breccia) as well as the volcanism itself. The manner of epiclastic deposition is not clear, but was probably alluvial. The tuff is similar to known examples of the Jackson Tuff, and is interpreted to be the same welded and ignimbritic tuff. It resulted from the collapse of an eruptive column during a plinian phreatomagmatic event. Based on stratigraphic position, this site and the associated rhyolitic plug are interpreted to be associated with the event that produced the Jackson Tuff, rather than the rhyolitic volcanic centers in areas 1 and 8. The rhyolitic volcanism here also definitely predates and is affected by the east-vergent thrusting in this part of the Jackson Mountains.

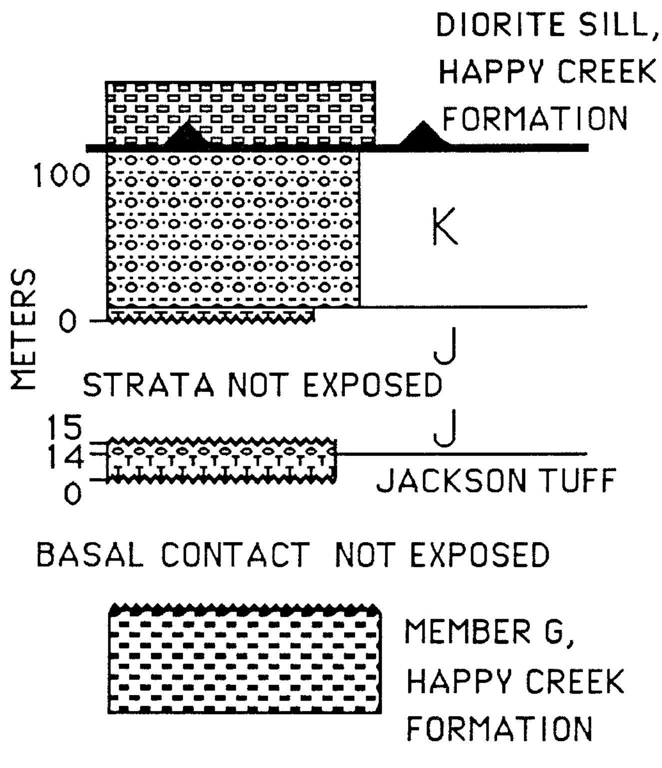
(7) Trout Creek

This section lies in a small valley in the lower headwaters of Trout Creek, on the west side of the Trout Creek Spur. Structurally, it is in an overturned syncline within the thrust belt, plunging north and truncated on the north by a high-angle fault. The exposure was not measured, and stratigraphic positions and thicknesses are approximate. See stratigraphic section 7 (fig. 40).

Stratigraphy The basal unconformity itself does not actually crop out. The lowest strata seen in the section (member J) were green siltstones. These are overlain by a thick sequence (member K) of red volcanic arenite, volcanic pebble conglomerate and siltstone. Boulders up to 1 m in size, of pink granite and of the Early Mesozoic Intrusive Suite are present, and the provenance of the rest of the detritus is Happy Creek. The red beds are stratified on a meter scale, and are cross-laminated and normally graded. Generally, thick cobble conglomerate beds are lacking. Bedding is laterally and vertically variable, and the clasts are rounded. This interval is 100 m thick or more.

The Jackson Tuff crops out to the west of the exposures of member I,

FIG.40
7 TROUT CREEK
STRATIGRAPHY OF THE KING LEAR FORMATION



structurally underneath the outcrop of member K discussed above. The bed is of uncertain but substantial thickness (probably greater than 10 m). The tuff is capped by a 1 m thick bed of conglomerate, with a framework of large rounded clasts of grey chert and quartzite and rhyolitic breccia fragments in a white tuffaceous matrix.

Age No data.

Depositional Environment This section is most similar to that measured in area 9, in the south Jackson Creek area. Member K is subaerial (hematitic cement) and alluvial (cross-lamination, scouring and grain-size). Because of the haphazard lateral and horizontal variability and the lack of the characteristic meandering channel facies and extensive overbank shales and siltstone it is interpreted as having been deposited in a braided alluvial fan. The normal grading records deposition during waning flood events. The deposits were probably distal, judging by the silty to pebbly range in grain size. Boulders the size of those seen are easily carried out into even such a distal environment by catastrophic debris flow events (McPhee, 1988).

The conglomeratic bed on top of the Jackson Tuff represents alluvial reworking of debris from the tuff (the rhyolitic breccia fragments and tuffaceous matrix) in with clastic sediment from member J (the cherty/lithic component). This supports contemporaneity of tuff eruption and the alluvial sedimentation of member J.

(8) Jackson Creek

Within the broad valley of Jackson Creek in the central part of the range, several knobs assert themselves west of the intersection of the Jungo and Iron King mine roads. These outcrops are within the central autochthonous zone of the range, and have not been internally transported by thrust faulting (though an anticline does run N30E through the region). On the eastern edge a thrust fault places Happy Creek and King Lear rocks above the rocks discussed in this section. The outcrop discussed here is bounded on the south and north by high-angle faults. See fig. 41.

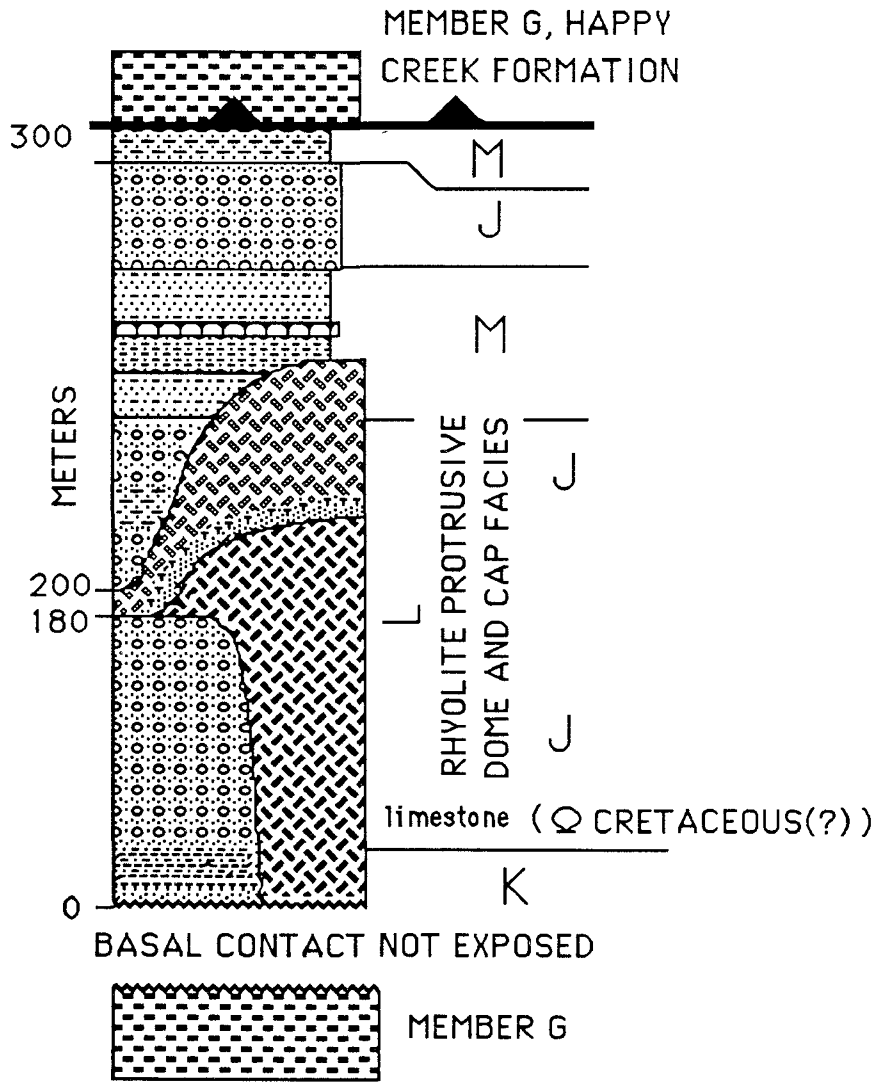
Stratigraphy The main stratigraphic framework of this area is that of a rhyolitic dome complex intruding into and overlapped by cherty/lithic and volcanic epiclastic sedimentary facies of the King Lear. The depositional contact with the Happy Creek is not exposed.

The rhyolite dome is cored by a very distinctive breccia unit. This breccia is composed of highly angular fragments a few mm to 25 cm in size, milled and broken with a size-frequency distribution such that there are equivalent numbers of particles in each size range. The fragments are usually hydrothermally bleached, in a framework in a very fine-grained matrix. Veins of pink, brown or black obsidian seam the rock. In some instances flow-banding and epiclastic textures from dome-related facies are discernable in these pieces. In thin section the breccia is made up of quartz and andesine grains and a framework of rhyolitic angular fragments in a devitrified glassy groundmass. The breccia itself can exhibit a size-based foliation. The colors are highly

FIG.41

8 JACKSON CREEK

STRATIGRAPHY OF THE KING LEAR FORMATION



variable in the pastel range, from pink to green to white. Some fragments (smaller and rarer) of siltstone, chertarenite, andesite, and chert as well as individual plagioclase grains are also present in the breccia zone. The obsidian dikes display all gradations of hydrothermal bleaching from glassy and fresh to highly devitrified and almost unrecognizable. In thin section one such sample was entirely composed of yellow glass with perlitic cracking; another had quartz and sanidine phenocrysts as well. The breccia dome overall has a mushroom-like shape.

Consistently present as a cap over the upper boundary of the dome (the lower contact is intrusive into King Lear sedimentary facies) is a set of three facies: (i) rhyolitic epiclastics; (ii) obsidian and tuff; and (iii) rhyolite flows. The epiclastics were shed entirely from the dome, and commonly consist of green sedimentary rhyolitic coarse-grained and immature arenite and conglomerate, in some sites containing accretionary pisolites. The pebbles of bleached tan to green rhyolite are rounded and in clast-support in a fine-grained arenitic matrix. The interval coarsens upwards at the base and fines again at the top, and locally grades into a green sedimentary rhyolite breccia, which locally are orange, brown and red in color. Lateral variability is pronounced, and the entire interval is only a few meters thick. In thin section the rhyolitic conglomerate at the base of the cap facies is composed entirely of volcanic rock fragments, subrounded to angular and bimodal - siliceous volcanic pebbles in a rhyolitic arenite matrix. Sanidine, biotite, andesine and quartz phenocrysts are present in the siliceous rock fragments. The thickness, degree of sorting and rounding and the grain size vary throughout this epiclastic cap facies.

Overlying the epiclastics in patchy outcrop, but distributed over the entire area, is black obsidian (rather fresh and only locally devitrified), which can display fiamme and grade into a white rhyolitic lapilli tuff. The obsidian zone in the cap facies is characterized by clear fiamme in a brown glass matrix with a few percent topaz and quartz grains. The fiamme are deformed and compacted.

On top of the obsidian is a thick and extensive rhyolite flow succession. The rhyolite is pink to orange to red to purple to grey, with quartz phenocrysts up to several percent by volume. There is contorted and deformed siliceous flow banding. In some places the flows have a few percent unfilled amygdules. Rhyolitic flow-breccias also occur at both the top and bottom. The texture is very fine-grained to glassy, and some of the breccia fragments in the flows are ductilely stretched. In thin section the rhyolite flows have topaz, quartz (large euhedral but embayed grains), sanidine, andesine, and biotite floating in a tan glassy matrix with siliceous, clear flow laminae and patches of tridymite growth. The sum of the phenocrysts is always less than 10%.

This whole rhyolitic complex is part of member L. Based on stratigraphic position it does not appear to be intimately related to the Jackson Tuff (though the lithologies are quite similar). Both the epiclastic facies (i) and the rhyolite flows (iii) are found in section 9 to the south, interbedded with member M (and separated from one another by 200 m of section, supporting the idea of resurgence and reactivation within the dome after some considerable span of time).

The dome has an intrusive contact into (and the lip of the top of the dome as well as the cap facies overlap) poorly-exposed sediments interpreted as belonging to members J and K. The sediments include arenite, thin tuff beds, red siltstone and shale and chert-pebble conglomerate (in stratigraphic order of appearance), which total about 180 m thick. The cap facies pinches out away from the central breccia dome, and are overlain by more unrelated sediments of members J and M. Though the actual depositional contact is not exposed, there is no evidence for a structural break.

The overlying section, approximately 300 m thick, is composed of (in stratigraphic order): (i) well-bedded chert-pebble conglomerate, lithic arenite and brown siltstone; (ii) reddish-brown to greyish-brown lithic arenite, cross-laminated to planar laminated, burrowed and well bedded; (iii) more chert-pebble conglomerate; (iv) red arkose with siltstone intraclasts and siltstone; (v) green siltstone; (vi) fine-grained green siliceous cross-laminated tuff; (vii) green siltstone and arenite; (viii) more chert-pebble conglomerate; and finally (ix) black shale and greyish-brown siltstone and arkose. The nature of outcrop did not permit an estimate of individual thicknesses. Intervals (i) to (iii) and (viii) are placed in member J, while (iv) to (vii) and (ix) are assigned to member M (this correlation is based on much better stratigraphic exposures to the south in area 9).

Across Jackson Creek and also in area 8, just east of the mouth of its north fork and just south of a high-angle fault, a poorly exposed rhyolitic lithology crops out. It is situated directly over member G of the Happy Creek. This outcrop is assigned to the Jackson Tuff; no other identifiable strata are exposed. A thrust places member F of the Happy Creek on top of this occurrence.

Post-depositional dextral movement along an east-west fault separating areas 8 and 9 is indicated by the stratigraphic discordance between those two sections across that fault. Significant basement relief is also implied by the greater stratigraphic thicknesses below the fingers of cap facies i and iii in area 9 to the south compared with this area. This relief is inherited from member G of the Happy Creek Formation, which is much thicker below 8 than 9 (volcanic constructional relief); hence 8 was deposited further up on a topographic high.

The northern bounding structure was active throughout King Lear deposition. For example, between the lowest lahar bed of member M and the top of the Jackson Tuff there is almost 800 m of section in area 9, but only 160 m in that interval in area 4 to the north. The fault had continued dip-slip activity post-dating the Jackson Tuff, unlike the other structures in this generation (apart from their local reactivation as tear faults during thrusting).

Age The rhyolite flows of the cap facies were collected and processed for zircon for U/Pb dating, but unfortunately no zircon was present. A collection from a thin grey limestone lens in the section (below and intruded by the dome) examined by J.B. Reeside is reported by Willden (1963) to be of Cretaceous(?) age. The fossils include gastropods, pelecypods, ostracodes, and various algal forms of a fresh-water origin.

Depositional Environment The lower, poorly exposed sedimentary sequence of I and J is alluvial (based partly on exposures elsewhere) and contained fresh-water lacustrine environments. It is worth noting that silicic volcanism was already a factor this early - thin rhyolitic tuff beds underlying the Jackson Tuff and predating the Jackson Creek dome (discussed here) are found throughout the central portion of the range.

The interpretation of the central breccia dome is as a very shallow to emergent protrusive rhyolite dome, about one kilometer across and with emergent relief of several hundred meters. The breccia core was repeatedly mobilized and extruded with infusions of new rhyolitic melt (as documented by the textural history of rhyolite intrusion, bleaching by hydrothermal activity and brecciation due to remobilization, then more obsidian infusion and repeated bleaching and brecciation). The paleosurface is interpreted to be where the dome abruptly widens and spreads out somewhat over non-rhyolitic sediments in a small lip. A small amount of rhyolitic epiclastic debris underlies this lip. The cap facies indicate (i) a period of alluvial epiclastic reworking and spalling of the dome with some ash emission (the latter indicated by the accretionary pisolites - Cas and Wright, 1987), followed by (ii) dome resurgence with the eruption of an ignimbritic tuff and then (iii) the extrusion of thick, viscous rhyolite flows. The latter covered the dome and flowed away from it, with an observed lateral extent of over 5 km or more and a thickness of at least several tens of meters. The dome was then gradually overlapped and covered by sediments. The Jackson Tuff but is not inferred to be genetically related to either the initial or resurgent phases of evolution of the dome complex; it is an earlier event. Though rather similar lithologically and in evolutionary sequence to the Buff Peak rhyolitic center, the Jackson Creek dome is quite distinctly stratigraphically lower than it, and higher than the Jackson

Tuff.

The sediments, which are inferred to overlap the dome and are found as xenoliths within it indicate that the latter had erupted up through a distal alluvial fan or alluvial flood-plain setting. The typical fluvial chert-pebble conglomerates of J, red and green coloration of the sediments and the abundance of finer shale and siltstone intervals may indicate a meandering alluvial flood-plain nature. The source appears to have switched between the andesitic epiclastics of K and the cherty/lithic provenance of J (both orogenically produced), and then between J and the plagioclase-rich silicic volcanic province of M, with reworked siliceous crystal tuff beds. Member N was not present in the section, though the upper plate of a thrust fault bounds the east side and top of the section. Member N is present on top of this thrust package, unconformably overlying a nappe of Happy Creek.

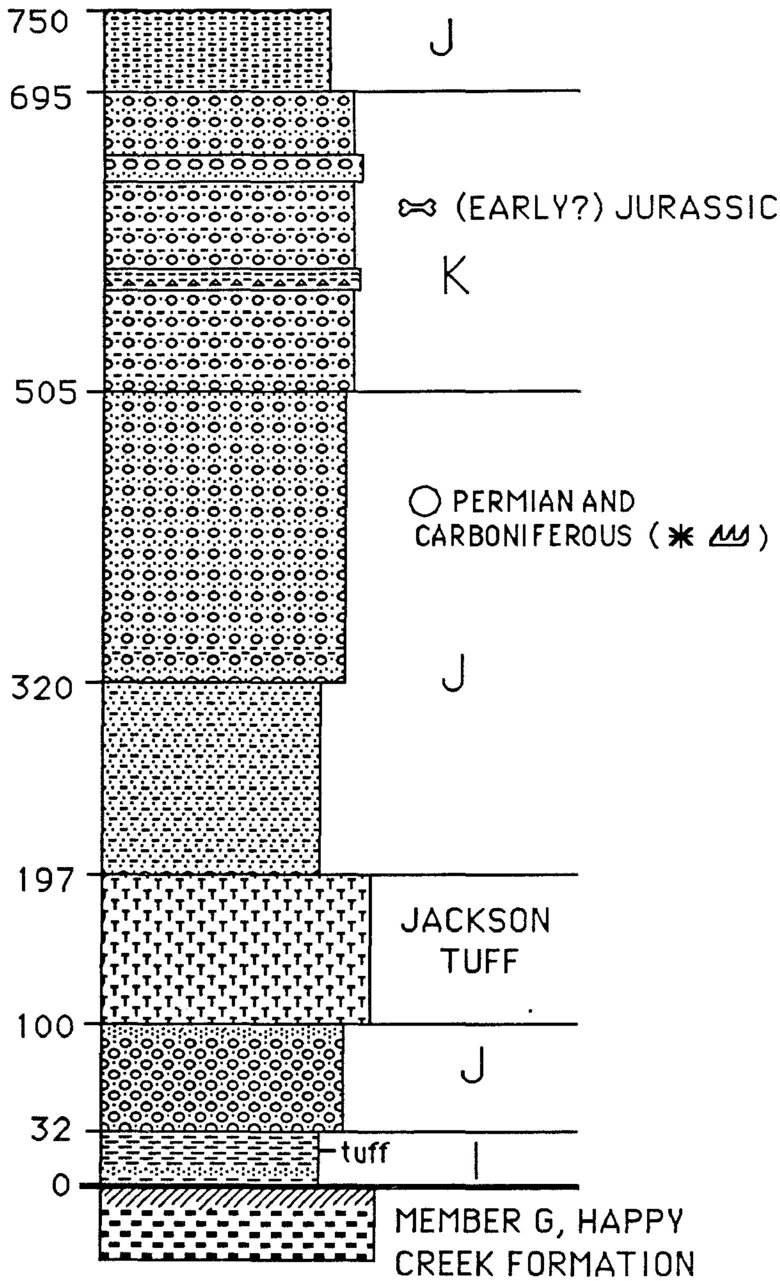
(9) South Jackson Creek

This is the individually most extensive and best exposed section of the King Lear Formation in the range. If any section were to be chosen for a type section, this one would best qualify. The area of exposure lies south of Jackson Creek, west of Trout Creek, and north of Hobo Canyon. It is bounded on the north by a high-angle fault, which separates it from area 8 (Jackson Creek), on the west by the basal conformable (member G) to disconformable (member F) contact with the Happy Creek Formation, on the east by Quaternary sedimentary cover, and on the south by the dacitic intrusive laccolith complex of area 10 (Bliss Canyon). The section dips homoclinally to the southeast and is not internally deformed, though several sills do extend into the strata from the south. The section was measured with a Jacobs staff. See fig. 42.

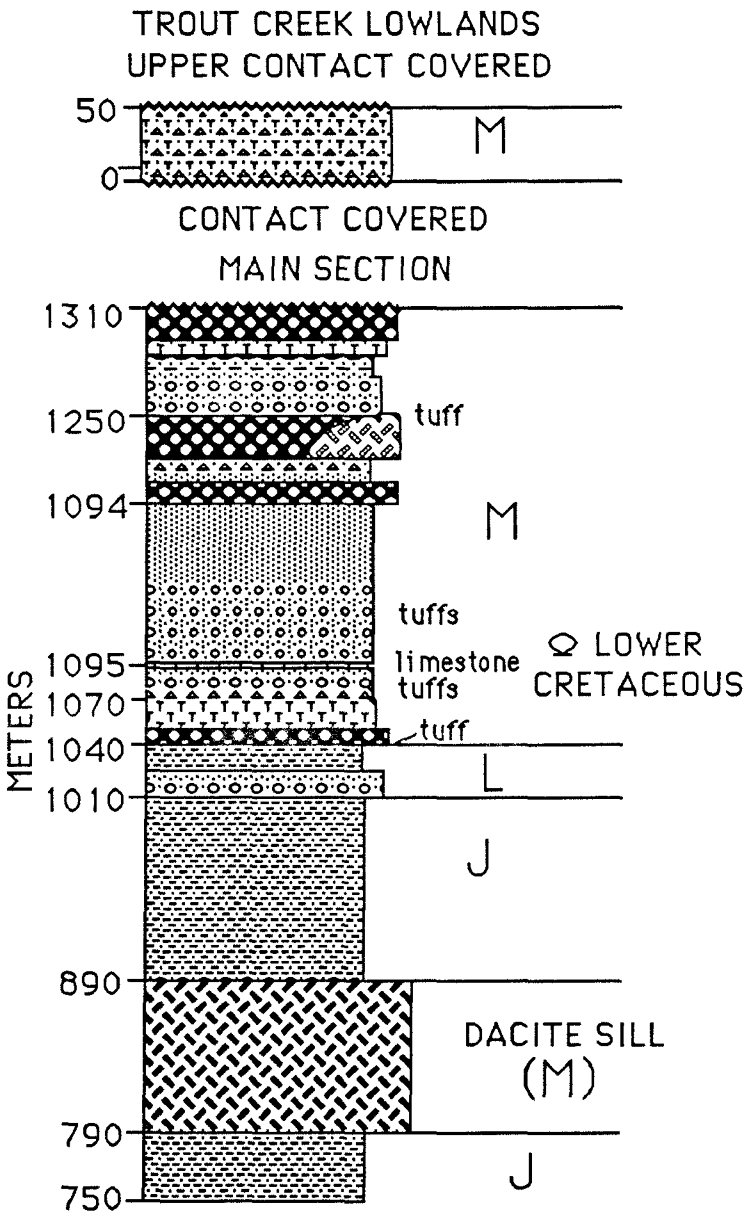
Stratigraphy At the base of the section in the northern part of the area, an 8 m thick tan volcanic conglomerate to pebbly arenite bed (with rounded clasts of Happy Creek andesitic rocks) overlies andesite flows, and is in turn overlain by another thick grey andesite flow and then by more King Lear sediments. The epiclastics pinch out to the north. The Happy Creek and King Lear Formations are thus actually interbedded in this vicinity. The sandstone petrographically is a volcanic arkose, composed of detrital andesine plagioclase, volcanic rock fragments and Fe-oxides with minor hornblende and trace quartz and apatite in a silty arkosic matrix. The texture has moderate sorting and angular grains (many of the plagioclase retain their original euhedral lath shape), and lacks clay size fractions.

Along the section line further south (on the divide between Jackson Creek and Hobo Canyon), the uppermost andesite of the Happy Creek has a subaerially weathered zone on top 6 to 8 m thick, but lacks the calcrete seams. Above is, however, 1.5 m of sandy to pure grey very fine-grained carbonate, veined, cracked and knobby-weathering - this is a calcrete horizon. Overlying this is 2.5 m of grey to green shale and siltstone with limestone concretions, then 3 m of red pebbly volcanic arenite and 12 m more of green to grey shale and siltstone, with limestone concretions best developed near the top. At 19 m is a 1 m bed of white to very light gray grainy vitric ash. This tuff is well-laminated with compacted glass shards at the base and shale rip-ups at the top. In thin

FIG. 42
9 SOUTH JACKSON CREEK
STRATIGRAPHY OF THE KING LEAR FORMATION



9 SOUTH JACKSON CREEK (CONTINUED)



section the tuff is made up of ductilely deformed and welded devitrified fiamme and xenoliths (siliceous volcanic and volcanic arenite) and is depositionally laminated (seen in the proportion of fragmental material). From 20 to 32 m is greenish-grey silty shale and very fine-grained arenite. This basal interval all belongs to member I.

At 32 m a change in provenance takes place, with cherty and lithic debris suddenly dominating. The base is a green chert pebble- to cobble-conglomerate with clasts of black and green chert (veined), quartzite, vein quartz, and rare granophyre. Above this 2 m thick bed and extending to 83 m in the section is similar pebble conglomerate to pebbly arenite. These beds exhibit clast-support and clast-imbrication. Another cobble conglomerate lens is found at 78 to 80 m. At 83 m the section abruptly fines, with green fine-grained chertarenite and siltstone (no pebbles) to 100 meters. The chert arenite in one thin section is composed of detrital grains (sorted, and subangular to subrounded) of monocrystalline quartz, mosaic and deformed quartz, chert, sedimentary rock fragments (argillite, probably not intraformational, but from the Bliss Canyon and McGill Canyon Formations as there is little mudstone at this level in the King Lear), and with a greenish-brown micaceous fibrous cement (inferred to be chlorite).

At 100 m a 97 m thick occurrence of the Jackson Tuff is present. The tuff is white to very light green and grainy and contains 5 to 10 % large euhedral quartz crystals and trace feldspar and hornblende grains. Columnar jointing is well-developed in the middle of the bed. Some welding is present in the basal 10 m, and along strike this zone can verge on the glassy. Above this zone the tuff is homogenous and massive.

The middle of the tuff in thin section is a crystal tuff, with 10 to 15 percent by volume quartz (large and corroded), with subordinate andesine, sanidine, biotite and hornblende grains floating in a homogenous devitrified very fine-grained siliceous groundmass (not glassy as at the base). The quartz is locally graphically intergrown with the alkali feldspar. The basal portion can be a black to dark green glass, with vitroclastic texures well-preserved, and crystals of andesine and quartz. The glass is overprinted by thin acicular needles inferred to be cristobalite and feldspar (Williams et al., 1982).

From 197 meters to 320 meters, the section is recessive and relatively poorly exposed. The interval is made up of green to greenish-gray siltstone and sandstone, laminated and fissile to thinly bedded and, in some instances, cross-laminated. The coarser-grained facies includes wackes, with poorly sorted and subangular detrital plagioclase and quartz grains and secondary volcanic rock fragments supported in a clay matrix. In the top 15 meters, discontinuous limestone nodule horizons are developed, and a meter-thick bed of grey sandy carbonate caps the interval. The carbonate bed is composed of calcitic micrite with a few percent feldspar, quartz and lithic grains (being replaced by calcite and chert) in a homogenous very fine-grained matrix. Overlying this is a 14 meter thick succession of green chert-pebble conglomerate (with the same sedimentary textures seen in this facies lower in the section). This passes upwards to 17 m of greenish-gray siltstone and very fine-grained chertarenite with thin lenses of chert-pebble conglomerate. Clasts in these lenses can reach 40 cm in size. This grades up to 152 meters more of green chert-pebble conglomerate with lenses of green chertarenite, extremely homogenous and featureless. The interval, from 197 to 505 meters, is member J.

Overlying J with a sharp contact is a thick succession dominated by volcanic epiclastic sediments - member K. The base is red quartzose breccia, 2 m thick with a red shaly matrix to white carbonate cement; a similar bed extends from 575 to 580 m. In thin section one of these breccias is composed of very angular but well sorted volcanic rock fragments of scoriaceous and plagioclase phryic dacite or andesite, in a very coarse hematitic or calcitic granular cement or a matrix of arkose (angular plagioclase grains). The volcanic rock fragments in this bed are commonly highly hematized, and are in a closed framework. Above the basal breccia are more red beds - dark red siltstone, volcanic arenite and volcanic conglomerate. Chert clasts are a component in the latter only at the base of this interval. Limestone nodules are common. Cross-lamination is present in the arenites, and clast-support texture is dominant. Diamictite beds are present at 540, 573 to 575, 615 to 645, and 695 to 697 m, with some channel cutting and scouring at their bases, and containing accretionary pisolites. The diamictite in one thin section is poorly sorted, and has angular particles and a sandy and silty matrix with volcanic and micrite rock fragments, detrital quartz, chert and plagioclase. This bed displays a two phase cement of hematite and then calcite. One conglomeratic lens at 656 m contains boulders of the pink monzonitic intrusive. Several thin chert-pebble conglomerates interfinger with the red beds at 525 and 571 m, and from 535 to 539 and 580 to 581 m there are also beds of white siliceous conglomeratic sandstone. The red bed succession extends to 697. Finer-grained facies are notably lacking (a small amount of green shale exists at 650 m, at the top of fining upwards sequence of a red volcanic arenite 5 m thick). I also found the distal end of the femur of a small carnosaurian dinosaur (a coelophysid - S. Welles, written communication, 1985) here.

At 697 m, K is overlain by member J. Green and grey siltstone and shale with minor amounts of green chert arenite and chert-pebble conglomerate begins, extending to 790 m. The succession is homogenous and rather fine-grained, and lacks a volcanic epiclastic contribution. A discontinuous limestone horizon (0 to 30 cm thick) is present at 720 meters, with an almost coqunoidal texture of small bivalve and other fossil fragments. Thin section analysis of this lithology revealed it to be a burrowed biomicrite (texturally a floatstone to packstone) with gastropod, ostracod and bivalve skeletal particles supported by or with interstitial calcite micrite. The burrows are recrystallized to a calcite spar. Large vertebrate (and probably dinosaur) bones are also sparsely scattered through this interval.

At 790 m, one of the dacitic sills intrudes the section. The sill is about 100 meters thick on the section line and thins to nothing northwards over about 1 km. The sill is made up of green plagioclase phyric dacite, with plagioclase laths 1 cm in size and 20 % by volume, and with minor (less than 5 %) quartz blebs of the same size. The matrix is very fine-grained to aphanitic, hard and siliceous.

Above the sill member J resumes, to 1011 m (the thickness of the sill is included in the cumulative stratigraphic thickness). Green and grey shale and siltstone characterize the interval. Coarser-grained lithologies are notably lacking.

Abruptly overlying member J is a thin interval dominated by rhyolitic volcanism and epiclastic sedimentation and placed in member L. Provenance is quite distinct from the reworked andesitic epiclastics of I, K and N or the lithic/cherty debris of J, though similar to the dacitic volcanics and epiclastics of M. Resting on J with a scoured base is a 16 m thick succession of rhyolitic conglomerate and breccia (the basal 1 m or so) grading up rapidly to white, light grey or light green rhyolitic arenite. Thin bedding and cross-bedding and upwards- fining characterize the upper part. The coarser-grained debris, though extensively bleached, is unmistakably of the same lithology as the rhyolite dome to the north in Jackson Creek, with siliceous flow banding and quartz and feldspar phenocrysts. The arenitic matrix is a very poorly sorted, angular arkose (plagioclase) with significant chert, quartz and sanidine grains and siliceous volcanic rock fragments, and chlorite-illite or hematitic cement; no clays were present. This is overlain by 12 m more of green and grey shale and rhyolitic arenite of member L.

Another sudden change in stratigraphy takes place at 1040 m. Member M is readily distinguished by dacitic volcanic and epiclastic sedimentary rocks. The lowest bed in this interval is a 4 m thick black siliceous crystal tuff, with an aphanitic almost glassy matrix and a significant proportion of angular feldspar, quartz and biotite grains, accretionary lapilli and xenolithic lapilli of scoriaceous volcanics, and flow and intrusive lithology fragments. The tuff has size-graded laminations, and is partially scoured and eroded on a 1 m scale by the overlying bed. The overlying bed, extending to 1053 m, is a dark green diamictite with a green siliceous aphanitic matrix and white feldspar grains. Scattered angular clasts are supported by the clay matrix, and consist of green plagioclase phyric dacite (identical to the sill intruding the section lower down), Fe-rich green biotite, and large andesine grains (also from the dacitic volcanic complex). The base is inverse-graded, and the aphanitic black cherty top lacks clasts altogether (and is very similar to the underlying tuff). From 1053 to 1069 m is another black crystal tuff identical to the first, with conchoidal fracturing. The crystal tuffs contain angular fragments of quartz, plagioclase and/or sanidine, biotite and chlorite and trace zircon and devitrified siliceous glass fragments floating in a devitrified to glassy siliceous groundmass. No exotic xenoliths were present. The black tuff grades upwards over 2 m to a light green cross-laminated tuffaceous arenite and then to 19 m of light green dacitic to rhyolitic breccia, conglomerate and arenite. The conglomerate in thin section has angular to subangular rhyolitic pebbles (with tridymite devitrification) in a volcanic arenite matrix with detrital andesine and quartz. Several thin cross-laminated to planar laminated gray to green siliceous and aphanitic tuffs are present in this interval, and one has paleo-rain drops preserved. At 1093 to 1095 m there is a thin grey limestone bed (laterally persistent over a 100 m), well bedded, burrowed and fossiliferous. The limestone is a fossiliferous micrite in thin section, with a large percentage of skeletal grains (small delicate bivalve and ostracod remnants) in a micritic matrix. Scattered arenaceous grains are present. This is overlain from 1095 to 1194 m by green fine-grained breccias, conglomerates, arenites and flows of dacitic provenance. Angular clasts of green plagioclase phyric dacite (sometimes bleached white), and detrital quartz and plagioclase grains characterize the rock. Accretionary pisolites occasionally appear. In thin section the sandstone is characterized by poorly sorted, angular and broken sericitized plagioclase, subordinate biotite and some heavy mineral (zircon and apatite) grains (identifiable as coming from the dacitic volcanic complex) in a cherty cement. There can be a substantial detrital chert component as well. The flows (thin and rare) have large euhedral plagioclase, biotite and trace apatite in a siliceous groundmass (with apatite and biotite microlites and replacive chert and calcite). The interval is massive and featureless except at the top, where (at about 1190) there is another green silicified crystal tuff with accretionary lapilli, cut by a channel filled by a fine wacke. At the very top there is grey to tan siltstone.

The next interval is quite different in nature, though still in member M. The interval is characterized by a sequence of very thick bedded and massive green diamictite. These rocks have a very fine-grained feldspathic green siliceous matrix supporting very poorly sorted scoriaceous and dacitic clasts. The clast population also includes rarer chert, rhyolite, diorite and andesite fragments. The tops of the diamictites are locally cross-laminated and sandy, lacking the clasts, and the bases are scoured. The beds are composite and hard to distinguish from one another. To the south, the diamictites overlie a thick but laterally restricted loaf-shaped rhyolite flow and flow breccia unit, pink to orange in color and with quartz (but no feldspar) crystals - this flow is more like those in member L than other facies in M, and is placed in L. The map distribution could be the original flow boundaries, or the lahar might have been deposited after some erosion of the rhyolite flow. The diamictite interval extends to 1250 m, containing from 1209 to 1220 m a sequence of bedded and cross-bedded light grey to green quartzose and dacitic arenite, coarsening at the top to a dacitic breccia.

Overlying the diamictites are more recessive epiclastic beds, from 1250 to 1305 m. The lower part of this succession is made up of a 2 m light green xenolith-rich tuff, then another 4 meters of diamictite as described above. From 1256 to 1295 m the interval is thin bedded (2 to 80 cm, averaging a few tens of cm) red siltstone, arkose, shale and dacitic pebble conglomerate. The beds are normally graded and cross-laminated at the bases (except some of the conglomerates, which are reverse graded and in red mud matrix-support). The bedding is locally very regular in the finer-grained facies and verges on varving. Compressional soft-sediment deformation is present, with small thrust faults offsetting the graded beds. In thin section the finer-grained facies contain very angular, poorly sorted detrital quartz and plagioclase grains and very abundant siliceous volcanic rock fragments plus Fe-oxides and biotite. They have a hematitic, calcite or chert cement (or some combination of the three). The feldspar in the arkose is derived from, and the arkose grades compositionally into, the plagioclase phyric dacite breccias and conglomerates. This interval is topped by a 4 m thick dark green siliceous crystal tuff with an aphanitic, conchoidally-fracturing matrix. This tuff is cross-laminated and graded, and is composite. The uppermost 7 m of continuous section exposed (to 1306 m) is another massive green diamictite. Member N is not present here.

To the east in the lowlands of Trout creek about 1 km is an unusual outcrop. The beds strike north-south and dip east. They are composed of a few tens of meters of silicified breccia and conglomerate, with angular to subrounded clasts (homogenous to flow-banded dacitic volcanic) in a siliceous white detrital matrix. The deposit can be very clast-rich and cross-bedded to thinly bedded, or in matrix-support and homogenous, and can even be graded. A thin section of one sample had siliceous and rounded volcanic pebbles plus large corroded quartz grains and broken fragments of plagioclase in clast-support in a microcrystalline and siliceous groundmass with unidentified microlites. Interbedded with this lithology are thin siliceous orange crystal tuffs with an aphanitic matrix. The lithologies are quite distinct, but are much closer to member M in nature than to anything else in the range. They are not thought to be Tertiary, though ascribed to that period by Willden (1963), as they are monolithologic and very similar lithologically to know occurrences of member M, they share the structural attitude of the King Lear, and they are unlike any known Tertiary strata in the range. They are inferred to be part of member M, and are also inferred to be slightly offset by a west-vergent thrust, which is exposed to the south. The exact stratigraphic position is thus uncertain - they could overlie the strata of M described above or they could be a lateral and more proximal equivalent.

To the north of sections 8 and 9, within the large canyon of Jackson Creek, is a major east-west high-angle fault. Significant syndepositional movement on this fault is inferred from the greatly increased stratigraphic thicknesses in the King Lear to the south of this fault as compared to sections to the north (area 4, on the large Mesozoic horst). Later dextral strike-slip offset (tear faulting during thrusting) also took place on this fault, and on the one between areas 8 and 9.

<u>Age</u> The dinosaur femur from member K has been given a provisional date of Late Triassic through Jurassic, and perhaps Early Jurassic (S. Welles, written communication, 1985). Several collections from the limestones in members J and M were examined but yielded no better age than Mesozoic or younger (J. Hanley, written communication, 1986). Willden (1963) reports a collection (presumably from upper member J) analyzed by J.B. Reeside in 1957 as being of fresh-water origin and containing an Early Cretaceous fauna (gastropods, and ostracods or charophytes). A sample from the Jackson Tuff was also collected and zircon separation for U/Pb geochronology attempted, with no success. Two dacitic crystal tuffs from M (from the bottom and top of the unit) were found to have small amounts of zircon, and dating will be attempted in the future. The underlying Happy Creek (member G) is Bathonian. Member I, lower J and member K are hence late Middle or Late Jurassic in age, and the upper part of the section extends into the Early Cretaceous.

Pebbles from one of the chert-pebble conglomerates in member J were identified by D.L. Jones, as reported by Russell (1981), as having Carboniferous radiolaria and Permian conodonts and radiolaria.

Depositional Environment The interstratification of andesite flows of upper member G (which are inferred to be stratigraphically equivalent to occurences of member H to the north) and volcanic arkose, placed at the base of member I, indicate some overlap in time between the final phase of Happy Creek volcanism and the beginning of the dominantly epiclastic member I of the King Lear. The volcanic arkose and conglomerate are clearly derived from member G. The detrital andesine, minor hornblende and Fe-oxides and trace quartz and apatite, as well as andesitic pebbles in the coarser-grained facies, all are derived from that unit. Member I is fluvial in origin; the larger-scale paelogeographic setting is less clear.

This gradational and conformable basal contact is not present along the measured section, where a somewhat more disconformable relationship is evident. Here a significant but not extensively subaerial weathering profile is developed on the Happy Creek Formation, down to less than 10 m depth. The basal limestone horizon in the King Lear Formation is entirely calcrete and of caliche origin, or could have been a lacustrine deposit modified by these processes. The overlying green (reduced) shale and siltstone in member I support the latter conclusion; these are overbank flood plain deposits. The single layer of red volcanic conglomerate in I is a meandering fluvial channel deposit. The member was laid down in an alluvial flood plain dominated by meandering streams (and not, judging by the fine grain size dominating, an alluvial fan or braided system). Siliceous volcanic activity in the area is documented by the tuff bed, which is a thin welded ignimbrite to crystal vitric tuff with vitroclastic depositional layering preserved. Deposition from a pulsed plinian eruptive column with a distal source (note the bed thinness, good sorting and lack of coarse fragmentals) is inferred for this tuff (Cas and Wright, 1987), which probably correlates with those seen elsewhere in member I.

The transition from I to J, as elsewhere in the range, reflects orogenic uplift and exposure of deeper stratigraphic levels including, recognizably, the Bliss Canyon and McGill Canyon units. This is inferred from both lithologic correlation of the detrital components and by age of the clasts. The black and green chert, quartzite, vein quartz and argillite clasts in the conglomerates are distinctive in provenance, as are the quartz (monocrystalline to elongate polygonized to recrystallized mosaic), chert and SRFs (sedimentary rock fragments) in the arenites. The quartz population reflects a metamorphic origin (Young, 1976). Member J is characterized by several distinct assemblages: (i) Green chert pebble to cobble conglomerates in clast support with pebble imbrication, lenses of chertarenite, and thick to massive bedding. Volcanic epiclastic contributions are lacking, though the member does interfinger with member K. The thicker intervals reflect stationary and therefore long-lived composite meandering fluvial systems, while the thin lenses within the finer-grained facies are single migrating fluvial channel-fill sequences. These deposits are largely missing in the upper occurrences of the member. (ii) Green and grey shale, siltstone and fine-grained arenites and wackes, and grey fossiliferous micrite. The clastic facies are homogenous to planar laminated and fissile or cross-laminated. The environment of deposition is inferred to be the alluvial overbank and flood plain, with traction and gravity sedimentation from debris-laden laterally extensive flood events. The micrite can be arenaceous, but is composed principally of microcrystalline calcite with varying amounts of fresh-water faunal and algal remains. A quiet and organically rich but restricted environment (based on the limited and hardy fauna) with a patchy distribution is inferred for these limestones, such as a flood basin lacustrine setting.

The Jackson Tuff in this area represents a very large plinian eruptive event, and was deposited as a crystal-rich vitric ash-fall. No good evidence for an ignimbritic nature is present along the measured section (the base is not well-exposed, however). Further south within this area, however, the base of the bed does have a glassy welded ignimbritic texture indicating hot deposition; slow cooling of the lower part of this tuff is indicated by the immediately post-depositional and frozen growth of acicular microlites within the glass (Williams _et_ _al_., 1982). The lateral extent (25 km) and thickness of this unit (100 m) point to an unusual event; even the thickest portions exposed lack significant fragmental debris and thus are probably not too proximal to the vent.

Member K resulted from a renewed influx of epiclastic debris from the Happy Creek Formation (including andesitic facies) and the Early Mesozoic Intrusive Suite (the pink granite), reflecting the nearby orogenically-induced exposure and erosion of high-level thrust sheets of that unit. Some thin lenses of chert-pebble conglomerate are also present, with a source lower in the uplifted and stripped local section. The extensive hematization and hematite cement in the red beds point to a subaerial semi-arid setting. The diamictite beds are debris-flows with muddy matrix support of poorly sorted and angular clasts; some of the flows were mud-poor (voids filled by later cement). Given the association with accretionary pisolites (which form in volcanic eruptive columns), these flows might be volcanically induced, and were certainly not too distant (5 to 10 km) from their source. The volcanic conglomerates (clast-support and bimodal with erosive bases) and cross-laminated volcanic arenites are fluvial in origin. Clasts in the conglomerates can be boulder size. Finer-grained rocks are rare in this area, and occur at the tops of fining-upwards sequences a few meters thick. These are probably channel abandonment and fill features. The overall setting is inferred to be a braided alluvial fan rather than a meandering flood plain (because of the abundance of debris-flows, and paucity of fine-grained clastics and distinct channel sequences), and indicates K was more proximal to its source area than member J.

The dacitic sill intruding the section is a hypabyssal intrusion, injected along bedding from the laccolith complex to the south. It is also identical to some of the flows in member M, and is inferred to be comagmatic with the extrusive source for that member. The sill (and laccolith) is the hypabyssal equivalent of the epiclastics, flows and pyroclastics in member M.

Member L is dominated by rhyolitic epiclastic sedimentation, though other components, such as detrital chert, are present. The stratigraphy indicates a large meandering fluvial channel (the basal conglomeratic thalweg lag deposit and the overlying cross-bedded point bar arenites) grading up into flood plain finer-grained facies. It is also possible that this succession actually would be better placed in the base of member M; the siliceous volcanic component was not sufficiently diagnostic to be entirely certain.

Member M is clearly distinct from the other members in provenance, with a pronounced dacitic epiclastic and pyroclastic heritage. The green diamictite beds with scoured bases are volcanic debris flows (lahars), having a siliceous and aphanitic feldspathic mud matrix and containing dacitic extrusive debris. The overlying cross-laminated zones were caused by fluvial reworking and sheet flooding after the emplacement of the lahar. Non-volcanic debris flows with a red shaly matrix also are present. These proximal facies point to the existence of significant volcanic constructional relief. The rest of the lower and middle part of the member is composed of homogenous volcanic arenite and coarser-grained facies of a braided fluvial origin, with common ash-fall tuffs (these have preserved rain-drop impressions, indicating subaerial setting). Short-lived lacustrine interludes in this part are indicated by the limestone lenses, which were deposited in quiet, patchy, organically productive fresh-water settings. The uppermost epiclastic part of the section is rather different, dominated by finer-grained thin-bedded epiclastics with normal grading, and by debris flows. Both of these facies were subaqueous mass-flow events within a large lake (distal Bouma T_{c-e} turbidite sequences for the former, and debrite facies F_3 for the latter - Stow, 1985; Lowe, 1982). The difference is that the turbidites disaggregated to a fluid low-density mixture at some point after entering the lake and were redeposited from traction and suspension, while the debris flows were flowing underwater or subaerially as plastic, coherent high-density bodies. The turbidites may represent more distal and the debrites more proximal facies of similar events. The tuffs in this part of the section are also water-deposited, with cross-lamination and graded bedding. The notable abundance of crystal and vitric tuffs throughout M (and the presence within those tuffs of related extrusive fragments - a trait indicating a nearby vent), together with the common lahars, dacitic to rhyolitic flows and the accretionary pisolite occurrences also indicate that dacitic volcanism was ongoing and proximal through the deposition of this succession. The tuffs in some cases immediately underlie lahars, and are probably part of the same eruptive cycle. The early pyroclastic phase may have been followed a short time later by related mass flows. Thrusting was also going on during the deposition of the upper part of M, as compressional soft-sediment deformation is present in the lacustrine turbidite beds; indeed the lake may well have been caused by ponding in front of the advancing thrust scarp. The member thins rapidly to the north over a distance of 10 km.

The overall setting for member M was of a braided epiclastic alluvial fan being shed from a dacitic to rhyolitic volcanic center of significant relief. Epiclastic fluvial reworking and mass flow of debris from this center was punctuated by tuff eruption, lahar emplacement and flow extrusion. Late in the history of the member, continued epiclastic sedimentation took place into and volcanism occurred near a relatively deep and extensive lake ponded in front of an active thrust front.

The outcrop of siliceous volcanic epiclastic facies in the lowlands of Trout Creek appears to be an even more proximal (colluvial) facies, deposited by fluvial and slope-wash processes in the upper portions of the alluvial fan in and around the volcanic center. Frequent explosive eruptions from this center are indicated by the crystal tuff beds.

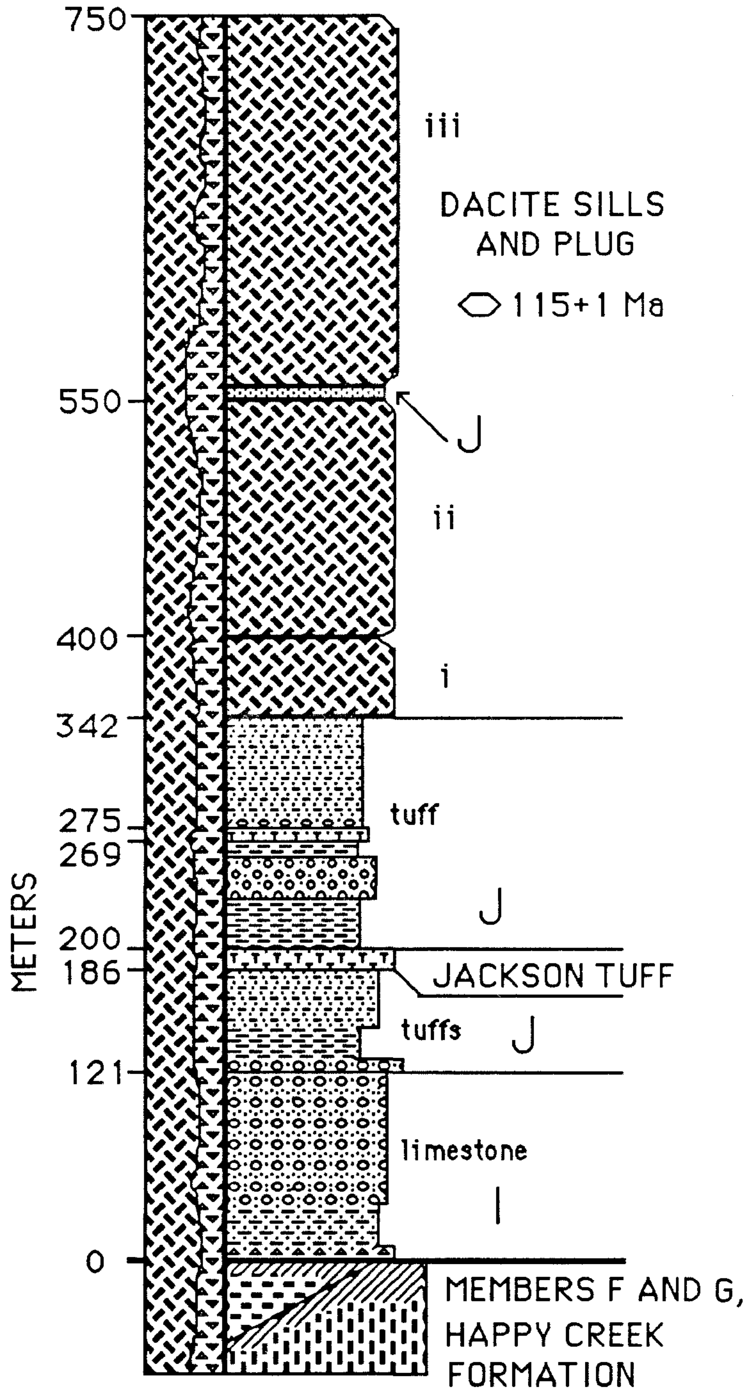
(10) Bliss Canyon

The area covers the outcrop of King Lear rocks in the headwaters of Bliss and Hobo Canyons. The section is generally homoclinal and east-dipping (with some thrusting at the very top of the sequence on the east), and includes both sedimentary and volcanic strata and related, synsedimentary volcanic hypabyssal intrusives in the form of a laccolith. The lower part of the King Lear was measured with a Jacobs staff along the divide between Bliss and Hobo Canyons. See fig. 43.

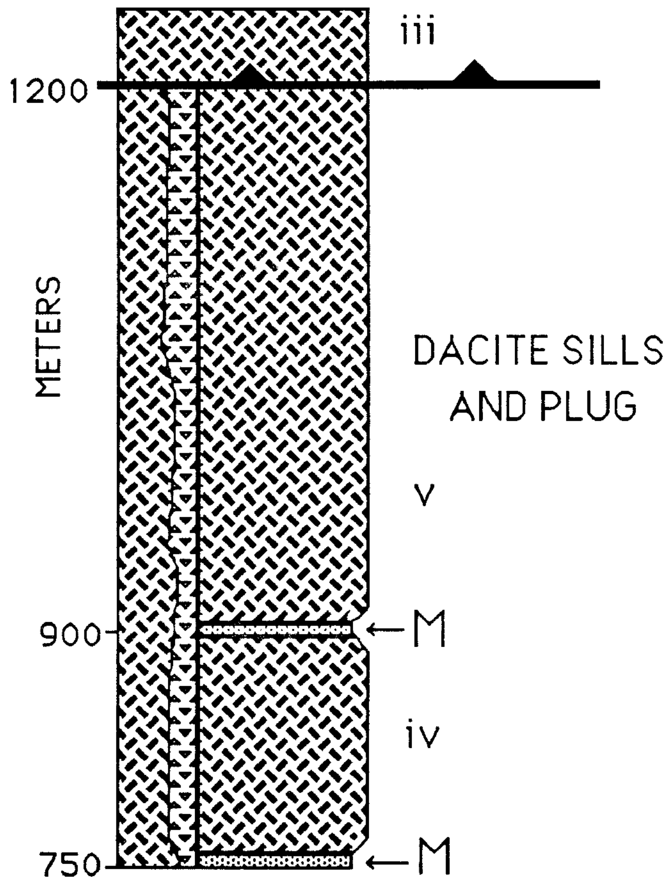
Stratigraphy Member I of the basal King Lear overlies and interfingers with member G (andesite) of the Happy Creek, or. where that unit pinches out, directly overlies member F (dacite and andesite). In the latter case, a deep subaerial weathering zone is developed on the dacite to a depth of tens of meters, with alteration (clay replacement of feldpsar especially) and reddening. Where member G is the substrate, the oxidation and alteration is shallower and less well-developed. One plagioclase phyric dark green volcanic flow of G a few meters thick is interbedded with member I of the King Lear, to the south of the section line.

The basal King Lear strata directly overlying the contact (a disconformity with a lacuna of unknown duration) is a 1 m bed of red volcanic breccia, and might actually represent a preserved soil zone. Two m of greenish grey shale comes next, then 2 m more of grey, brownish weathering, coarse-grained volcanic arenite, ripple cross-laminated and

FIG. 43
10 BLISS CANYON
STRATIGRAPHY OF THE KING LEAR FORMATION



10 BLISS CANYON (CONTINUED)



with carbonized wood remains. This arenite contains detrital chlorite, altered plagioclase, quartz, chert, biotite and muscovite in a calcitic replacive matrix. Over this is a 32 m thick sequence of grey shale and siltstone with thin interbeds of volcanic arenite, coarsening upwards to grey and green volcanic arenite and conglomerate, and grey shale. The volcanic arenite and conglomerate (with rare interbeds of shale) continue up to 121 m (all member I). In the upper 20 m of I two thin silicic tuffs are interstratified. This interval is very rich in large silicified and well-preserved wood chunks, locally 25% of the rock by volume. Whole logs, as well as leaf imprints, are present. The sediments are andesitic epiclastics, with volcanic rock fragments and detrital feldspar grains dominating. Textures are angular and poorly sorted. Large imbricated conglomerate lenses were not seen, and facies were laterally and vertically variable. At 72 m is a horizon of micritic grey limestone nodules, non-fossiliferous but arenaceous (with detrial plagioclase, volcanic rock fragments and quartz). A thin green siliceous aphanitic dike with amygdules also intrudes the section (probably related to the dacitic volcanic center).

At 121 m member J, made up of cherty/lithic clastic sediments, abruptly overlies member I. An 8 m thick chert-cobble conglomerate is the basal unit, with normal grading over the interval to a sandy top. The conglomerate has a coarse-grained and clean lithic arenite matrix, clast-support and imbrication, internal scouring and cross-bedding, an erosive and abrupt base and good sorting of the clasts. The clast population in outcrop includes 40% black chert, 50% green chert, 5% white vein quartz and 5% volcanic rocks (this latter component disappears upwards in the section), and the clast to matrix ratio is 2:1 with close packing. This conglomerate grades upwards to 2 m of coarse- to fine-grained green chertarenite, and then to 5 m of red shale, fissile and burrowed. The chertarenite is sorted with subrounded to subangular particles and has detrital chert, quartz (monocrystalline to mosaic to undulatory), metamorphic rock fragments (quartzose mica schist), and minor trachytic andesite rock fragments and detrital plagioclase, overly compacted with an illite cement. Overlying the red shale is a 2 m thick red lithic wacke, with green glassy volcanic fragments floating in a red siltstone matrix, then 11 m of green to red shale and another 6 m of the same lithic wacke. The wackes have matrix support of glass shards and minor plagioclase, quartz and Fe-oxides by red hematized silty clay. This wacke bed is rich in carbonate nodules, including septarian types. A thin tuffaceous sandstone, full of glass shards and 1 m in thickness, comes next, followed by 26 m of green siltstone and very fine-grained arenite, fissile and featureless other than the presence of more calcareous nodules. The tuffaceous sandstone contains poorly sorted but sub-rounded to rounded volcanic rock fragments and angular glass shards, quartz (corroded and embayed) and altered plagioclase with extensive micrite replacement of the rock. From 171 to 186 m is grey coarse-grained chertarenite, with angular grains and a rusty matrix.

At 186 m the Jackson Tuff overlies this interval of member J. The tuff is much thinner than to the north - it thins gradually from 100 to 70 to 14 m over less than 2 km. It also lacks the basal glass present less than 1 km to the north. Here, the base is made up of greenish welded ignimbrite and grades up to a white or very light grey ash tuff.

The tuff contains sanidine, quartz and plagioclase in a matrix devitrified to a very fine-grained quartz and feldspar mosaic. The ignimbiritic banding survives, with clear quartzose vs. cloudy brown laminae.

Member J continues above the tuff, starting with 31 m of green and grey shale, 27 m of coarse-grained wacke and pebbly arenite, and 8 m more of green shale. The wacke is grey and quartzose with a muddy matrix but grain support and has trough cross-bedding. It coarsens upwards to the pebbly arenite, which is cross-laminated. At 266 m is a 3 m bed of green quartzose fine-grained wacke with a muddy matrix, and then a tuffaceous wacke bed 6 m thick. This latter bed contains sand-sized glass-shards in a shale matrix. Next is a 6 m bed of chert-pebble conglomerate with a sandy matrix, grading rapidly upwards to cross-laminated green chertarenite, and then to 14 m of green sandy siltstone and shale. Another tuffaceous arenite, 2 m thick, overlies this. From 275 to 342 m the section is all green shale, siltstone and fine-grained arenite, recessive and homogenous. This interval is baked and reddish next to the dacite plug, which terminates outcrop of the section to the east and south.

The dominant feature of the King Lear Formation in this area is the dacite laccolith complex (member M), including a central stock and a number of sills spreading out from the plug at various levels within the sedimentary sequence. In plan it is a Christmas-tree laccolith (Corry, 1988). The stock is a highly elliptical cylinder, trending N20E, and 2.8 km long by 1.3 km at its widest. It intrudes the strata at a high angle. The sills are generally concordant to the surrounding strata, with flat bottoms and convex tops, and displace the sedimentary section upwards. On a scale of kilometers, however, the sills do cut several hundred meters down-section as they are followed to the south (in area 11).

Lithologically, the core of the plug is made up of green plagioclase + quartz ± biotite phyric dacite. The phenocrysts average several mm to a cm or more and the groundmass is very fine-grained to aphanitic. Xenoliths of darker green dacite porphyry or andesite are common. Some areas in the plug are bleached and altered. Aphanitic green dikelets cut the dacite and do not share the alteration and sericitization. Around the margins, the plug has a crumbly-weathering light-colored cataclastic banded nature, with a very strong foliation marked by color as well as by grain-size and abundance variations within a population of angular fragments in a siliceous groundmass (from sand-sized to 50 cm). The fragments have been highly brittly milled and sheared. This foliated marginal facies, though variable in attitude, is generally steep and subparallel to the edges of the plug, and on the order of 50 - 800 m in width. Small green aphanitic dikes cross-cut this fabric also. The contact zone with the country rock is chilled, aphanitic and almost glassy with a volumetrically less abundant population of small plagioclase or large rounded quartz phenocrysts.

The dacite sills can be mapped as five distinct bodies, characterized by the stratigraphic level intruded into, color, and phenocryst population. These sills (stratigraphically from lowest to highest) are: (i) green with plagioclase, quartz and biotite or hornblende phenocrysts, 50 m maximum thickness; (ii) green with plagioclase and lacking quartz or biotite, 150 m maximum thickness; (iii) green with plagioclase ±

quartz and biotite, locally foliated and with a bleached top, up to about 200 m in vertical extent; (iv) red with plagioclase and lacking quartz and biotite, a maximum of about 150 m; and (v) green with plagioclase and quartz and with a bleached top and a thickness of perhaps 300 m.

All the sills attain their greatest thickness due east of the plug, and have a lateral preserved extent of 6 km, thinning to both north and south. They either are traced into the plug or are cut by the marginal breccia envelope. In general, the sills contain up to 30% plagioclase, 15% quartz and 5% biotite in an aphanitic siliceous groundmass (plagioclase is ubiquitous, quartz common and biotite rare). The plagioclase appears as large, euhedral and zoned crystals, the quartz as large rounded blobs and the biotite as large plates. Thin section examination showed that some of the quartz (which is highly embayed and corroded) is actually tridymite though most is clear, monocrystalline ã-quartz. The plagioclase is andesine. Fe-oxides and brown biotite can be significant phases as well, and occur as cubic crystals and euhedral hexagonal plates, respectively. The biotite plates locally postdate and partially enclose plagioclase laths. The coarser-grained samples have an intersertal aphanitic matrix with aligned microlites of plagioclase, quartz, and Fe-oxide. Much of the groundmass in the glassier samples has also been devitrified to tridymite, in radial spherulites. Sill border facies are aphanitic and occur as xenoliths in the main bodies of the sills. The xenolith population also includes fine-grained densely plagioclase phyric felsite. Sill contacts can be distinguished by the marginal facies, or by screens of sedimentary facies (green siltstone and chert pebble conglomerate from member J or arkose, lahar, green siltstone and tuff from M - the latter containing detrital clasts identical in lithology to the sills). The upper parts of several of the sills are bleached and altered extensively. This also occurs at some places in the country rock. Sills at their distal fringes may finger out to several apophyses. A well-developed foliation identical to that in the marginal facies of the plug is present in several sills, with zones of very well foliated, milled and sheared dacitic breccia subparallel (though rather variable) to the sill margins. These breccias are composed of dacite fragments from within the sill. In thin section these foliated igneous breccias have a cataclastic texture, with angular fragments of dacitic sill lithologies, andesine, and quartz (plus minor apatite and chlorite after biotite) in a volcanic siliceous groundmass with patches of epidote replacement and chlorite growth. Microdikes of grey, or more commonly, pink glass crosscut these foliated areas in particular. These dikes are at a high angle to the banding and only a few mm thick. Other, thicker, quartz- and plagioclase- bearing grey fine-grained dikes also crosscut the sills.

One sill (i) also has distinctive flaser structures, where narrow zones a few cms in width have been ductilely sheared. These east-dipping shear zones are characterized by wispy, strung-out and ribboned quartz phenocrysts, grading out to aligned and then random fabrics. The plagioclase grains are broken and milled, but not ribboned. Within the shear zones are epidotized and slickensided surfaces. A down-dip lineation is also developed with the shear foliation.

Thin sections of the shear zones show the cataclastic texture. The quartz is strained, undulatory and elongated or even ribboned to broken and fragmented, and the andesine is deformed only in a brittle fashion. These components float in a weakly to well foliated cataclastic matrix of ground-up quartz, plagioclase and dacite fragments; even the undeformed phenocrysts are subparallel to the foliation. These shear zones, dipping east 35°, are inferred to be related to thrusting along parallel surfaces.

Age One of the sills (iii) has been dated by U/Pb geochronology on extracted zircons at 115 $\pm$ 1 m.y. The intruded King Lear sediments (members I through lower M) must therefore be older than this. M is inferred to be near in age to sill or slightly older, as the volcanic source active during deposition of the unit is represented by the laccolith complex of which the sill is an integral late stage part. Thrusting, which took place during or very soon after the emplacement of the laccolith and epiclastic sedimentation (as witnessed by both the subsolidus flaser zones in the lowest sill and by the soft-sediment compressive deformation of the volcanic strata of upper M), was therefore also active at this time (the Aptian).

Depositional Environment A significant lacuna reflecting subaerial exposure exists between member I (the lowest King Lear) and the Happy Creek where the highest exposures are from member F (dacite). Where I overlies member G of the Happy Creek the disconformable nature of the contact is much less evident, implying a shorter time period between those two units. In fact, at one site here G and I are stratigraphically intercalated.

Member I was deposited in a subaerial alluvial setting. The overall green instead of red coloration might have been due to the proximity of the intrusive complex, which could have caused reduction of the iron in the neighboring sediments. The abundant silicified wood, the stratigraphic architecture (highly variable, which, following Walther's Law, implies juxtaposed and swiftly changing depositional environments) and depositional textures (poorly sorted and subangular) lead to the conclusion that this was a braided alluvial fan. The shaly horizons are taken to represent sheet flood events, while the coarser-grained facies were laid down within the braided stream systems. A more distal setting within the alluvial fan is indicated by the lack of larger clasts and mudflows. Siliceous volcanic eruptions during this time period in the area are documented by the ashfall tuffs in the section (which are rather characteristic of member I).

Member J overlies I with an abrupt and erosive lower contact, implying a sudden change in provenance. Quite active continuing volcanism is again indicated by the numerous tuffs, tuffaceous sandstones and tuffaceous wackes in the section (including the Jackson Tuff). The basal chert pebble conglomerate lens is laterally continuous over at least several hundred meters, and has all the characteristics of a large active meandering fluvial channel (clast imbrication and support, fining upwards, good sorting of clasts with a clean sandy matrix, internal channel scouring and cross-bedding - Selley, 1985). The coarser-grained intervals higher in the section, though not diagnostic, share some of these characteristics. From 275 to 279 m there is a small fluvial channel with basal lag and rippled arenite. At 231 to 258 m the cross-bedded arenite and wacke represents point bar environments within a larger fluvial channel. The source for the cherty/lithic clastics in member J was a regionally metamorphosed terrane, as seen in the strained and partially recrystallized fabrics in the quartz grains (Young, 1976) and by the inclusion of schistose debris. The homogeneous shaly and silty to fine-grained chertarenitic lithologies, which dominate the section over the basal conglomerate, are foodplain facies, laid down during regionally extensive flood events. The tuffaceous arenites represent fluvially or sheetflood-reworked pyroclastics, while the tuffaceous wackes are mudflows or even lahars caused by and involving similar pyroclastic deposits. Small ephemeral flood-basin lakes are also indicated by the horizons of limy nodules (heavily altered by diagenetic redistribution of the calcite; depositional textures are preserved in some while others are entirely diagenetic).

The Jackson Tuff in this area only exhibits the upper two of the three lithologies seen in it elsewhere (these are basal glass, middle welded ignimbrite, and upper crystal-vitric tuff, with gradational contacts). The ignimbritic nature is indicated by the laminated green aphantic lithology of the base of the bed. The crystal vitric tuff contains the same crystal population seen everywhere in this bed-quartz, sanidine and andesine plagioclase. The somewhat further evolved character of the devitrification is probably due to the adjacent hypabyssal volcanic center.

The laccolith is interpreted as such because of its geometry and intrusive relationships (Corry, 1988). It consists of the central zoned stock and at least five major sills (some of these could be composite). The outer cataclastic marginal facies in the plug could be either explosive or peperitic (though evidence for peperitic involvement of the surrounding country rock is lacking), or alternatively could have been formed by flow within this feeder zone. In the latter case, this (chaotic?) flow would entrain, shear and fracture the brittle shell of the plug, resulting in a foliated cataclastic breccia subparallel to the margins. The central core facies of the plug would have been more fluid or a later intrusion. It is the preserved conduit to the sills and the now-vanished volcanic center; as such it may postdate (and is inferred to truncate) the sills. The alternative explanation for the cataclastic zones are that these igneous breccias are explosive in origin, caused by major phreatomagmatic events; however, the fabrics occur in the sills as well. A point in favor of this hypothesis is that there are no ductile fabrics observed within the breccia zones, as might be expected of aphanitic rhyolite xenoliths reinvolved with flowing magma. In addition, I judge the breccia envelope to be too thick and the brecciation too well-developed and consistent to have formed by simple flow-induced cataclasis. The best conclusion is that this fabric is indeed explosive in genesis, and may have been caused by one or more phreatomagmatic events.

The sills are highly tabular and generally concordant, and, as they displace strata upwards by inflation are inferred to have convex upper surfaces. In detail, the fact that the sills cut down section southwards implies (if each one intruded along a surface where the density of magma and of surrounding rock was the same - Corry, 1988) that the section was greatly thinner to the south. This thickness variation is inferred to be caused by relief on the top of member G of the Happy Creek Formation; the areas of thinner King Lear sections were deposited on top of the topographic highs within the Happy Creek, and thicker sections in the valleys. These topographic highs are inherited volcanic constructional forms. The laccolith intruded into the thickest part of the King Lear, flanked to the north and south by basement highs. Each sill is thought to represent a distinct episode, as they do vary in color (controlled by the oxidation state of Fe^{++} or Fe^{+++} in the groundmass) and phenocryst content. Similar foliated cataclastic zones within them are also thought to be caused by phreatomagmatic activity involving groundwater, though this is not certain.

The flaser fabrics found as discrete parallel surfaces within one of these sills are very distinct from the thick zones of xenolithic foliated flow breccias discussed above. Ductile deformation of the quartz and of the matrix, but not of the plagioclase crystals (which were brittly deformed) indicate shear at rather high but not magmatic temperatures. This subsolidus (but not protoclastic) deformation took place relatively soon after the intrusion of the sill. As these shear zones are parallel to the thrusts to the east, they are inferred to be genetically related.

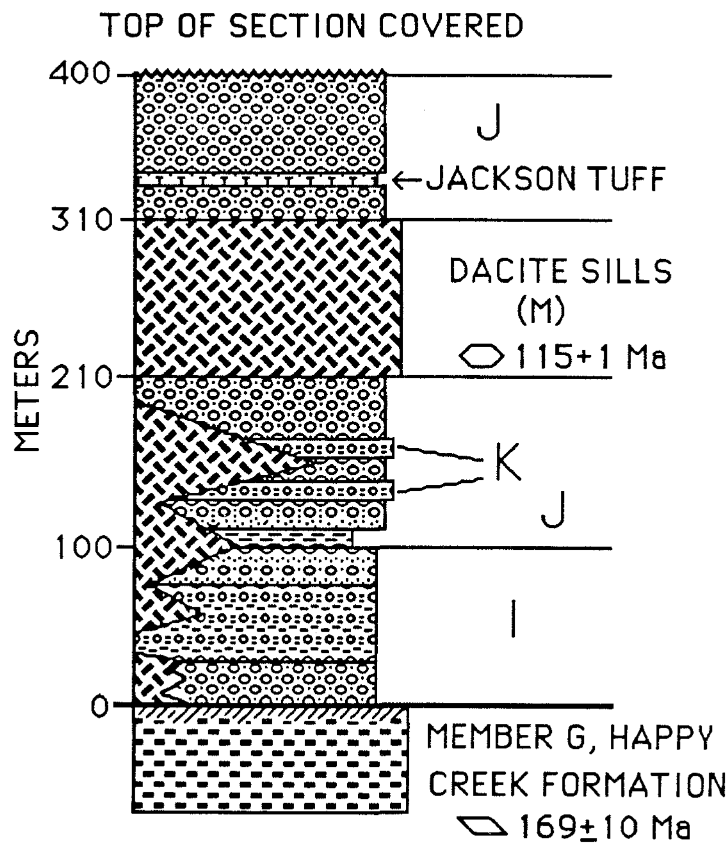
The orientation of the long axis of dacite plug, of the feeder dikes for the Jackson Tuff (plus its locations of maximum thickness), and of the dikes related to the Buff Peak rhyolite domes are all the same. These and the Jackson Creek rhyolite dome also all lie along a trend parallel to that orientation (N20E). The implication to be garnered is that there is some fundamental deep-seated geologic feature controlling the location and orientation of all these silicic volcanic occurrences, or that the stress-field controlled their location. In the latter case, the stresses are not compatible with kinematic analysis - see chapter 8. This lineament could a buried fault or fracture system within the crust, running up the center of the range at N20E, and of uncertain origin. The plutons and minor dike swarms within the Happy Creek north of the mouth of Jackson Creek and in the Trout Creek Spur are also parallel to this orientation.

(11) Clover Creek

This area, northeast of King Lear Peak, is within the drainages of Clover and Louse Creeks. The section dips homoclinally to the east. It is bounded on the west by the basal unconformity (near the range divide), on the north by the dacitic hypabyssal volcanic center (area 10), and on the south by a high-angle fault. To the east, Tertiary basalt flows and Quaternary alluvium overlap it. This region is part of the unthrust-faulted central autochthonous buttress of the range. See fig. 44.

Stratigraphy The basal contact with member G of the Happy Creek is not well exposed in this area; it appears to be a conformable or disconformable contact, but often has sills injected along it, obscuring relationships. The base of the King Lear is made up of red to grey volcanic pebbley arenite. The basal beds are overlain by red to green volcanic conglomerate. The subangular coarser fractions of the conglomerate float in a mud-matrix, and the rock is very poorly sorted and massively bedded. These diamictite beds lack fragments of coarse-

FIG. 44
11 CLOVER CREEK
STRATIGRAPHY OF THE KING LEAR FORMATION



grained plagioclase phyric dacite such as occur in the laccolith complex and related facies, but do contain Happy Creek lithologies. Further up in the section is fine-grained volcanic arenite to volcanic conglomerate, still subangular, moderately sorted, with normal grading and cross-bedding. This whole interval is part of member I and is on the order of 100 m thick.

Green siltstone and chert-pebble conglomerates are found stratigraphically above this epiclastic interval; the latter are generally imbricated, in clast-support, cross-stratified and well-bedded. These are placed in member J, which is extensively intruded by dacite sills. True thickness can not be estimated reliably, but is in the order of several hundred meters. A large dip-slope of chert-pebble conglomerate at the top of the member (formerly called the Pansy Lee Conglomerate by Willden, 1963) above Louse Creek contains a clast population of grey quartzite (65%), black chert (30%), green chert, argillite and vein quartz (totalling 5%). The clasts are rounded (up to 15 cm in size) to subangular (the smaller size fractions, presumably from broken and fragmented larger cobbles) and imbricated with some ferruginous cement in the depositional voids. Measurements on the imbrication fabric were taken for paleocurrent analysis (and are discussed later - see fig. 46). The matrix here is a very coarse-grained lithic arenite with no fines, and is cross-bedded. The clasts are well sorted and the lithology is highly bimodal. Silicified wood and wood imprints are locally abundant.

In the upper part of this succession the Jackson Tuff crops out. The tuff is a few meters thick and composed of white rhyolitic ash-fall tuff with quartz grains. Large planar intraclasts of the tuff are found reworked in the immediately overlying chert-pebble conglomerates, and the superjacent arenite beds also have a minor tuffaceous component, and contain reworked slabs of the tuff. In thin section, the tuffaceous arenite above the tuff contains subequal amounts of brown siliceous reworked glass grains, metamorphic rock fragments (phyllitic - dark brown, foliated and with fine-grained white mica), quartz and chert grains, and has a hematitic cement. No andesitic or plutonic (Happy Creek) or dacitic (member M, King Lear) debris was present; provenance was solely from the McGill Canyon and Bliss Canyon Formations.

Several thick beds of volcanic diamictite (andesitic volcanic clasts floating in a muddy matrix) are intercalated with member J near the bottom, and are placed in member K. They also lack detritus identifiable as having a source in the dacitic volcanism of M. These beds are siliceous and dark green, poorly sorted, and contain accretionary pisolites.

Numerous sills (and several small pipes) of member M intrude the strata of this area, spreading out from the dacitic laccolith complex to the north in area 10. These sills pinch out to the south, and can be as much as 150 m thick in this area. In one area on the divide in the headwaters of Clover Creek, the intrusive relationships between these hypabyssals and the base of the sedimentary facies of the King Lear (member I) are well exposed. Here, light grey pebbley volcanic arenite is intruded by pillows of greenish-grey aphyric dacite with peperitic relationships. The sediment is very contorted and ductilely disturbed and contains entrained angular clasts of plagioclase phyric dacite. The sedimentary rock is locally baked where caught up between the pillows. The pillows have vesicles in the outer rind, and the edges of these features are also altered where in contact with the sediment. The pillows are up to 2 m long and irregular to elliptical in shape, and contain xenoliths of the country rock as well.

In several other areas of sill emplacement, the intrusives are rather variable in lithology. The dominant rock type is coarse-grained green to red plagioclase ± quartz ± biotite phyric dacite with an aphanitic to very fine-grained groundmass. The plagioclase is present as large laths, the quartz as amorphous elliptical blebs and the biotite as discrete plates (all several mm up to 1 cm in size). The intrusive facies are bleached and altered (the plagioclase sericitized and the groundmass silicified), particularly near the margins. The marginal phases (dark green) are also substantially finer grained to aphyric, and can be present as angular xenoliths in the interior phases. Pink glassy and quartzitic veins crosscut the sills in some areas, and breccia zones occur (as in section 10). Epidote and magnetite seams, and siliceous flow banding are also present. The breccias can grade into the surrounding sill facies. The tops of the sills are characterized by finer grain size, extensive bleaching and silicification, red to pink to grey coloration, cross-cutting veins and brecciation. The various sills are also distinguishable and were mapped by the proportion of the described facies they contain, and by their intrusive contacts with one another and with the sediments of the King Lear. The sills do not typically intrude the Happy Creek strata immediately underneath the King Lear but are confined to levels at and above the unconformity, and also are not seen south of the high angle fault bounding the south edge of the area.

It is worth noting that in one area some of these sills (intruding member G of the Happy Creek) are folded by the younger, east-vergent western thrust system, and so must in part predate that orogenic episode.

Age One of the sills from the dacite laccolith complex to the north was dated at 115 $\pm$ 1 m.y. by U/Pb zircon geochronology; members I, J, and K and the Jackson Tuff all must be older than that. Member G of the Happy Creek on King Lear Peak just to the south was dated by Rb/Sr geochronology as 169 $\pm$ 10 m.y. (2 Ú) by Russell (1981). The interval of the King Lear exposed here is thus between Bathonian and Aptian in age.

Depositional Environment Member I is made up both of alluvial stream deposits and debris flows. The thick poorly sorted mud matrix-supported diamictites are proximal debris flows (but not lahars - no evidence for volcanism concurrent with these deposits was found, and the epiclastic volcanic component is from the older strata of the Happy Creek). The type of alluvial environment is not certain with the degree of exposure present, but the normal grading, clast-imbrication and cross-bedding with clast-support does point to some type of fluvial setting. Proximity to high relief is indicated in particular by the debris-flows, which vary substantially in thickness through the area (thinning northwards).

Member J was deposited in a meandering distal alluvial fan to flood plain, and reflects the abrupt change in provenance linked to tectonic evolution of the sourceland. The thick and laterally continuous conglomerate beds with internal large-scale cross-bedding and pebble imbrication and with significant interbeds of green siltstone are clearly meandering (not braided) fluvial channel and overbank facies; such an environment is supported by the plant fossils associated with these facies.

The Jackson Tuff is not as diagnostic as in other areas, but is a rhyolitic crystal-rich vitric tuff, and is inferred to be rather more distal (because of the thinness of the bed and lack of ignimbritic textures). The diamictite beds attributed to K and interbedded with J are debris flows, but the presence of accretionary pisolites (which form due to rainfall in the ash cloud generated by phreatomagmatic columns- Cas and Wright, 1987) strongly indicates that these diamictites were volcanically induced and should be considered to be lahars though lacking in an intraformational volcanic component. Similar pisolites are associated with the rhyolitic volcanism of members M and L.

The dacitic sills appear to have intruded the sediments of the King Lear while the latter were still unconsolidated and wet, judging by the peperitic textures witnessed (the emplacement of pillows of dacite in the ductilely deformed sediment, which also contains angular clasts of the dacite that are not present in the normal sedimentary assemblage for that unit). The rheology of the strata (a thick sequence of structurally weak and well-bedded King Lear sediments within a basin of significant relief and on top of very strong and massive Happy Creek volcanics) limited the sites of sill injection to within the King Lear part of the sequence. The sills also disappear to the south away from the plug coring the laccolith. Hydrothermal activity related to the sill intrusion is indicated by the bleaching, silicification and alteration seen within the sills and affecting the intruded sediments near the contacts.

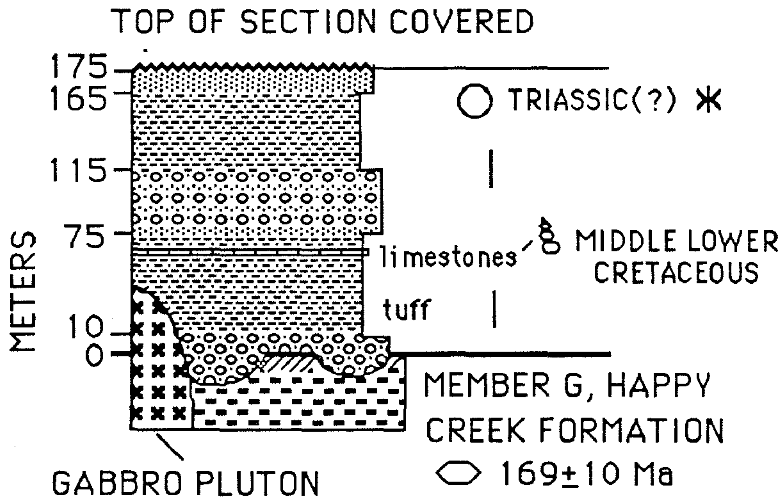
(12) King Lear Peak

This area lies on the slopes east and south of King Lear Peak, where the western east-vergent thrust system has placed Happy Creek over King Lear and tightly folded the King Lear section. It is the southern boundary of the thesis mapping area. A large high-angle fault (reactivated as a sinistral tear fault during thrusting) separates this area from that to the north. The strata are not affected by dacitic hypabyssal volcanism, but the base of the formation in the extreme south of this region is intruded and contact metamorphosed by a gabbroic to monzonitic pluton. Tertiary basalts overlie the Mesozoic section with angular unconformity. See fig. 45.

<u>Stratigraphy</u> Only member I, greatly increased in thickness, is found in this area; higher stratigraphic levels are not present. The basal unconformity is well-exposed, with member G of the Happy Creek weathered and oxidized. The normally green fine-grained plagioclase phyric andesites are reddened and degraded, with red lisagon staining and grey carbonate seams dominating as the unconformity is approached. The King Lear and Happy Creek are parallel but with evidence of a hiatus, so the relationship is actually disconformable. Thicknesses of strata were paced out and exposure was quite good.

Red boulder and cobble conglomerate channel-fill deposits occupy channel cut into the Happy Creek at the base of member I. Clast

FIG. 45
12 KING LEAR PEAK
STRATIGRAPHY OF THE KING LEAR FORMATION



lithologies include the ubiquitous pink granite, and a variety of andesitic lithologies from the Happy Creek. The large clasts are rounded, while smaller ones are subrounded to angular. The matrix is a volcanic arenite and is cross-laminated in the finer-grained lenses. The beds are massive but overall normally graded and are in clast-support. Several of these channels exist with erosional relief of 10 to 15 meters. Elsewhere the basal red volcanic conglomerate is missing, to only several meters thick. Sedimentary breccia lenses of magnetite gravel are also present.

Overlying this basal conglomerate is 65 m of green shale and siltstone, with light grey limestone nodule horizons and beds up to 1 m thick. The nodules are irregularly elliptical in shape and 30 to 50 cm in size, and locally amalgamate to form mottled beds with highly irregular margins; both are sandy. These occurrences are concentrated at the bottom of the shaly interval close to the contact with the Happy Creek, and also in at least three horizons near the top. The shale and siltstone are homogenous, with no outstanding characteristics. There was at one location with the lower part of this interval a greenish-brown felsic crystal-rich tuff, with an aphanitic siliceous matrix; the grains are highly angular and the bed more than 1 m thick.

On top of this is another 40 m of red cross-laminated volcanic arenite and volcanic conglomerate, characterized by subrounded detritial particles, clast-support in the coarser facies, thick bedding and poor sorting. Interstratified with this lithology are lenticular and laterally impersistent thick conglomeratic beds with mud matrix-support and reverse grading at the base.

Next is 50 m more of green shale, siltstone, and fine arkosic wackes with shale intraclast rip-ups. These lithologies are still rather featureless. Overlying this is a 1 m thick bed of cross-laminated medium-grained, well-sorted and rounded light grey volcanic arkose. This bed is the base of a sequence containing more red beds similar to those lower in the section. The bed is at least a few meters in thickness; the top is not seen.

The large scale architecture in this member is quite laterally persistent, with only minor thickness variations (except in the basal cobble and boulder-conglomerate). Nowhere in this entire succession is a major component of cherty/lithic debris involved (such as would characterize member J), though a chert pebble in a conglomerate of uncertain stratigraphic position was found and dated (D.L. Jones in Russell, 1981).

In the Bull Creek area just east of Navajo Peak, a large zoned pluton (gabbro core to monzonitic shell) of the Early Mesozoic Intrusive Suite intrudes the strata described below. The sedimentary rocks are overprinted. A contact-metamorphosed volcanic conglomerate in thin section exhibited epidote, chlorite and calcite overprinting the sedimentary textures, with incomplete recrystallization of the detrital quartz and plagioclase grains. The latter now have irregular serrated boundaries and a mosaic pattern. The metamorphic grade thus appears to be albite-epidote hornfels. Some mining took place in this contact metamorphic zone, and Cu-mineralization (azurite and malachite) was evident in the spoil piles.

Age The monzontic phase of the pluton is a likely rock type to yield zircons (not sampled in this study), and U/Pb geochronology on this unit might well give a minimum age on the base of the King Lear and member I, which is greatly needed. J.B. Reeside reported on a collection of fossil gastropods from one of the limestone horizons in member I (Willden, 1963) as probably being of the middle Lower Cretaceous, and of a fresh-water origin. Russell (1981) also reported Early Mesozoic (probably Triassic) radiolaria in a chert pebble.

Depositional Environment The basal unconformity, with the pronounced subaerial weathering, caliche formation, and the incised channels, indicates a hiatus in deposition in this area between the Happy Creek (member G) and member I of the King Lear. This may be connected to the unusual thinness of member G in this area; the area is on the fringe of stratovolcano construct in G.

These are probably some of the best exposures of member I in the range. Some of the outstanding characteristics within the coarser-grained epiclastic red beds are: lateral continuity of bedding (except for the diamictites), moderate sorting and rounding, pebble imbrication and clast-support, strong bimodal size distribution, normal grading at the top of thick beds, ripple cross-lamination within the arenites and trough cross-bedding in the conglomerates, large-scale erosive channelling (at the disconformity on the bedrock), general lack of fines, and the red to green color and lack of organics. All of these features clearly indicate a meandering fluvial channel facies in a semi-arid environment. The volumetrically important and continuous finer-grained facies represent an overbank flood-plain setting (perhaps within low-lying, long-lived flood basins), with scattered lakes indicated by the limestone horizons. The extensive calcrete deposits within this interval and the low sedimentation rates implied by their development indicate that deposition may have taken place over rather long periods of time (10,000 years per bed - Reineck and Singh, 1980). The overall depositional environment of these facies was an alluvial flood plain (or a distal alluvial fan). The diamictites are interpreted as lobes of debris-flow deposits, which is slightly unusual in this environment and implies some relief in the vicinity (as well as supporting a semi-arid setting). Debris flows can travel significant distances but generally need some slope and hydraulic head to propagate (Collinson, 1987).

Even with the top missing, member I here is quite substantially thicker than just to the north in area 11, across a high-angle fault. Although this fault was reactivated as a sinistral tear during the late phase of east-vergent thrusting, it is inferred to be a reactivated earlier structure because of the stratigraphic changes across it in member I. This structure could have provided short-lived and local fault scarp relief for the generation of the mudflows. The thickening also seems too much to be accounted for by simple facies changes, given the relatively flay-lying alluvial floodplain setting. In its initial phase, I interpret it as influencing sedimentation as a growth fault with south-side down, and as part of the generation of these features active during members H (Happy Creek), I and lower J (King Lear). Member G of the Happy Creek does thin rapidly in this area, however, and inherited topographic relief also might have caused the thickening of member I from sections 11 to 12, rather than activity across that high-angle fault.

Paleocurrent Data

Pebble imbrications were measured in four different areas within the King Lear Formation (see fig. 46). The measurements were taken on the dip direction of the plane defined by the ellipsoidal pebbles. As pebbles are imbricated so that this direction should point upstream, a large enough collection of these measurements should give statistically valid paleocurrent information. Specifically, the imbrication measurements will point upcurrent and towards the source area. As the chert and quartzite pebbles in member J tend to the flattened ellipsoid much more than the relatively rounded volcanic epiclastic pebbles and cobbles of member I, the measurements for J will be more reliable. As all of these sites are of low-sinuosity meandering fluvial origin, a single site can give a direction substantially away from the sourceland because of stream sinuosity, but a set of several sites should give a better indication.

The four sites are from upper member J of area 3, lower member I and lower J of area 5, and upper member J of area 11. The extent of agreement among the four sites is rather reassuring that the measurements are telling us something real. Area 3, member J gives 82° ± 36° (the uncertainity is 2Ú, and n is the number of measurements, = 20); area 5, member J yields 151° ± 16° (n = 16); area 11, member J gives 91° ± 20° (n = 16); and area 5, member I yields 138° ± 17° (n = 23). The sourceland for the cherty/lithic sediments deposited in member J and even for member I (in this particular area) was therefore to the south and east; this is

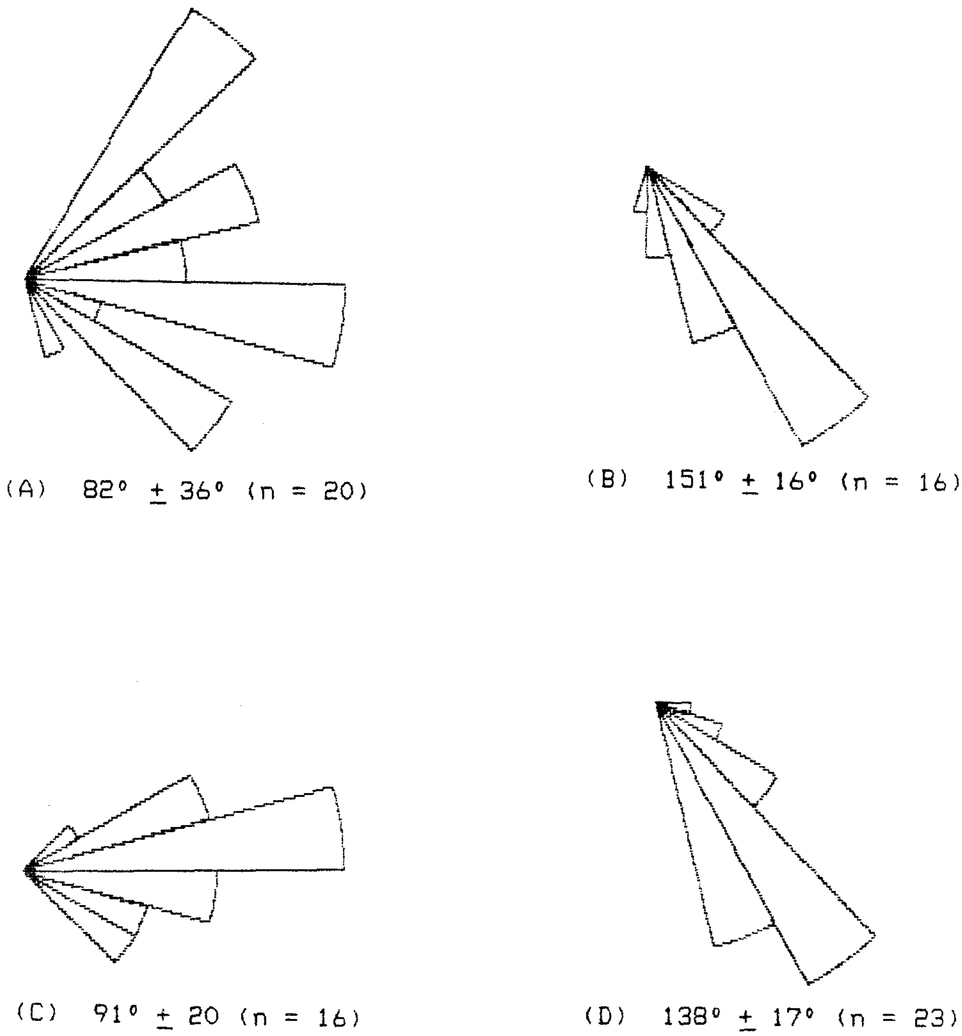


FIG. 46

Pebble imbrication measurements from fluvial conglomerate facies of the King Lear Formation, Jackson Mountains, NW Nevada. The dip azimuth of the pebbles indicates the upcurrent direction during deposition, and the position of the sourceland. The paleocurrent data comes from: (A) area 3, upper member J; (B) area 5, lower member J; (C) area 11, upper member J; and (D) area 5, member I.

(not coincidentally) the direction the first phase of thrusting and the resulting orogenic highlands would have been propagating from. These thrusts trend about N20E (therefore verging to the north and west), and the thrust front would have shifted closer and closer to the field area until the thrusts impinged on and deformed it. The paleocurrent data is a further indicator that the west-vergent thrusts were active during and provided the sourceland for member J, K and N; and perhaps I as well (particularly outside the pre-existing to synsedimentary fault-bounded basins).

Depositional, Volcanic and Structural Framework and Evolution of the King Lear Formation

The facies and structural architecture of the King Lear Formation, as discussed above, is extremely complex. A number of overlapping factors have influenced depositional patterns within the unit, including inherited topography, high-angle faults with a major dip-slip component, ongoing andesitic to rhyolitic volcansim in several distinct centers, and two distinct phases of thrusting. Because of the good preservation and extent of outcrop, the relationships documenting and clarifying the roles of these various phases and factors, though rarely obvious, can be discerned in some detail.

The basal contact of the King Lear is highly variable. At the end of deposition of the Happy Creek Formation, the top of member G defined several distinct topographic highs formed by volcanic constructional processes, with angles inferred to be in the range of 10° to 15° and relief of 800 m or more. The King Lear was deposited in and over this

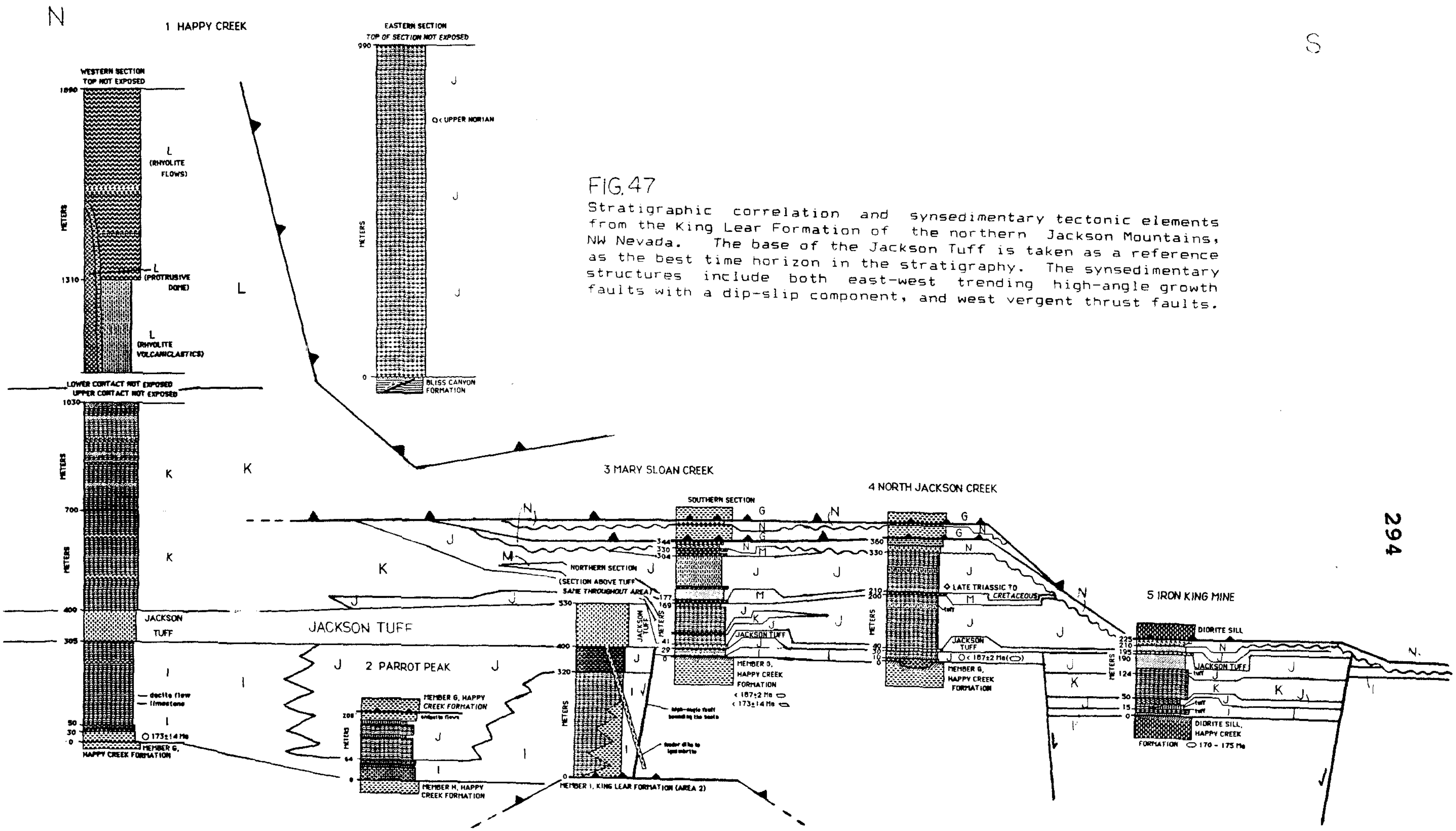


FIG.47

Stratigraphic correlation and synsedimentary tectonic elements from the King Lear Formation of the northern Jackson Mountains, NW Nevada. The base of the Jackson Tuff is taken as a reference as the best time horizon in the stratigraphy. The synsedimentary structures include both east-west trending high-angle growth faults with a dip-slip component, and west vergent thrust faults.

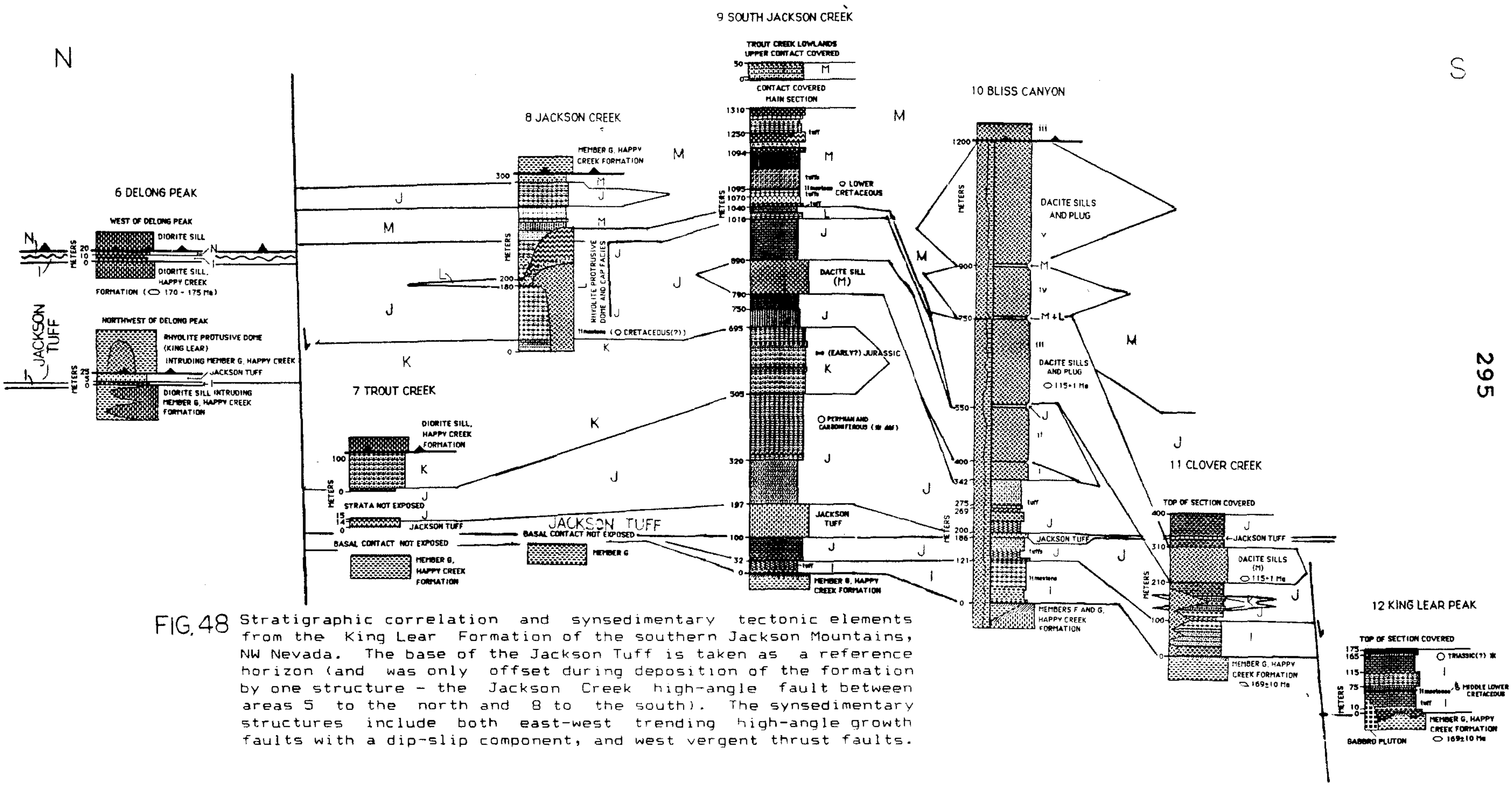
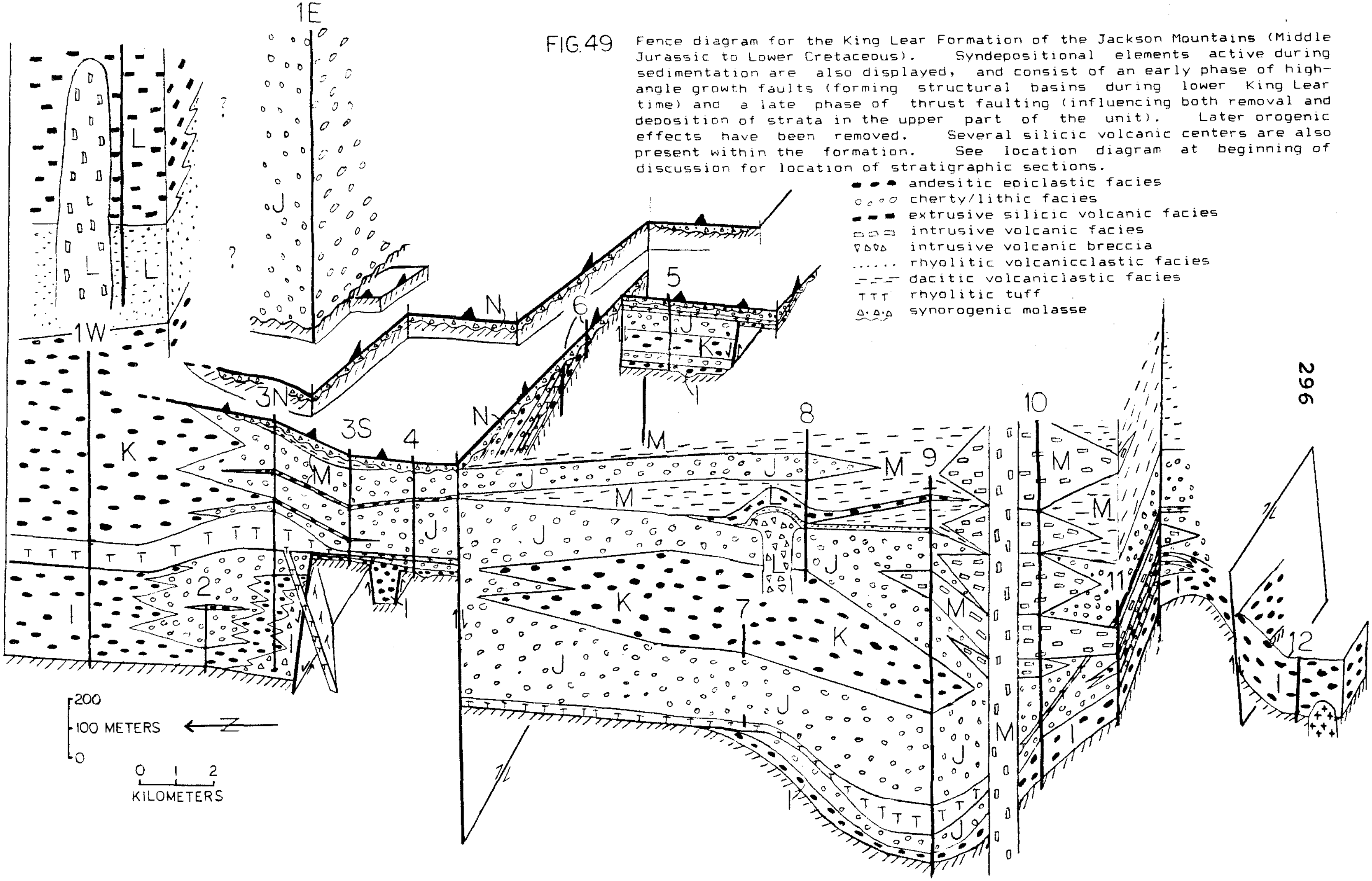


FIG. 48 Stratigraphic correlation and synsedimentary tectonic elements from the King Lear Formation of the southern Jackson Mountains, NW Nevada. The base of the Jackson Tuff is taken as a reference horizon (and was only offset during deposition of the formation by one structure - the Jackson Creek high-angle fault between areas 5 to the north and 8 to the south). The synsedimentary structures include both east-west trending high-angle growth faults with a dip-slip component, and west vergent thrust faults.

FIG.49 Fence diagram for the King Lear Formation of the Jackson Mountains (Middle Jurassic to Lower Cretaceous). Syndepositional elements active during sedimentation are also displayed, and consist of an early phase of high-angle growth faults (forming structural basins during lower King Lear time) and a late phase of thrust faulting (influencing both removal and deposition of strata in the upper part of the unit). Later orogenic effects have been removed. Several silicic volcanic centers are also present within the formation. See location diagram at beginning of discussion for location of stratigraphic sections.



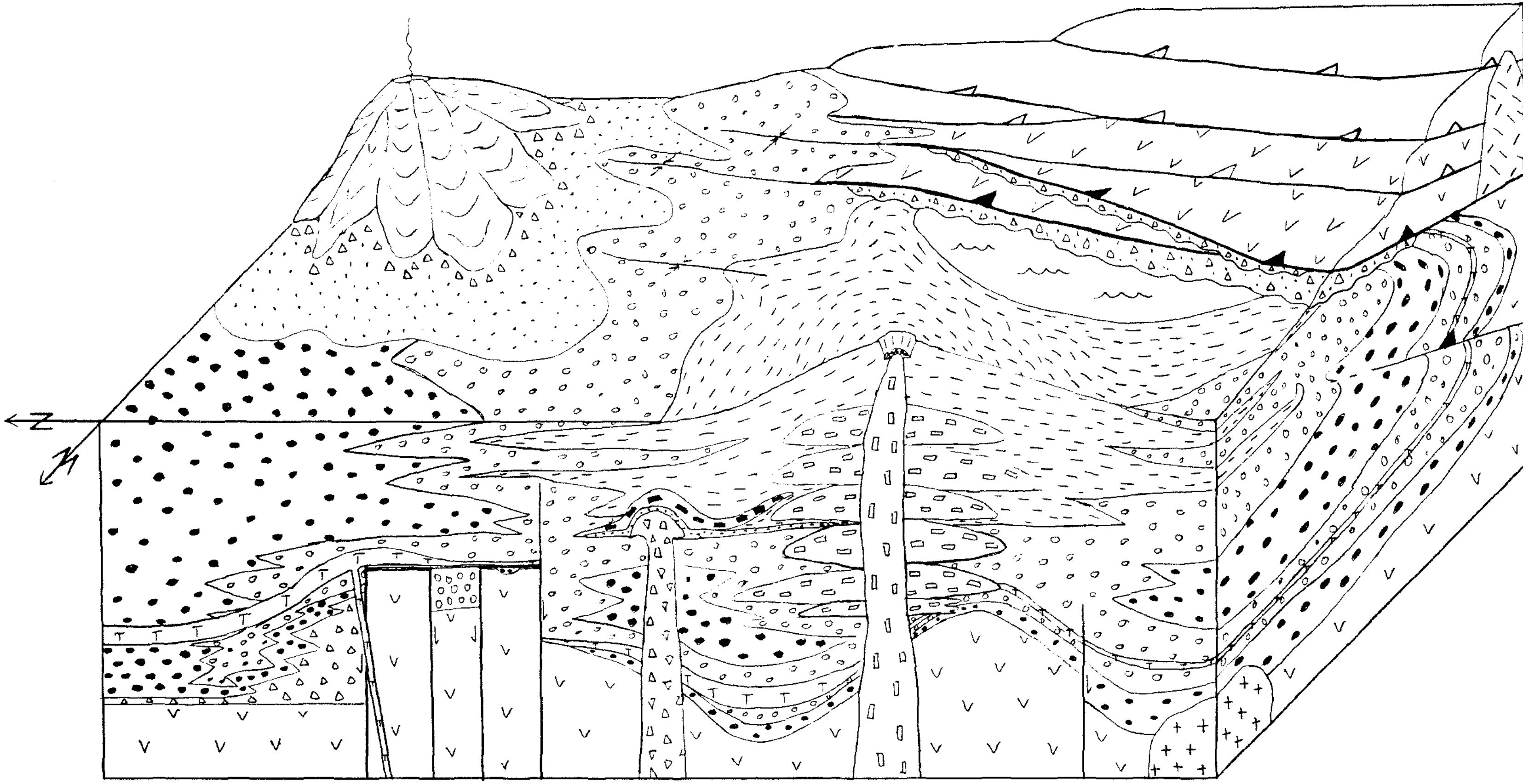


FIG. 50 Block diagram displaying the depositional, volcanic and structural elements of the King Lear Formation of the Jackson Mountains, NW Nevada during the Early Cretaceous. Facies symbols same as for fence diagram.
△ inactive thrust ▲ reactivated thrust

basement relief, and in a number of places there are dacitic to andesitic flows interbedded with and dikes intruding member I (the base of the formation). This indicates a locally gradational transition from volcanism-dominated to molasse-dominated depositional regimes. Elsewhere, a substantial hiatus is indicated by the disconformable nature of the contact. Deep subaerial oxidation and weathering profiles, including lisagon staining and caliche veins and lenses, characterize the contact in these areas (particularly where member I overlies older strata of the Happy Creek, such as member F). This type of weathering is characteristically semiarid. Within the eastern, west-vergent thrust belt the King Lear locally overlies the Happy Creek (and even the Bliss Canyon) with angular unconformity, and overlaps some of the thrust faults. These relationships are from younger syn- and post-thrusting exposures of the King Lear.

In other places, active tectonic elements (initiated during members G and H of the Happy Creek) formed scarps and growth faults that juxtaposed the King Lear and Happy Creek Formations structurally, or caused the King Lear to overlap the Happy Creek along a steep buttress unconformity.

Member I of the King Lear is quite thin to missing on the fault-bounded highs resulting from this activity. These horsts might have 10 m or less of the unit, except where fluvial channels have been incised into the Happy Creek basement and subsequently filled with bouldery deposits. The corresponding basins within the grabens might have up to several hundred meters of the unit. Member I is characterized primarily by provenance - it consists entirely of debris traceable to the Happy Creek Formation and Early Mesozoic Intrusive Suite and made up of andesitic, dioritic and monzonitic debris, which had already undergone magnetite and epidote replacement prior to erosion and deposition. The unit is almost always red (except in several areas where later nearby intrusions have induced reduction and it is now green). On the horsts, member I is composed of a thin gravelly screen over the basement, made up of conglomerates with a volcanic arenite matrix and sometimes missing altogether. This facies was deposited by braided streams and sheetfloods across a pediment surface floored by the Happy Creek. In the grabens, breccias lie against the bounding faults where exposed, and are slightly sheared by them. These talus to upper alluvial fan breccias (characterized by rockfall, debris flows and slopewash features) grade distally and upwards into conglomerate, arenite and siltstone. The latter beds were deposited in somewhat more distal alluvial fan and floodplain settings formed by low-sinuosity and meandering streams (though, as indicated by localized presence of debris flows and the general coarseness and immaturity of the sediment, not very distal). The source for the epiclastic debris is inferred at first to have been both the fault scarps produced by the east-west trending high-angle faults, and the inherited topographic relief. The upper part of the member (including the condensed sections on the horst) also had a source lying to the southeast in the initial stages of the compressional orogenic highlands (as indicated by paleocurrent data, by local interfingering of I and J, and by the fact that the large monzonite boulders found in these levels would require significant unroofing and exposure). Siliceous vitric tuffs are present in several places in the section, as are andesitic and dacitic flows, and reflect minor ongoing volcanism in the vicinity as well. The transition from the Happy Creek to King Lear Formations is inferred to have been caused by a change from a primarily constructional volcanic regime to one controlled by several phases of tectonically-induced sedimentation. These phases must reflect fundamental stages in regional tectonic evolution.

Member J usually has a very sharp contact with I, though in some localities intermixing of clasts in a narrow zone at the top of I or bottom of J is present. J is distinguished by a completely different clast population, consisting of green and grey to black chert and quartzite and white vein quartz; andesitic debris of Happy Creek provenance is missing almost completely. Indeed, clasts in J have been dated at Triassic (radiolaria and corals), upper Norian (conodonts), Carboniferous (radiolaria) and Permian (radiolaria and conodonts). These clasts are very similar in lithology (as well as in age) to rocks in the Bliss Canyon (Triassic) and McGill Canyon (Late Paleozoic) Formations. These units certainly served as the source for the debris in member J. In order for this process to take place, these formations would already have to have been tectonically unroofed and exposed for erosion. Paleocurrent data in member J indicates a source to the south and east, within the approaching orogenic highlands (which had to already have begun forming in order to provide the detritus). The member is characteristically green (possibly as it lacked the Fe-oxides present in the Happy Creek in such abundance), and consists of conglomerate to shale lithologies. Many of the clasts are multicyclic and are reworked directly out of the Bliss Canyon and McGill Canyon Formations, where conlomerates form a significant proportion of the successions. The facies in J were laid down in an alluvial flood plain environment, still dominated by low-sinuosity meandering streams, but more distal than in member I (lacking debris flows, and with a higher percentage of finer-grained overbank facies). Upwards in the section, depositional environments become more distal yet. Tuffs still punctuate the section, as well as andesitic flows in one area and volcanic debris flows in another, indicating that some volcanic activity was continuing in the region. The lower part of the formation (members I and J) in several areas has growth fault relationships with some of the high-angle faults, while the upper part (above the Jackson Tuff) overlaps the faults with no evidence generally for syndepositional activity. One fault (the Jackson Creek structure) was active throughout the deposition of most of member J, inferred to have been overlapped only by the very top of the formation. East of Buff Peak, member J even unconformably overlaps the thrust faults and strata as low in the local section as the lower member of the Bliss Canyon Formation, and probably was deposited as an allochthonous or para-autochthonous piggyback basin on the allochthon.

The Jackson Tuff lies within the lower portion of member J, or where J is missing is taken as the boundary between I and K. The tuff is ignimbritic at the base, grading to a crystal-rich vitric ash-fall tuff, and is thickest nearest its feeder dikes or in depressions (both the grabens and the inherited topographic basins) where it ponded. These feeders cut, and the tuff itself generally overlaps, the family of high-angle faults (except for the Jackson Creek structure, or where the high-angle faults have been reactivated). The tuff formed by the collapse of an unusually large plinian eruptive column (both by thickness and lateral extent), and makes an excellent time horizon across the entire range. Three facies are seen in the Jackson Tuff: (i) basal glass; (ii) middle welded ignimbrite; and (iii) upper crystal-rich vitric ash-fall tuff. Facies iii is always present, but ii and especially i are restricted to the more proximal (to the fissure source) or valley-ponded occurrences.

There are also, in the middle of J in several locations and generally above the Jackson Tuff, interfingering lenses of andesitic epiclastics assigned to member K. J fingers out to the north within the large graben into a thick succession of K. Member K is identical to upper I in provenance, and was laid down in a middle alluvial fan environment by braided streams, sheetfloods and debris flows. The depositional environment becomes more distal upwards in the interval. As elsewhere in the King Lear, the climate was semi-arid (judging by the hematitic cement and calcrete formation in the red beds, by the tree fossils, and by the facies architecture). These red beds mark the uplift and erosion within the thrust zone of more Happy Creek rocks, perhaps as the thrust system propagated closer and new faults broke the surface and exposed basement nearby. The facies in K are noticeably more proximal than those in J, as well as being typically of braided and high-discharge origin as opposed to the low-sinuosity meandering and lower discharge facies of J. Continued uplift and exposure of deeper levels (or, alternatively, overlapping and burial of the source to member K) might then have lead to renewed deposition of facies of member J in the middle and southern range. The fact that J pinches out entirely to the north and west indicates a difference in the sourceland history in that area; apparently the Bliss Canyon and and McGill Canyon units were not exposed in the drainage feeding the northern fault-bounded basin (at least until later in the history of the formation, when the piggyback basin containing J east of Buff Peak formed). The implication is that the thrust belt may not have exposed deeper stratigraphic levels here, due perhaps to substantially less shortening. Alternatively, drainage patterns might have been such that debris from the thrust belt exposures of the McGill Canyon and Bliss Canyon units was diverted to the south. However, a causative mechanism to prevent sediment from entering a basin of substantial topographic relief is not evident, and an alternative source to the north or west for the epiclastic debris of K would be necessary. More tuffs are present in this member as well; at least minor levels of volcanic activity persisted throughout the formation.

Locally, volcanic centers in the upper part of the formation interfinger with upper members J and K and dominate the depositional patterns. Three such centers are present in the range - around Buff Peak, in the lowlands of Jackson Creek, and east of Bliss Canyon. The first two are rhyolitic, and are part of member L. The last one is dacitic, and belongs to member M.

Near Buff Peak, member L at the base consists of a thick but laterally restricted interval of white rhyolitic volcaniclastics interpreted as vent-proximal base-surge pyroclastic deposits. The upper part of the member is made up of thick glassy quartz$\pm$andesine$\pm$sanidine rhyolite flows. A large protrusive rhyolite dome with a breccia carapace in the superjacent thrust plate is interpreted as the subsurface extension of the actual vent for the Buff Peak center. The rhyolite protrusive dome and related dikes in this area cut the higher thrusts (an earlier phase?) and, with the pyroclastic and flow rocks are folded and cut by the frontal structures as well (a reactivated later phase?). As the lower contact was not exposed, it was not possible to tell whether the rhyolite bedded facies of L overlie K conformably or overlie thrusted rocks of member K, Happy Creek and Bliss Canyon with angular unconformity. The former conclusion is favored.

In the lowlands of Jackson Creek, a second but substantially smaller rhyolite protrusive dome intrudes members J and K and interfingers with member M. The exposed part is entirely an intrusive rhyolite breccia. The dome has a rhyolitic cap succession, with a basal epiclastic horizon indicating dome quiesence and reworking of the surface of the dome. Dome reactivation is indicated by an overlying welded tuff and then a quartz $\pm$ sandidine $\pm$ andesine $\pm$ topaz rhyolite flow interval. The epiclastic and flow successions extend several kilometers laterally, but are separated by several hundred meters of member M - a further indication of dome resurgence. After the flows were extruded, the dome was gradually overlapped and buried by members M and J. This volcanic center, like the others in the King Lear Formation, lies within one of the pre-existing basins where the King Lear section is much thicker.

To the south, east of Bliss Canyon, is a large and complex dacitic volcanic center. The subsurface portion is made up of an elliptical feeder stock, again with a breccia shell, and a series of thick sills. The overall structure is of an andesine$\pm$quartz$\pm$biotite dacite porphyry Christmas-tree laccolith. The laccolith intrudes all levels of the King Lear (though not to any extent the Happy Creek), including its own extrusive equivalents in member M, and has peperitic relationships with the King Lear. The sills are generally concordant but cut down-section to the south overall. Here basement topography caused the King Lear succession to be much thinner and topographically higher over the low-angle buttress unconformity. As the sills were inferred to be intruded at a constant depth (that is, a zero-density-contrast surface), they would cut down into lower stratigraphic levels over the basement highs. The extrusive portion of member M consists of proximal alluvial fan to colluvial epiclastic deposits, lahars (volcanic mudflows), crystal tuffs, and rare flows. The epiclastic facies are characterized by the presence of dacite porphyry rock fragments and copious detrital plagioclase and quartz, but generally lack andesitic or cherty/lithic debris identifiable as coming from the Happy Creek, Bliss Canyon or McGill Canyon Formations. Deposition of the epiclastics was by braided streams, sheetflood and debris flow. Member M interdigitates with and grades into member J away from the volcanic center, and particularly in the region of the horst. The most distal facies are characterized by lahars and tuffs. Synsedimentary and synvolcanic thrusting is indicated by flaser zones in the sills, and by compressional soft-sediment deformation in the lacustrine turbidites that cap the section in Trout Creek.

Additional evidence for the synsedimentary nature of thrusting is presented by member N. This member occurs immediately below several of the frontal thrusts, and conformably overlies with conformity or overlies angular unconformity either various levels of the King Lear, or even (within the thrust zone) the Happy Creek. The member is quite thin where present, and was deposited in colluvial fan and fault scarp talus environments (and dips gently away from the thrust frontal scarp along lateral accretion bedding surfaces). Uplift, deformation and removal of portions of the King Lear due to the impinging thrust front followed by the deposition of sedimentary fault scarp breccias from the thrust front characterize N. The member would then have been overrun by the thrust plate as it continued to advance along the paleosurface. The fact that these sedimentary breccias occur in several distinct structural positions indicates either that thrusting was considerably out-of-sequence, or that several distinct phases of thrusting existed with reactivation (as the substrate to the upper-level occurrences of the breccia had already been uplifted, exposed and stripped down before the overriding thrust plate was emplaced over its breccia). The thrusts also reactivated many of the high-angle structures as tear faults, although this did not affect the sedimentary record (except perhaps in N) as it was post-depositional.

Member I contains clasts identical to intrusive bodies dated at 173.3 $\pm$ 14.3 m.y. (two dioritic and monzonitic stocks, Rb/Sr - R. Kistler, written communication, 1988), 187 $\pm$ 2 m.y. (monzonite stock - U/Pb on zircon), 170 - 175 m.y. (a diorite sill in the upper stratigraphy of the Happy Creek -U/Pb on zircon), and 169 $\pm$ 10 m.y. (member G of the Happy Creek - Rb/Sr, Russell, 1981). Outside the grabens, member I also nonconformably overlies the 170 - 175 m.y. diorite sill, and conformably overlies and interfingers with member G of the Happy Creek. Some of the high-angle faults active during the deposition of member I in the basins are cut by the plutons dated at 173.3 $\pm$ 14.3 m.y., while offsetting that pluton dated at 187 $\pm$ 2 m.y. The basin formation and deposition of member I within those basins thus began in the Middle Jurassic (during or after the Aalenian, but certainly by the Bathonian). Some of the faults continued activity into the Late Jurassic. Up on the horsts, deposition of member I began some time during the Bathonian or Callovian (very soon after member G, and the diorite sills that intrude G). As the sedimentary record indicates that thrusting and high-angle fault activity overlapped temporally though separated paleogeographically (the thrust-generated cherty/lithic molasse of lower J has growth fault relationships in the basins), thrusting in the first phase of activity on the eastern, west-vergent thrust system had begun by the end of this period. Some thrusts in Trout Creek Spur (higher up in the structural stack, and associated with exposures of member N) are also stitched by a pluton dated at 162 $\pm$ 1 m.y. (U/Pb on zircon), so that the initial phase of thrusting had begun (and largely ended) before that time. In addition, some of the exposures around DeLong Peak were topped by breccia of N and related thrust packages at very low stratigraphic levels, still quite early in the history of the King Lear. The first phase of thrusting may have been over by the Callovian or Oxfordian, during the deposition of lower member J and member K (with a Late Triassic to Jurassic age). However, middle to upper member J, also molasse shed from the orogenic highlands (though of more distal environments), has fossil dates of Triassic to Early Cretaceous, Cretaceous(?), and Early Cretaceous. Molasse sedimentation in the King Lear therefore continued from the late Middle Jurassic (Bathonian to Callovian) through to the Early Cretaceous, after the first phase of thrusting apparently ceased. The orogenic highlands persisted for some time, though becoming less pronounced with time. No evidence for a major hiatus exists anywhere within the formation (except at the base of N). Reactivation of the thrusts during at least the Aptian (and starting, perhaps, somewhat earlier) is indicated by the synthrusting features of M (seen in the laccolith dated at 115 $\pm$ 1 m.y., and in epiclastic sediments of lower Cretaceous age), by the overlapping of thrusted older units by member J (near Buff Peak), and by the relationships exhibited by N (the frontal thrusts and related occurrences of N are from this second phase; sites higher in the structural stack are inferred to be from the first phase). Whether thrusting was continuously active between the Aptian and Callovian is not clear, but is not probable on kinematic grounds - too little shortening has taken place.

The silicic volcanic center around Buff Peak is of Early Cretaceous age, overlapping in time with the second phase of thrusting. The dacitic volcanic facies of M are 115 $\pm$ 1 m.y. (the sill) or older, as the sill intrudes them. The center was fairly long-lived, judging by its stratigraphic extent, and was active from the middle to later Early Cretaceous. The small rhyolite dome and related facies of member L in Jackson Creek started just before or at the same time as M, but ceased activity earlier than M did. As this dome intrudes and overlaps Jurassic to Lower Cretaceous strata, it was active in the middle part of the Early Cretaceous as well. The copious tuffs in the King Lear indicate ongoing nearby silicic volcanic activity from the Middle Jurassic through to the late Early Cretaceous.

CHAPTER SIX:
THE EARLY MESOZOIC INTRUSIVE SUITE
OF THE JACKSON MOUNTAINS, NW NEVADA

A related suite of stocks, dikes, and sills intrudes all the bedded units of the Jackson Mountains, except the McGill Canyon Formation. The igneous rocks in the suite cover a wide range in composition and have a variety of intrusive settings and cross-cutting relationships with the stratigraphic and structural phases, therefore aiding greatly in reconstructing the geologic history of the range. These crystalline intrusive lithologies are placed in the Early Mesozoic Intrusive Suite.

The bodies of the Early Mesozoic Intrusive Suite discussed in this chapter are inferred to be genetically related to one another and to the volcanism expressed in the Happy Creek Formation, and to a lesser extent in the King Lear. This is based on petrography, geochemistry, intrusive setting, and temporal and spatial overlap. Hypabyssal, glassy intrusive facies of the King Lear, which are intimately associated with the volcanic centers expressed in the formation, are discussed separately in chapter five. The few dikes in the McGill Canyon Formation, tied to the volcanic flows in that unit, are treated with them in chapter two.

The intrusive bodies included in the Early Mesozoic Intrusive Suite are (see fig. 51 for their locations - they are listed counterclockwise from the north tip of the range): (1) the Happy Creek stock; (2) the Deer Creek Peak stock; (3) the Parrot Peak pluton; (4) the Harrison Grove stock; (5) the Mary Sloan Creek dike swarm; (6) the Bliss Canyon

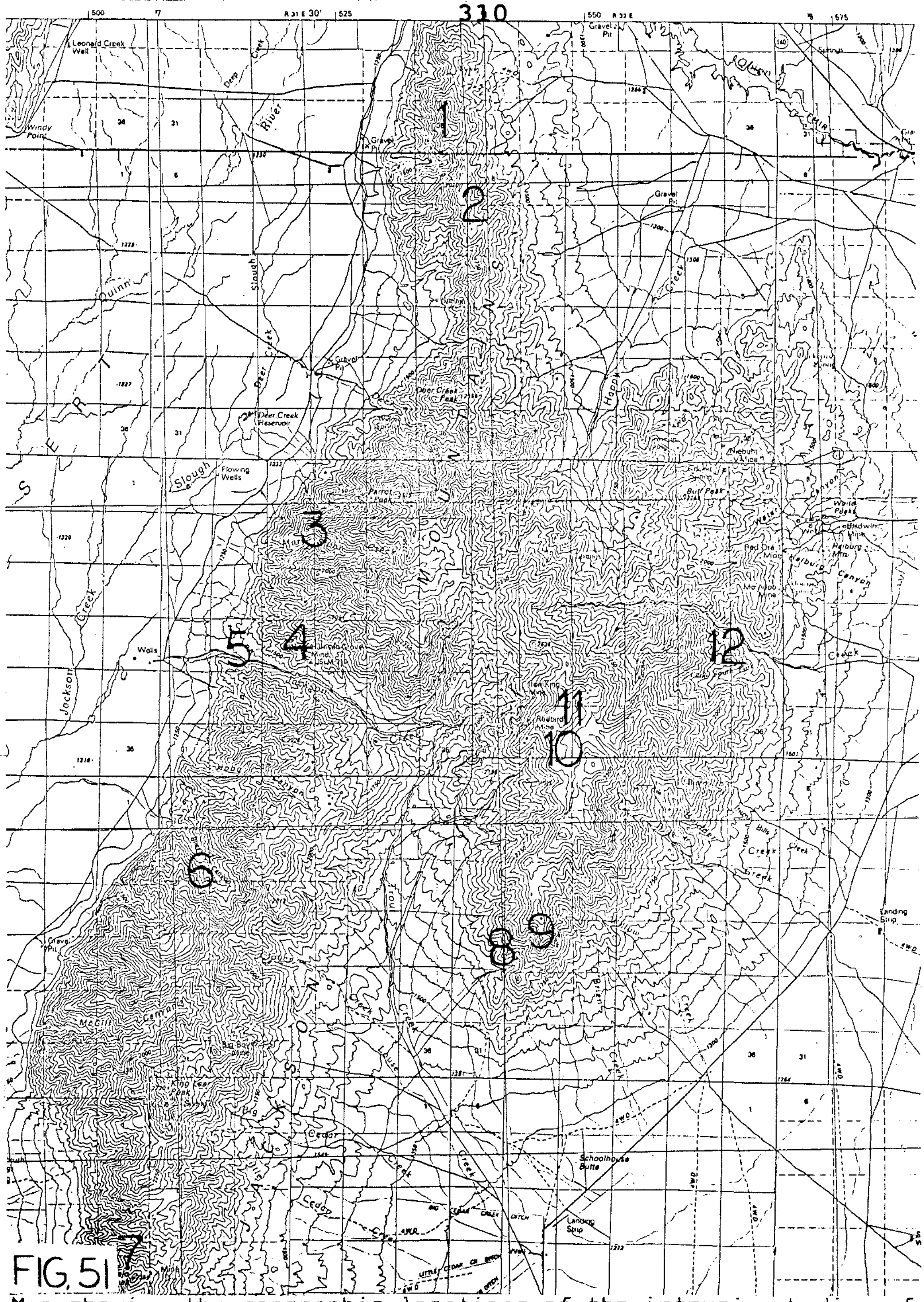


FIG. 51

Map showing the geographic locations of the intrusive bodies of the Early Mesozoic Intrusive Suite in the Jackson Mountains. The grid gives boxes one mile on a side. Localities (after the nearest named geographic entity) are: (1) Happy Creek stock; (2) Deer Creek Peak stock; (3) Parrot Peak pluton; (4) Harrison Grove stock; (5) Mary Sloan dike swarm; (6) Bliss Canyon stock; (7) Navajo Peak stock; (8) Trout Creek stock; (9) Trout Creek Spur stock; (10) diorite sills in the Happy Creek Formation; (11) gabbro sills (also in the Happy Creek Formation); and (12) Bottle Creek suite of dikes and sills.

stock; (7) the Navajo Peak stock; (8) the Trout Creek stock; (9) the Trout Creek Spur stock; (10) the diorite sills intruding the Happy Creek Formation in the eastern part of the range; (11) the gabbro sills, also intruding the Happy Creek Formation in the eastern part of the range; and (12) the Bottle Creek suite of dikes and sills (including samples from the Jungo Hills to the south, an area not part of the mapped thesis area). These bodies will be dicussed in the order listed.

For each body or group of bodies listed, the analysis will include: (i) the field appearance and petrography; (ii) the intrusive and cross-cutting relationships; and (iii) the age data pertinent to that locality. At the end of the chapter, the intrusive framework, setting, and evolution of the Early Mesozoic Intrusive Suite as a whole will be discussed.

The geochemistry of the Early Mesozoic Intrusive Suite will be discussed in a later appendix covering the geochemistry of all the igneous rocks in the range. This section also includes the igneous phases in the McGill Canyon, Happy Creek, and King Lear Formations and the Tertiary volcanic assemblage.

The Happy Creek, Deer Creek Peak, and Bliss Canyon stocks and the diorite sills intrude only the Happy Creek Formation. The Parrot Peak, Harrison Grove, Trout Creek and Trout Creek Spur plutons, the Mary Sloan Creek and Bottle Creek dike suites, and the gabbro sills intrude the Bliss Canyon Formation as well as the Happy Creek. The Navajo Peak stock intrudes the base of the King Lear Formation as well as the Happy Creek. The stocks and dikes on the western side of the range are cut by thrusts of the western belt (and in one case intrude a thrust). Thrusts of the eastern belt imbricate the sills but are stitched by the stocks and dikes. The high-angle faults offset the sills and the older stocks, but are in turn cut by the younger intrusive bodies. Xenoliths and detrital clasts of the intrusives (mostly diorite, tonalite, granodiorite and granite) are found in the King Lear and uppermost Happy Creek Formations, but not the lower Happy Creek. These relationships, where the plutons or the units they intrude are dated, allow the bracketing of these major depositional, magmatic, and tectonic phases in the geologic evolution of the area.

Contact metamorphic aureoles are found around the Happy Creek, Parrot Peak, Harrison Grove, Navajo Peak, and Trout Creek Spur stocks. The diorite sills and their immediate country rock surroundings are extensively hydrothermally autometamorphosed and altered as well.

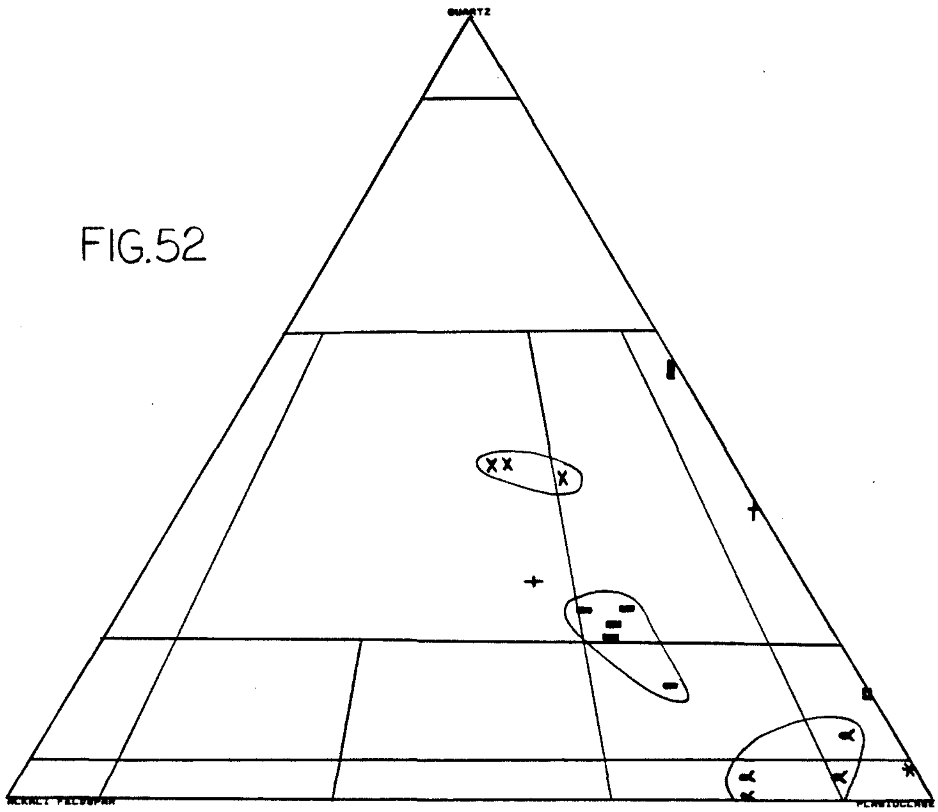
(1) Happy Creek Stock

This body is located in the extreme northwest of the range and covers about 20 km^2. It intrudes the Happy Creek Formation (members A, B and E) on the north, south and west, and the older Deer Creek Peak pluton on the southeast. On the east it is overlapped by Tertiary basalt flows. Willden (1963) mapped it as Jd (Jurassic diorite) together with the Deer Creek Peak body, which it intrudes. Russell (1981) did not map this far north, and the state map (Stewart and Carlson, 1978) does not show it at all. The contacts are sharp, with abundant diking and brittle veining of the country rock. The border zone of the stock is full of angular stoped blocks and xenoliths of andesite or of the Deer Creek Peak stock, and septa or pendants of andesite, up to 2 km in length and a third that in width are found within the interior. On the west side of the pluton along the range front the contact zone becomes a megabreccia, with very large blocks brittly plucked out and contained in the edge of the stock. This zone grades into an extensively diked region, over several tens of meters. The dikes are very brittle in nature, with sharp, angular contacts. The andesite immediately adjacent to the margin of the stock (and within the stoped blocks) is mildly metamorphosed. Amphibolite and epidote veins characterized the contact zone as well.

In the field, the Happy Creek pluton was characterized as quartz monzonite (after Streickeisen, 1973), with roughly 30% plagioclase, 10% quartz, 20% mafics (hornblende and FeTi-oxides) and 40% alkali feldspar. It is fine to medium-grained and phaneritic, pink to pinkish-grey in color, unaltered and fresh, and equigranular to slightly porphyritic in texture. The core appears to be slightly more mafic, with some of the hornblende taking on a poikolitic texture. The margin phases are granitic, and lacking almost entirely in mafics. The pegmatitic dikes and apophyses in the contact zone were also classified in the field as being granitic in composition, and are coarse-grained.

Petrographically, the dikes are indeed granite (see fig. 52). They contain quartz, alkali feldspar and albite in subequal amounts, with minor chlorite and trace tourmaline and sphene. The texture is subhedral granular. The stock itself is granite to granodiorite, containing quartz, alkali feldspar, andesine plagioclase and hornblende, with minor FeTi-oxides and biotite and trace zircon and apatite. This phase is also subhedral granular to subophitic, generally fresh and with the mafics occuring together in large clumps. In some areas there is noticeable

FIG.52



Petrographic modal analyses from the stocks of the Early Mesozoic Intrusive Suite of the Jackson Mountains, plotted on a Q-A-P diagram for the classification of plutonic rocks (after Streickeisen, 1973). Modal percentages were visually estimated. The numbers in front of the pluton names gives the location on the map and the text section discussing that intrusive body.

X (1) Happy Creek stock (granodiorite to granite)

(2) Deer Creek Peak stock (granite)

(3) Parrot Peak stock (quartz diorite)

(4) Harrison Grove stock (granodiorite to quartz monzodiorite)

(6) Bliss Canyon stock (tonalite)

* (7) Navajo Peak stock (gabbro - plagioclase An value >50)

(8) Trout Creek stock (tonalite)

(9) Trout Creek Spur stock (quartz diorite, diorite, and monzodiorite)

alteration, with sericitization of the plagioclase and with the biotite being replaced by chlorite. The plagioclase has normal to oscillatory zoning.

Two high-angle faults cut the Happy Creek Formation in this area. The fault that intersects the stock is truncated by it, and therefore predates intrusion. The highly discordant contacts, lack of significant regional metamorphism in the country rocks, simple plan, brittle diking and stoping, and lack of complex internal structure or of igneous fabrics all indicate a shallow crustal origin within the "epizone" of Buddington (1959), at a depth of less than about 4 to 6 km. This accords with the estimate of the thicknesses of the overlying stratigraphic column at the time of intrusion (composed of the Happy Creek and King Lear Formations) of a maximum of about 3 km, and probably less. Note that the body is dated as intruding during or just before the early phases of King Lear sedimentation, before more than a few hundred meters or so of the unit had accumulated.

Together with the Parrot Peak stock, this body was dated by Rb/Sr geochronology at 173.3 ± 14.3 m.y. (R. Kistler, written communication, 1988).

(2) Deer Creek Peak Stock

Lying on the eastern flank of the northwestern extension of the Jackson Mountains, this body covers about 5 km^2. It is intruded by the granitic to granodioritic Happy Creek stock on the west, and along the contact is full of granitic pegmatite dikes. In turn it intrudes the Happy Creek Formation (member E) on the south and southeast, and to the east it is unconformably overlapped by Tertiary basalt flows. Clasts and boulders of this body are present in the Tertiary gravel lens beneath the basalts in the saddle just south of the outcrop area. Where the Deer Creek Peak stock intrudes the Happy Creek along preserved portions of its margin, it is much finer-grained and very xenolith-rich. The angular xenoliths are of green andesite.

In the field, the body was classified as a quartz diorite to monzodiorite, with plagioclase and unidentified mafics and minor quartz and alkali feldspar identified. In one thin section, a more silicic and less abundant granite phase was identified, with quartz, alkali feldspar, andesine, hornblende, and minor FeTi-oxides and biotite. The plagioclase is normally zoned from An_{43} to An_{24}. The crystallization sequence was plagioclase hornblende alkali feldspar quartz. The mafics tend to occur in clots, and the quartz as interstitial to poikolitic patches. The texture is subhedral granular.

This body must be older than 173.3 $\pm$ 14.3 m.y. because it is intruded by a stock dated at that age. It also must be substantially younger than 204.6 $\pm$ 13.6 m.y., because it intrudes high up in a section of the Happy Creek that has been given that date by Rb/Sr geochronology at its base (R. Kistler, written communication, 1988).

(3) Parrot Peak Pluton

Situated north of the mouth of Jackson Creek in the western part of the range, this body was mapped by Willden (1963) as TKg - Tertiary to Cretaceous granite (together with the Harrison Grove stock). Similarly, Russell (1981) mapped both of these intrusives as Kgd - Cretaceous granodiorite. On the state geological map (Stewart and Carlson, 1978) they are both listed as Mzgr - Mesozoic granite. The body is a separate entity from the Harrison Grove stock, and covers about 18 km^2.

On its western side, the pluton is truncated and overthrust by nappes containing Bliss Canyon Formation rocks. Around its northern, eastern and southern margins it intrudes the Happy Creek and Bliss Canyon Formations. The pluton is inferred to cut the high-angle faults expressed just to the east in the headwaters of Mary Sloan Creek, as these faults extend toward the margin of the stock without appearing to offset it. The Mary Sloan Creek dike swarm cuts through the intrusive, and several large dioritic apophyses are injected by the pluton into the dike swarm. The pluton is also elongated slightly in the direction of the swarm.

The stock is made up of dark green diorite, with poikilitic hornblende and plagioclase. The marginal zone, tens of meters thick, is a finer-grained, lighter-colored dioritic phase. Near the thrust faults the body has a local schistose fabric.

This body and the Happy Creek stock are colinear together on a Rb/Sr plot at 173.3 $\pm$ 14.3 m.y., as mentioned earlier.

(4) Harrison Grove Stock

The Harrison Grove stock is found north of the Harrison Grove Mine and the lower reaches of Jackson Creek near the mouth. In the previous mapping efforts it has been lumped together with the Parrot Peak pluton, but they are distinct in petrography and geography (see discussion for that body). The area covered by the body is about 2.5 km^2. It intrudes the lowest Happy Creek and upper Bliss Canyon Formations, and is also overthrust by the western thrust system. A substantial contact metamorphic aureole is associated with this body, the Parrot Peak pluton and the Mary Sloan Creek dike swarm. One of the high-angle faults cuts across and offsets the pluton. The pluton is brecciated and mineralized along this fault plane.

The stock was classified in the field as a quartz monzonite to monzodiorite, with plagioclase, hornblende, alkali feldspar, and minor quartz, muscovite and FeTi-oxides seen. The texture is phaneritic and fine- to medium-grained to slightly porphyritic, with larger plagioclase laths in a finer-grained matrix. The color is pink to pinkish-grey. The plagioclase is often green and zoned, and the alkali feldspar is pink. Some zoning in the intrusive is present, with the alkali feldspar, muscovite and quartz component diminishing inward from the margins. The stock is cut by a number of dikes, and locally is foliated and highly altered with abundant epidote veining. In these altered areas the plagioclase grains are chalky and the mafics replaced by chlorite and epidote.

Petrographically, the body is a granodiorite to quartz monzodiorite. An assemblage of plagioclase and hornblende with minor to major amounts of quartz, alkali feldspar and biotite, minor FeTi-oxides (and uncommonly orthopyroxene), and trace apatite and zircon characterizes the pluton. The plagioclase is andesine (An_{35} to An_{50}), and can be highly turbid or sericitized. In some cases, oscillatory to normal zoning of the plagioclase is present. The hornblende is green to greenish-brown, subophitic to poikolitic, and is replaced by a combination of chlorite, epidote and calcite. The biotite is also replaced by chlorite in some samples. The alkali feldspar can be sanidine, orthoclase, or microcline (with perthitic textures) depending on the sample. The quartz formed as a late-stage interstitial phase, subophitic to poikolitic. The texture is anhedral or subhedral granular to slightly porphyritic.

Epidote seams commonly criss-cross the igneous textures, and sprays of tourmaline (elbaite) overprint them as well. In some samples a chloritic cleavage is developed. The alteration predates the foliation, as the replacive chlorite, for example, can be kinked and deformed and the tourmaline offset by microfaults.

As with the other plutons in this area, an epizonal emplacement site is indicated by the discordant and sharp contacts of the body, lack of internal igneous foliation or lineation, associated brittle diking, stoping and veining, restricted contact aureole, simple internal structure, and inferred stratigraphic overburden at the time of intrusion (Buddington, 1959). This pluton could have been intruded at a depth of no more than 2 - 3 km when intruded, as that would be the maximum thickness of the Happy Creek Formation at that time.

U/Pb geochronologic analysis of this pluton has given a concordant date of 187 $\pm$ 2 m.y.

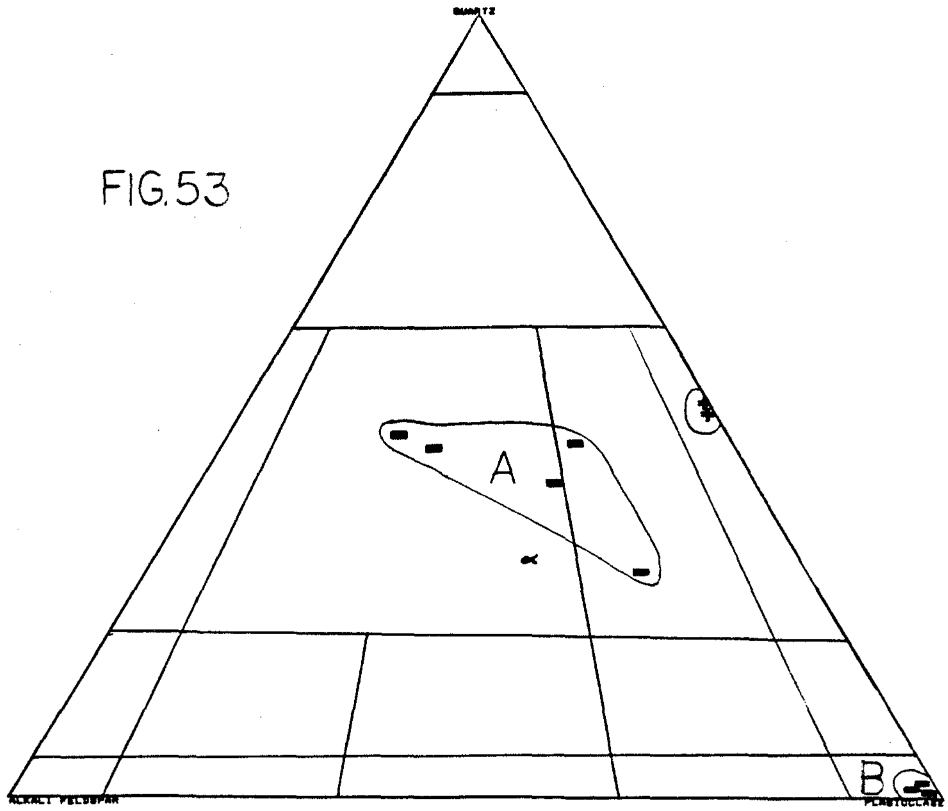
(5) Mary Sloan Creek Dike Swarm

This family of intrusives is found on the western range front, north and south of the mouth of Jackson Creek. The dikes are generally too thin to be mappable. They intrude the Bliss Canyon and Happy Creek Formations and the Parrot Peak and Harrison Grove plutons. The western thrust system truncates and deforms the dikes, which were nowhere seen to in turn cut it. The dikes are found in both the lower and upper plates and in the footwall. Along the thrust fault zones themselves, granitic dikes are ductilely deformed and boudinaged into pods strung out along the fault. Elsewhere, the dikes are commonly structurally foliated to schistose.

In places the dikes make up as much as 40% of the rock by volume, though generally less than 10%. The swarm thins and disappears to the north and south along strike (with an extent along strike of 15 km), and to the east as well (a width of 3 km or less). The orientation is quite systematic and constant throughout the swarm at N05E to N25E with moderate to steep eastward dips; minor diking also exists trending N75E. The dikes are planar and have sharp and highly discordant boundaries, and are oriented parallel or at only a slight angle to one another. Spatially the dikes are associated with a significant development of contact metamorphism and with a set of plutons (the Happy Creek, Parrot Peak and Harrison Grove stocks - this set of associations is not coincidence). Some of the thicker dikes are finger-like apophyses coming off of the first two of those plutons. With these exceptions, the dikes generally are several meters in thickness.

In the field, at least five different lithologies were seen to make up the dike swarm (fig. 53). Each appears to correspond with extrusive facies in the Happy Creek Formation, or with an intrusive phase of the Early Mesozoic Intrusive Suite plutons. These lithologies include: (1) Pink granitic coarse-grained phaneritic and equigranular dikes with alkali feldspar, quartz, plagioclase and biotite or hornblende. These

FIG.53



Petrographic modal analyses from the dikes of the Early Mesozoic Intrusive Suite of the Jackson Mountains, plotted on a Q-A-P diagram for the classification of plutonic rocks (after Streickeisen, 1973). Modal percentages were visually estimated. The numbers in front of the dike suite names gives the location on the map and the text section discussing that intrusive suite. Note the similarity of field A in the Mary Sloan Creek dike composite dike swarm to the nearby Harrison Grove stock, and of field B to the adjacent Parrot Peak stock.

▬ (5) Mary Sloan Creek composite dike swarm (granodioritic to granitic - field A, and dioritic - field B)

∝ (9) Granitic dike related to the Trout Creek Spur stock

(12) Dikes of the Bottle Creek/Jungo Hills dike suite (tonalitic)

pegmatites are the most common phase, and some are associated with (and intrude) the Harrison Grove stock in particular. They are also rather similar to the Happy Creek pluton, which has copious diking associated with it as well. Petrographically, the granitic set of dikes is anhedral or subhedral equigranular in texture, and contains alkali feldspar, albite plagioclase, and quartz with minor hornblende, biotite, muscovite, FeTi-oxides, epidote, chlorite and tourmaline (elbaite). The plagioclase can be extremely turbid and sericitized, and the hornblende and biotite is replaced by chlorite, calcite, epidote and FeTi-oxides. Some samples are foliated, with a spaced fracture cleavage developed and overprinting the replacement assemblage.

(2) White tonalitic to quartz dioritic fine-grained dikes, with plagioclase and quartz. This phase is the one seen in the actual thrust zone. These dikes are similar to the Bliss Canyon stock and to the Bottle Creek suite of dikes.

(3) Green andesite to dacitic dikes, porphyritic with plagioclase $\pm$ quartz and hornblende phenocrysts. The dikes have a fine-grained to glassy (though devitrified) groundmass, trachytic texture, and can have flow-foliated and chilled margins. The margins of the enclosing country rock are also slightly metamorphosed. These dikes are very similar to the flows in members E, F and G of the Happy Creek Formation, and are probably comagmatic. In thin section the andesite dikes contain andesine plagioclase, clinopyroxene or hornblende and minor quartz and alkali feldspar. The texture is intergranular and the matrix is not as fine-grained as in the lithologically similar Happy Creek flows.

(4) Diorite, with plagioclase, hornblende and/or pyroxene, medium-

grained and grey in color. This phase is most similar to the Parrot Peak stock. The diorite dikes petrographically are characterized by andesine plagioclase, hornblende, clinopyroxene and FeTi-oxides and can have a glassy to very fine-grained groundmass with good alignment of the phenocrysts. These dikes also are altered, with sericitization of the plagioclase and replacement of the mafic phases by the Fe-Ti oxides, epidote, calcite and chlorite.

(5) Gabbro, dark grey and coarse-grained, with pyroxene and plagioclase. The gabbro resembles the sills seen elsewhere intruding the Happy Creek Formation.

Due to its variability and extent, this dike swarm indicates a long-lived and complex magmatic history. All the plutons discussed so far but the Happy Creek are cut by at least some of the dikes. Although the opposite relationship was not observed, it is inferred to exist as some dike families do not crop out within certain plutons. The dikes are concluded to cover the same age span of the Happy Creek Formation and the Early Mesozoic Intrusive Suite stocks that they resemble, and are correlated to. The intrusion of the swarm probably spanned the age range of Early to Middle Jurassic.

The systematic orientation of these dikes probably reflects the regional state of stress. The implication to be drawn from this is that during dike injection $\acute{U}_1$ was oriented vertical; $\acute{U}_2$ was N20E and horizontal; and $\acute{U}_3$ was perpendicular to the plane formed by these two. This plane of least stress (perpendicular to $\acute{U}_3$) is characteristically that along which dike swarms are injected (Jaeger, 1969; MacBirney, 1979). Perhaps not coincidentally, this is parallel to the trend observed in the silicic volcanic features of the younger King Lear Formation.

(6) Bliss Canyon Stock

This body intrudes the Happy Creek Formation (members C and E), and is truncated and structurally overridden by an east-vergent thrust slice containing the Bliss Canyon Formation. It was previously mapped as unit KJi2 of Russell (1981), and as Tf (Tertiary microdacite intrusive) of Willden (1963). It is only about 1 km^2 in extent. The intrusive contact is irregular but sharp, with a very narrow and minor contact aureole. Red jasper veins cut the intrusive, which was classified in the field as a white, fine-grained tonalite. The exposure can be seen to be the top of the intrusive body, as portions of the overlying carapace are present and the contact is curved up over the stock in a dome-like fashion.

Petrographically, the stock is actually a tonalite, and is composed of albite and quartz, with a minor mafic phase (either hornblende or biotite - the mineral is too fine-grained to tell). The texture is very fine-grained and anhedral granular to porphyritic, with sparse large plagioclase phenocrysts.

The body is undated, and is a good candidate for U/Pb zircon geochronologic analysis. It must postdate the strata it intrudes (which are Lower to Middle Jurassic) and predate the thrusts (which may be Aptian, but certainly need to be better dated).

(7) Navajo Peak Stock

This body lies at the extreme south of the map area, making up Navajo

Peak and its immediate environs. It was mapped as TKd (Tertiary to Cretaceous diorite) by Willden (1963), and extends far to the south; the area within the mapped region for this study is about 10 km^2. This body is truncated on the west by a thrust fault of the western system, with a slaty volcaniclastic interval inferred to belong to the McGill Canyon Formation in the upper plate. However, the pluton also stitches another east-vergent thrust on its northern border, and cuts across already folded strata (see chapter eight for a discussion of these relationships). The stock intrudes members E, F and G of the Happy Creek, and even member I at the base of the King Lear. This is the only place in the field area where the King Lear is intruded by the Early Mesozoic Intrusive Suite. A limited but significant contact aureole is present around the body.

The Navajo Peak stock is more complex than any of the others in the range. The core is made up of a coarse-grained dark green monzogabbro, with pyroxene, plagioclase and minor alkali feldspar identified in the field. The grain size, color index, and composition of the mafic phase vary somewhat through the body. The core grades into a partially enclosing shell of uneven thickness and made up of lighter-colored, finer-grained diorite. This is in turn surrounded by patches of pink, coarse-grained granite, fine- to medium-grained and composed of plagioclase, alkali feldspar and quartz. The granite phase contains bleached xenoliths of the gabbro and diorite. Granitic, pegmatitic and aplitic veins also intrude the more mafic central phases of the pluton as well as the country rock. Magnetite seams are also present.

In thin section, one sample of a less mafic portion of the central grabbroic phase was found to contain labradorite plagioclase, biotite, hypersthene and minor quartz and Fe-Ti oxides. The texture was medium-grained and subhedral granular, and the plagioclase normally zoned with cores of An_{68} and rims of An_{40}. The orthopyroxene is partially replaced by biotite, and both have been partially altered to chlorite. This particular sample is a quartz gabbro.

This zoned pluton must postdate member I of the King Lear (which is upper Middle Jurassic), and overlap the time of thrusting. The thrusting here might have occurred in two phases as elsewhere in the range (late Middle Jurassic and Aptian), with the intrusion emplaced at some time in between. Alternatively, both thrusts might be Aptian and the stock the same age. I strongly favor the first alternative. The granitic carapace would make a suitable candidate for geochronologic dating, as this lithology within the range has proved productive for zircon.

(8) Trout Creek Stock

This body is located on the western side of the tip of Trout Creek spur, and covers less than 1 km^2. It was mapped together with the Trout Creek Spur stock by Willden (1963) as Jurassic diorite, and separately by Russell (1981) as Cretaceous or Jurassic felsite. The contacts range from sharp and discordant to an intrusive breccia zone, with no noticeable contact aureole. The stock intrudes and stitches the eastern thrust system, including an imbricated sequence containing the Bliss Canyon/Happy Creek contact, the lower Happy Creek Formation, and both the diorite and gabbro sills. It intrudes the Trout Creek Spur stock.

The Trout Creek stock in the field is a fine-grained to very fine-grained, white to very light grey or tan tonalite, with plagioclase and quartz. In thin section it is composed of quartz and andesine plagioclase, minor hornblende and muscovite, and trace zircon and Fe-Ti oxides with an anhedral granular (i.e., granitic) texture.

Zircons have been extracted from the body, but have not yet been analyzed for geochronology.

(9) Trout Creek Spur Stock

This stock covers part of the spine of the tip of the Trout Creek Spur, with an area of about 5 km^2. Both Stewart and Carlson (1978) and Russell (1981) classified it as Cretaceous or Jurassic diorite, and Willden (1963) as Jurassic diorite. The contacts are highly discordant and the stock intrudes the imbricated Bliss Canyon and Happy Creek Formations in the eastern thrust system, stitching several of the faults. The stock also cuts and dikes a gabbro laccolith and may intrude the uppermost McGill Canyon Formation.

One area where the intrusive contact was unusually well-displayed was near the top of the pluton, and displays a 4 m-wide zone of intense brecciation. The breccia is made up of green, yellow or pink and glassy to very fine-grained clasts, and pieces of the surrounding andesite. The country rock also has pink aphanitic dikes and veinlets right near the contact. The breccia zone grades into a coarse-grained monzodiorite (alkali feldpsar, hornblende, plagioclase and trace sphene); this phase is also cut by pink glassy dikes. This breccia margin is present at several localities. Though the contact was chilled, no contact metamorphism was seen at this or other sites in this pluton. Elsewhere, the border of the pluton is 10 m of pink and grey aphanitic to very fine-grained aplitic material, and is breached by late-stage dikes.

The interior of the pluton is principally monzodiorite, very coarse-grained and rich in pyroxene. The stock and its companion dikes lack the fracturing, foliation and epidote veining commonly associated with the thrusted rocks (including the gabbro and diorite sills), but are instead quite fresh. The dikes, which are quite abundant near the stock, are up to 3 m in thickness and are granitic pegmatites.

Petrographically, the stock contains normally zoned plagioclase (andesine cores to oligoclase rims), hornblende, alkali feldspar (microcline, sanidine and orthoclase are all present in different samples), minor quartz, biotite, chlorite, Fe-Ti oxides and trace sphene, apatite, and zircon. Augite was present in some samples as well and is rimmed by hornblende, and sphene is locally quite abundant. Replacement of hornblende and biotite by chlorite and sericitization of the plagioclase has taken place, though not extensively. Some of the alkali feldspars have microperthite developed. Textures are subhedral granular to poikolitic, with plates of hornblende enclosing plagioclase. The order of crystallization was plagioclase Fe-Ti oxides sphene hornblende alkali feldspar chlorite quartz. The hornblende in particular is often clotted. The lithology of the stock occupies the range diorite to monzodiorite to quartz monzodiorite.

The dikes in thin-section are granite - composed of equal amounts of plagioclase, orthoclase, and quartz, and crystallized in that order. They are fine-grained to very fine-grained and anhedral granular, and lack mafic phases. They cut the pluton as well as the country rock.

In one sample, brittle deformation was evidenced by extensive kinking of the albite twinning in the plagioclase, and indicates some later deformational phase in this area. The stock itself was definitely intruded at a very high crustal level (epizonal), judging by the lack of contact metamorphism, and by the chilled and brecciated margin, brittle diking, high degree of discordance, simple plan and lack of internal igneous fabrics.

This stock is dated at 162 $\pm$ 1 m.y. by U/Pb geochronology on zircons.

(10) Diorite Sills, Eastern Jackson Mountains

These sills are restricted, with a minor exception, to the eastern Jackson Mountains, in the middle and southern portions of Trout Creek Spur. Russell (1981) included this area in what he termed an "intrusive center", though not recognizing that the intrusives were a number of distinct imbricated tabular and concordant mafic intrusions. The sills intrude the Happy Creek Formation, are in turn intruded by the gabbro sill phase, and then all of these units have been imbricated in the eastern thrust system. The stack was then intruded, by the Trout Creek Spur stock. This family includes at least six separate bodies. Two are small (about 70 and 100 m thick and 2 km in width) and are found on top of one another in members A and E south of Bottle Creek. Another small body (as much as 200 m thick, but truncated by a gabbro sill on one end and by a thrust on the other) intrudes the Bliss Canyon (and possibly the McGill Canyon) Formation at the tip of Trout Creek Spur.

An additional very thin sill (only about 10 m thick) intrudes along the contact of member G of the Happy Creek and member I of the King Lear just north of Deer Creek Peak. This last sill contains pink aplitic veins up to 10 cm across.

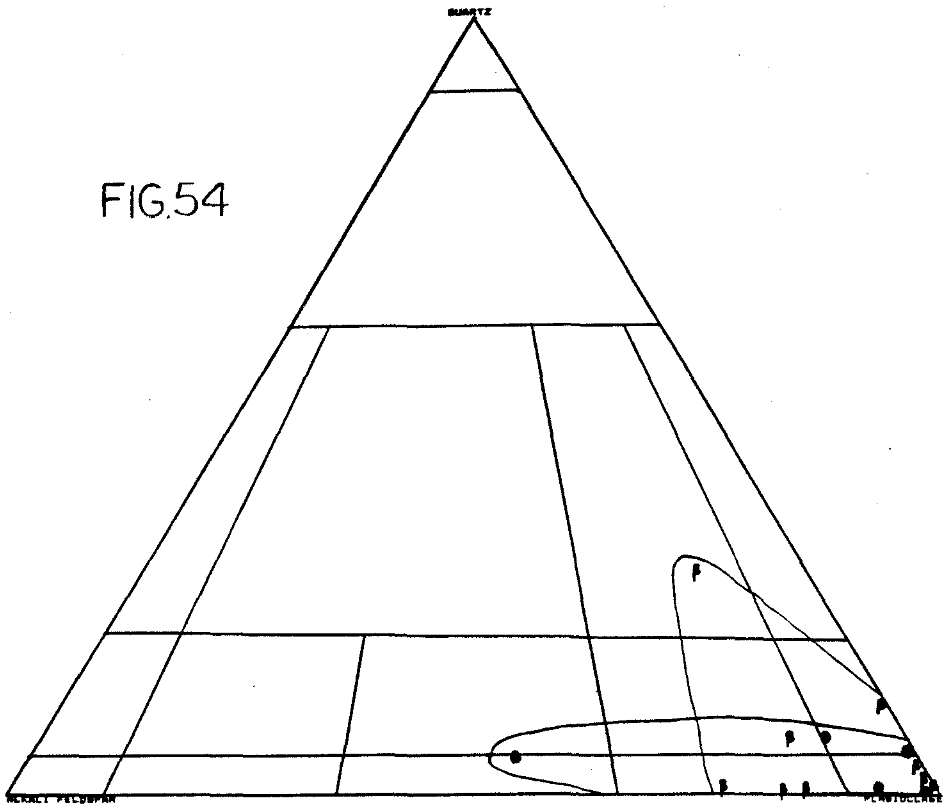
The last two bodies are both very thick (660 and 720 m maximum thicknesses) and are at least 8 to 10 km in width (the actual original width is difficult to determine because of the imbrication, but could be greater than this estimate). These larger bodies intrude the upper part of the Happy Creek Formation (member G - see the stratigraphic sections, fence and block diagrams for the Happy Creek, as the sills are included in that stratigraphic analysis). They were exposed and eroded before deposition of the overlying King Lear Formation (the basal King Lear overlies them nonconformably with subaerial weathering, and also contains clasts identical to the altered and fresh phases of the sills). Because of this relation, they must have been emplaced and hydrothermally altered at very shallow crustal levels with an overburden of only several hundred meters. Xenoliths of diorite are also abundantly present in the volcanics of member G of the Happy Creek, indicating overlap of andesite extrusion in that member and of this dioritic phase of sill intrusion.

Locally along the intrusive contacts igneous breccias are developed, with xenoliths of green altered andesite. There are baked contact zones within the intruded andesites. The sills do not appear to have generated abundant diking or apophyses, even internally. The contacts (where observed - they are usually obscured) are irregular on an outcrop scale, but megascopically are smooth and planar. On both the northern and southern fringes of the lower large sill, the body fingers out concordantly into the country rock. The fringes and contacts tend to be finer-grained. It is possible that these bodies are composite intrusives; there is a fair degree of internal heterogeneity.

The sills were all characterized as fine- to medium-grained diorites in the field (fig. 54), light to medium grey when fresh and medium to dark green when altered (the more common situation). The diorite was observed to contain plagioclase, hornblende and/or pyroxene. The texture varies from equigranular and phaneritic to subophitic and poikolitic, with hornblende or pyroxene plates enclosing the plagioclase. A restricted phase within the interior of the lower large sill, consisting of medium-grained quartz diorite, with plagioclase, clinopyroxene, biotite and minor quartz, is present at several localitites on DeLong Peak.

In thin section, the diorites contain oligoclase to andesine plagioclase, a variable mafic component (containing some combination of hornblende, hypersthene, and augite), and minor Fe-Ti oxides. A minority of the samples also have significant amounts of one or more of the following minerals: alkali feldspar (orthoclase), biotite and quartz. The mafic component can be composed of the following combinations: hornblende; opx and hornblende; opx; cpx; cpx and hornblende; or opx and cpx. In some cases the hornblende is replacing and rimming pyroxene, and chlorite is replacing and interleaved with biotite. Textures are subhedral or anhedral granular (and more rarely somewhat porphyritic), with a color index (pre-alteration) of 20 to 50. The crystallization sequence was plagioclase pyroxene hornblende biotite alkali feldspar quartz. The sills are primarily dioritic with excursions into the quartz dioritic, monzodioritic, quartz monzodioritic and granodioritic fields.

FIG.54



Petrographic modal analyses from the two sill suites of the Early Mesozoic Intrusive Suite intruding the eastern part of the Jackson Mountains, plotted on a Q-A-P diagram for the classification of plutonic rocks (after Streickeisen, 1973). Modal percentages were visually estimated. The numbers in front of the sills suite names gives the location on the map and the text section discussing that intrusive suite.

(10) Dioritic sills (plagioclase An values < 50)

(11) Gabbroic sills (plagioclase An values > 50)

Both the diorite sills and their country rock (andesite flows of the Happy Creek Formation) are hydrothermally metamorphosed to varying degrees. This event is assumed to be autothermal because of its spatial association with the sills; the aureole extends out several hundred meters or less from the sill contacts, and the great bulk of the larger sills has been affected. The metamorphism is characterized by chalky alteration of the plagioclase, replacement of the mafic phases by epidote and chlorite, and brittle fracturing and brecciation with epidote, magnetite and hornblendite veining. The magnetite mineralization is later than and cross-cuts the other veins. Accordingly, the diorites are normally green and heavily seamed with a network of veins, with igneous textures sometimes highly obscured. The gabbro sills intruding them lack and postdate this hydrothermal autometamorphism.

Alteration of the diorite in thin section is also commonplace and intense, with chlorite, epidote, calcite and Fe-Ti oxides replacing the hornblende, biotite and both pyroxenes. The plagioclase is highly replaced by very fine-grained sericite or can be brown and turbid. Chlorite and epidote veining is commonplace, and quartz and calcite veins are also seen. The quartz veins postdate the others. The volcanic country rocks are altered in a very similar fashion. Several samples of sill also have natrolite, a Na-rich zeolite, which in one case overprints and postdates a phase of deformation. This deformational phase in that sample is characterized by brecciation and brittle milling of the diorite in irregular pockets and seams with no through-going foliation, and the matrix to the breccia zone still looks igneous. This may be an explosive texture. Similar breccias were observed at scattered sites elsewhere in the sills; they may be parts of diatremes.

Actually, a more technically correct term for the sills might be "laccolith," a term implying forceful but generally concordant intrusion with uplift of the roof and a plano-convex to doubly convex lens shape (Corry, C.E., 1988). There is no clear distinction widely accepted between sills and laccoliths, though a ratio of diameter to thickness of 10 to 1 has been proposed (Billings, 1972). As the tabular and concordant bodies in this area are about 600 to 700 m in maximum thickness and have lateral extents of 5 to 10 km, these bodies are on the borderline. The sills (except the one near Deer Creek Peak) do largely overlap each other. I will continue to call them sills, with the caveat that the majority may actually be part of one or more shallow and very large Christmas-tree laccoliths (which would include the gabbro sills). The major one would be around the DeLong Peak area, and a second satellite system might exist at the tip of Trout Creek Spur. As a result of the imbrication, the original geometry is somewhat obscured, and no feeder system was found.

The lower larger sill has yielded a highly discordant and rather preliminary U/Pb age on extracted zircons of 170 to 175 m.y. (this is a tentative upper intercept - the lower intercept is probably in the Late Jurassic or Early Cretaceous).

(11) Gabbro Sills, Eastern Jackson Mountains

These sills overlap spatially with the diorite sills, but are distinguished in the field by their much coarser grain size, comparative lack of alteration, and generally more mafic nature. They also consistently crosscut and intrude the diorite phase, though they cannot be much younger because they are imbricated by thrusts plugged by the 162 ± 1 m.y. Trout Creek Spur monzodiorite pluton, and one sill is itself intruded by that body. Two large and several smaller sills are included in this suite. One of the larger occurs in the area in and east of DeLong Peak and has a maximum preserved thickness of 490 m, while the other is located at the tip of Trout Creek Spur and is at least 600 to 800 m thick. The Trout Creek Spur gabbro sill is particularly thick and stubby, and should properly be called a laccolith - it can be seen to definitely inflate and uplift the stratigraphy (including a diorite sill) in the roof along an S-shaped contact. The DeLong Peak intrusive is a sill proper, as its lateral extent is at least 7 km. Two other sills, only 20 to 30 m thick at most but 3.6 km long, intrude one of the diorite sills in the headwaters of Bottle Creek. They finger out into several thinner apophyses at their edges. A single gabbro dike, possibly part of the feeder system, was seen intruding the Bliss Canyon Formation near Burro Bills Spring. The lower stratigraphic position occupied by these sills despite their younger age, at a time when volcanism was still adding to the Happy Creek pile, might be due to their higher density (they are distinctly more mafic and calcic). Sills and laccoliths intrude along the horizon where the density contrast between country rock and magma is zero.

The gabbros are coarse-grained to very coarse-grained, greenish brown to brown in color, and have a color index of 30 to 80. They were observed in the field to contain plagioclase and a mafic phase, either pyroxene, biotite or hornblende. In places the texture is poikolitic, with large hornblende plates. In others there is a well-developed igneous foliation, expressed as the parallel arrangement of the plagioclase phenocrysts at a high angle to the margins. The texture is always highly crystalline and equigranular, though the plagioclases are the earliest phase to crystallize.

The petrographic examination of the gabbros indicates that they contain labradorite plagioclase, hypersthene, clinopyroxene (augite or diopside), hornblende, and minor biotite, alkali feldspar (microcline), quartz, and Fe-Ti oxides. The hornblende may rim the clinopyroxene. Textures in thin section were subhedral granular to subophitic, and the quartz and alkali feldspar can by myrmeketically intergrown. Some more-evolved phases, enriched in quartz, biotite and alkali feldspar (though the plagioclase is still calcic) and with trace apatite are also present in the sills. The crystallization sequence was plagioclase pyroxene hornblende Fe-Ti oxide biotite alkali feldspar quartz. Limited alteration of the pyroxenes, hornblende and biotite to chlorite has taken place, and in some cases the plagioclase is also turbid. In general, however, the textures are better preserved than in the diorite sills, though near the thrusts there can be some brittle deformation (i.e., kinking and fracturing of the plagioclase) and limited epidote veining. The lithology in the sills is gabbro to quartz gabbro to monzogabbro to monzonite (a slightly more potassic and more quartz-deficient trend than seen in the diorites).

The green hydrothermal alteration and veining characteristic of the diorite is missing, but some magnetite veins do crosscut the gabbros. There is no diking of the gabbro sills or of the country rock by related late-stage phases. The margins are fine-grained and dark green or bleached, and may have igneous breccias along them. The country rocks at the margins are slaty and indurated by contact metamorphism, and have been mineralized, over a zone 30 m in width. Some stretching of the country rock, and epidote veining, is also present along some contacts. Miarolitic cavities, indicative of shallow intrusion, are present in one of the stratigraphically lower sills.

Near the Trout Creek Spur pluton, granitic and pegmatitic dikes intrude the gabbro. The monzodiorite pluton is chilled against and cuts the gabbro.

The gabbro sills must be younger than 170 to 175 m.y. (the age of the diorite sills) and older than 162 $\pm$ 1 m.y. (the age of the cross-cutting pluton).

(12) Bottle Creek Dike Suite

Intruding the imbricated Bliss Canyon and Happy Creek Formations on the eastern flanks of the range is a suite of tonalitic dikes. These bodies are fine-grained and white to very light grey, but are quite distinct from the rhyolite dikes of the King Lear Formation (member L) in the same area as the tonalites are phaneritic and crystalline in texture. The dikes are also found in the same structural and stratigraphic setting in the Jungo Hills, south of the mapped area. Dike intrusion overlapped thrusting - some are deformed and foliated, and others are definitely stitch fault planes and crosscut the deformational fabrics. This relationship also exists in the Jungo Hills (R. Speed, oral communication, 1984).

These intrusives contain quartz, plagioclase, hornblende and biotite. The plagioclase has pronounced oscillatory zoning, is sodic (albite to andesine), and is graphically intergrown with the quartz. The texture is anhedral granular. Where the dikes predate deformation there is pervasive fracturing, kinking of the plagioclase and undulatory straining of the quartz. Petrographically they are tonalites.

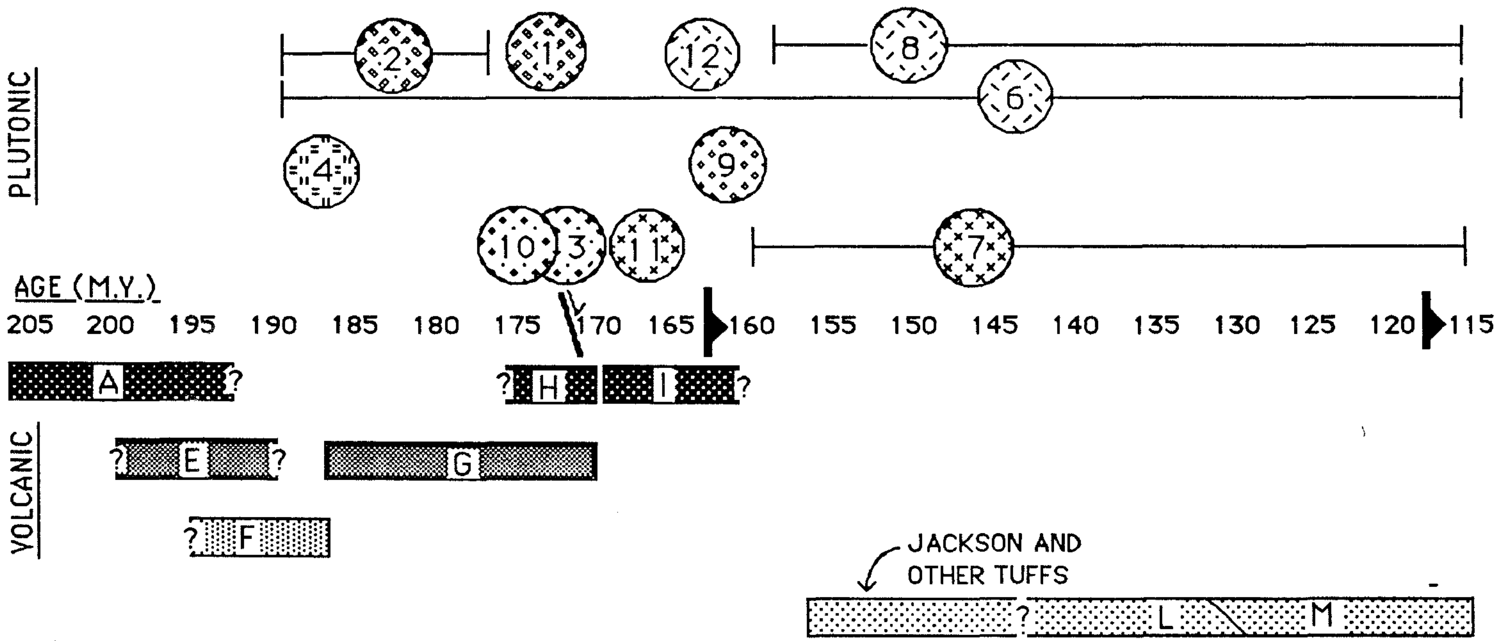
Zircons have been extracted from several dikes (one pre- and one post-thrusting), but have not yet been dated. The first and major phase of thrusting on the eastern system took place in the period 160 to 170 m.y. ago (and probably in the latter part of that period) based on other constraints, so the dikes must overlap this time period (see chapter 8).

Setting and Development of the Early Mesozoic Intrusive Suite

There is no obvious systematic trend seen in the petrology of the intrusives alone as a function of time or of geography. Plutons intruded later can be either more mafic or less than those intruded earlier. The sills are all highly mafic, deficient in quartz and somewhat less so in K. The gabbros are more calcic and mafic than the diorites. The stocks vary greatly, forming a general trend from diorite to granite, but are also rather K-poor. Each stock is internally fairly uniform (with the exception of the Navajo Peak body). Those dikes that can be seen to be comagmatic with certain of the stocks (all but the Bottle Creek suite) tend to be slightly more silicic and potassic than the parent body, but otherwise similar.

The volcanic record in the Happy Creek and King Lear Formations, which is a much more complete record of magmatic activity in the area, does show distinct temporal trends (see fig. 55). Volcanism in the Happy Creek started out as basaltic andesite in member A, to andesitic in E, to andesitic and dacitic in F, back to andesitic in G and to basaltic andesite in H. In the King Lear in member I, volcanism was basaltic to andesitic, then there was a hiatus of unknown duration and silicic tuffs were deposited. In member L some time later volcanism was rhyolitic, and slightly later again in member M dacitic to rhyolitic.

The record preserved in the plutonic bodies is much patchier, but when compared temporally with the volcanic record some correlation is evident. The granitic to granodioritic Deer Creek Peak and Harrison Grove stocks on the northwestern part of the range overlap in time with the andesitic to dacitic volcanism in member F, though the slightly less evolved Happy Creek stock was intruded during a phase of much more mafic volcanism (members G and H). The dioritic Parrot Peak and monzodioritic Trout Creek Spur stocks and both sill complexes were also intruded during that more mafic episode, and so do correlate fairly well. The ages of the tonalitic Bliss Canyon and Trout Creek stocks are not well tied down enough for a similar comparison, though they might be related to the rhyolitic and dacitic activity in the Jackson Tuff and members L and M. In the Pueblo Range to the north, a major silicic volcanic episode took place in the Late Jurassic (Roback, 1988); this may be the same event. The tonalitic sills around the Jungo Hills and Bottle Creek might correlate to the Jackson Tuff, but are too old for the later silicic phases in members L and M of the King Lear. The gabbro at Navajo Peak does not appear to have any equivalent in the volcanic record. Generally, the intrusive and extrusive phases are concluded to have been



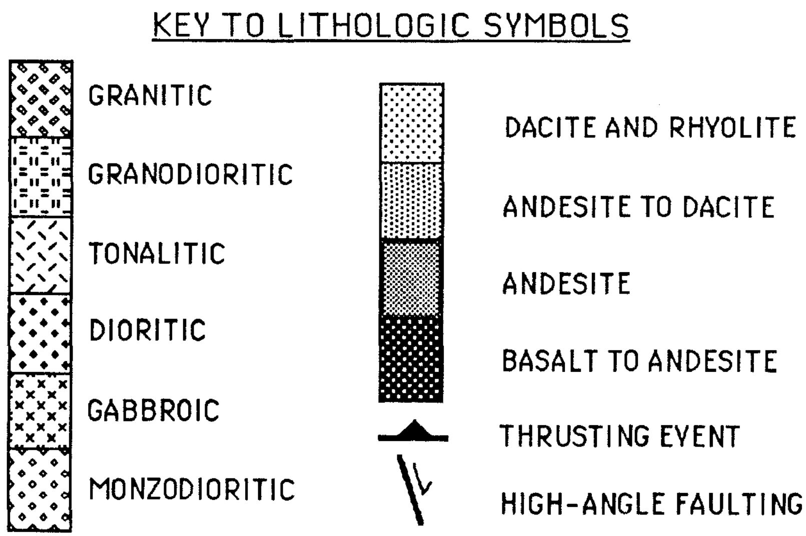


FIG.55

The chronology of the volcanic and plutonic rocks of the Early Mesozoic Intrusive Suite and Happy Creek and King Lear Formations, Jackson Mountains, NW Nevada.

The letters give the member, and the numbers give the plutonic body, as discussed in the text in the appropriate chapters for these units.

comagmatic. This is fully supported by cross-cutting field relations.

There is some evidence, then, for the existence of evolutionary trends in the Mesozoic igneous activity in the Jackson Mountains. From about 205 to 185 m.y. ago, igneous activity became more silicic and sodic. At about 185 to 180 m.y., a trend back to more mafic activity began, culminating around 165 to 170 m.y. After a minor hiatus, highly silicic igneous activity began again, from some time at or soon after 160 m.y. through to 115 m.y. The extent of continuity in this last phase is not clear yet, and there appears to be a major hiatus in volcanism in the early Early Cretaceous.

All of the intrusives in the Early Mesozoic Intrusive Suite were intruded in the epizonal (shallow crustal) zone of Buddington (1959). They are characterized by sharp and discordant margins with stoping, brittle diking and veining, chill zones and only minor contact metamorphism, and have simple internal plans. With stocks of known age the stratigraphic overburden at the time of intrusion can be determined, and is always less than 3 to 4 km. The bodies are not hypabyssal (with the exception of a few of the dikes), as they are fine- to very coarse-grained, crystalline to phaneritic, and lack a aphanitic or glassy matrix; their size and resultant slow cooling could account for this.

Other aspects of the evolution of the area are indicated by the morphology of the intrusives. Over the same time period in the Middle Jurassic, sills were being intruded into the stratigraphy in the eastern part of the range, and stocks and related dike swarms were characterizing the western side. No differences existed in the strata into which they were intruded that might explain this. It should also be noted that on the eastern side, thrusting took place very soon after the sills were intruded and was then stitched by stocks. And on the western side, high-angle faulting (with the structures trending east-west, and both strike-slip and dip-slip components) overlapped in time the intrusion of the stocks and dikes. The conclusion is that the mode of intrusion for these similar magmatic phases was caused to differ by the local stress state. Stocks and dikes are known to intrude preferentially in areas of neutral to tensile stress. Sills are injected in areas of more compressional stress conditions (Jaeger, 1969; MacBirney, 1979). However, within complex wrench zones a variety of structural regimes and hence intrusive geometries can exist (Sanderson and Marchini, 1984).

The orientation of the principal stresses were coaxial in the different regions (dominated by sills on the east, and stocks and dikes on the west), though relative magnitudes varied. More specifically, on the east the sills were injected in a region with a local stress state of Ú$_1$ horizontal and with an azimuth of about S70E, Ú$_3$ vertical and Ú$_2$ at right angles to both. The earlier to synchronous high-angle faults in the central and western portion of the range are inferred to have been dominantly conjugate faults related to wrenching, with oblique-slip offset (see chapter eight for a detailed analysis). The intrusion of elliptical stocks in this area, and their elongation direction, is compatible with that stress state. The dike swarm may have been caused by dilational jointing around the stocks, but does not appear to have a systematic relationship to the stress state.

The later stocks (Late Jurassic and Early Cretaceous) in the Trout Creek Spur area and at Navajo Peak, cutting the sills and the thrusts, document that that the regional compressive stress state (due to the eastern thrust system) had been replaced by a more neutral one. The later western thrust system (cutting the stocks) has synorogenic sills intruded within the King Lear Formation (actually a laccolith in member M), once again indicating a regional compressional stress state.

CHAPTER SEVEN:

METAMORPHISM IN THE JACKSON MOUNTAINS, NW NEVADA

The development of metamorphic minerals and textures in the Mesozoic and Late Paleozoic strata of the Jackson Mountains is generally restricted to contact metamorphism around some of the larger plutons of the Early Mesozoic Intrusive Suite, and to more regionally extensive dynamothermal metamorphism in the structurally higher thrust sheets in western thrust systems. Substantial hydrothermal alteration also took place along the family of high-angle faults, and around the diorite sills.

This chapter is divided up into sections on: (i) the contact and hydrothermal metamorphism around the intrusive bodies; (ii) hydrothermal alteration along the high-angle faults (D1); (iii) metamorphism associated with the eastern thrust system (D2); (iv) metamorphism in the western thrust system (D3); and (v) hydrothermal alteration along the Basin-Range normal faults (D4). Each section analyzes the metamorphic mineral assemblages, the textures, the cross-cutting relationships of the metamorphic episodes with other phases, and the evidence for timing and cause of the metamorphism. Turner (1981) was used for interpretation and assignment of metamorphic facies.

Contact Metamorphism Around Intrusions

Baking of the immediately adjacent country rock is a feature of most of the intrusions in the range. Only a few develop significant contact aureoles, however. This latter group includes the gabbro sill at the tip of Trout Creek Spur, the larger diorite sills, the Navajo Peak gabbroic stock, the Harrison Grove granodiorite stock, and the Parrot Peak diorite. Apparently, the more mafic and deeper intrusives (though all are epizonal) are more likely to have developed contact aureoles. Higher temperatures of mafic magmas, and slower cooling at greater depths, are inferred to cause this relationship. For a discussion of the petrology and intrusive relations of these igneous bodies, see the chapter six on the Early Mesozoic Intrusive Suite.

Around the Navajo Peak zoned stock (with a gabbro core to granite shell), the King Lear Formation (member I) has undergone contact metamorphism. This is characterized by extensive replacement by epidote, chlorite, calcite, and quartz, with recrystallization of plagioclase feldspar. The original detrital textures are obscured, with ghosts of pebble clasts the only remaining original feature. No preferred fabric was evident in the replacement assemblage. The altered zone is only tens of meters wide, and also exhibits Cu-mineralization, with azurite and malachite. The metamorphic grade indicated by the assemblage observed is low and in the albite-epidote hornfels zone (Turner, 1981), because actinolite is lacking. Low to moderate temperatures and low pressures are inferred; the latter is supported by the nature of the intrusive (epizonal) and the inferred stratigraphic overburden (less than several km).

The Trout Creek Spur gabbro laccolith (thicker and relatively less wide than a sill) has a contact zone about 30 m in width. The inner 12 m is greenish and slaty, with a related cleavage parallel to the contact.

The contact is locally Cu-minerlized. The outer 18 m is reddened, lacks the Cu-mineral-bearing veins, and is slaty. The depositional textures in the Bliss Canyon Formation (the country rock) are still discernable, though overprinted to some extent and even ductilely stretched during inflation by the sill. This last feature is seen in a deformed siliceous and calcareous metaconglomerate lithology. The carbonate beds are pink and dolomitized, and epidote veins cross-cut the wall rock. This is also a low to moderate temperature, low pressure contact aureole, probably also albite-epidote hornfels.

The Harrison Grove granodiorite and Parrot Peak diorite stocks are located in the same area as, and both contribute to and are cut by, the Mary Sloan Creek dike swarm. A high heat flow in this area is evident from the mineral assemblages and textures in the Happy Creek and Bliss Canyon Formations. The metamorphism is inferred to be caused by all three units. The dike swarm (despite the size of the dikes) would have had a disproportionately high heat flux in its area if the dikes served as conduits and feeders, with magma flowing through them for extended periods.

In the Happy Creek Formation in this area, the volcanic flows are overprinted by the replacive growth of plagioclase, chlorite, epidote, elbaite tourmaline, quartz, opaque phases, and calcite. More rarely, hornblende also occurs as a metamorphic mineral. The tourmaline replaces the chlorite, and the mafic volcanic phases have gone to chorite, epidote, calcite and opaque phases. The plagioclase grains are recrystallized, and the groundmass overprinted by all of the metamorphic assemblage listed above. Original volcanic textures are usually still preserved. This assemblage in a mafic lithology is indicative of the lower albite-epidote hornfels facies, except where hornblende is present and the metamorphic facies is higher hornblende hornfels.

Tuffaceous cherts in the Bliss Canyon are heavily recrystallized, with epidote, chlorite and calcite overprinting the detrital features, while the argillites become slaty and the carbonates are completely recrystallized to marble. Tremolite, wollastonite and diopside are present in one very proximal and highly metamorphosed sample - a calcareous meta-conglomerate - together with quartz, calcite, and epidote. This particular assemblage is typical of a $CaCO_3$-SiO_2-rich lithology in the inner contact zone of the hornblende-hornfels facies.

This replacement is partial and patchy and lessens in degree away from the intrusives. At its culmination near the Harrison Grove granodiorite, the textures are medium-grained, and granoblastic. No preferred metamorphic fabric exists - the minerals are randomly oriented - so the event was not dynamothermal. The nature of the assemblage, lack of a preferred fabric, and spatial relationship to the intrusives indicates that the event is indeed a contact metamorphic aureole. At least one later foliation, with kinking and fracturing of the minerals, does cross-cut the metamorphic textures. These late foliations are characterized by chlorite growth and brittle textures, and in places are a spaced fracture cleavage (early phase - D1) to slaty cleavage (later phase - D3).

The diorite sills east of DeLong Peak in the Trout Creek Spur also have associated alteration and replacement. This episode is also characterized by extensive chlorite and epidote veining. Brittle brecciation and veining is a very common secondary texture in the diorites and their immediately adjacent wall rock. Epidote, chlorite, quartz, plagioclase, calcite, and an opaque phase are the normal metamorphic assemblage around and in these intrusives. Several samples have metamorphic hornblende. Some of the cleaner and less overprinted samples of both diorite and andesite also have natrolite, a Na-rich zeolite. Chlorite (which often replaces the mafic phases) is being replaced by biotite in some cases. The andesite flows of the Happy Creek Formation intruded by the diorite share the same metamorphic assemblage and textures; these features disappear over a 100 m or so span away from the intrusive contacts. The contact metamorphic facies are once again the albite-epidote to hornblende hornfels, indicating conditions of low to moderate temperature and very low pressure. The extensive brittle network veining, the autometamorphism of the diorite, and the width of the contact aureole indicate that hydrothermal activity was an important part of the event. The metamorphic textures are in places deformed and overprinted by both the high-angle fault (D1) and thrust fault (D2) phases of deformation.

Where the hydrothermal alteration in and around the diorites is not pronounced, a zeolite-grade background level of very low-grade regional or burial metamorphism is inferred to have existed. This is not at all uncommon in undeformed thick volcanic and sedimentary provinces. Alternatively, it could be the preliminary stage of a regional event.

In all the areas discussed above, the metamorphism is inferred to be caused by the nearby emplacement of an intrusive body. Hence the formation of the contact aureole will have been at the same time, or very soon after, that intrusion (most intrusions have some age constraints).

Local Metamorphism Associated With D1 - High-Angle Faulting

The generation of structures assigned to D1 are high-angle faults, trending generally east-west and with both dip-slip and strike-slip components. The faults were active during deposition of the uppermost Happy Creek (members G and H) and lower King Lear (members I and J) Formations, and were reactivated later as tear faults during thrusting.

During the first incarnation of these structures, extensive hydrothermal activity took place along the fault zones, with associated alteration and mineralization. This event is characterized by Fe and SiO_2 enrichment of the host rock along the fault planes and in the adjacent graben-fill. Red jasper, magnetite and both specular and earthy hematite are diagnostic, with epidote nodules as well. Extreme replacement of the matrix and mafics in the host volcanic rocks by hematite, and extreme sericitization of the plagioclase are diagnostic features. The iron minerals and jasper are often banded. Replacement by calcite, quartz, chlorite and sphene also occurred. Much of the mining in this area was for these particular deposits, as at the Iron King, Delong and Redbird claims (Shawe et al., 1962).

The replacement predates the deposition of the King Lear, as that unit is not affected and also contains clasts of lithologies altered and replaced by this particular metamorphic episode (including sedimentary magnetite breccias). The mineralization therefore took place early in the first stage of activity. The phase also predates thrusting, as the deposits are offset. It also postdates the intrusion and autometamorphism of the diorite sills, as magnetite veins cross-cut that event. Member H of the Happy Creek was synchronous, because it contains Fe-mineralized debris and is itself overprinted. The fault planes acted as localized conduits for hydrothermal fluids, probably with a heat source in subvolcanic intrusives (such as the epizonal stocks of the Early Mesozoic Intrusive Suite). The high-angle fault conduits were also definitely contemporaneous, as the wall rock was hematized before and after cataclastic fault brecciation (i.e., both the country rock and the fault breccias are mineralized, and mineralized zones brecciated). The nature of the alteration is that of an epithermal deposit, less than 1500 m deep (Evans, 1980), which agrees with the inferred setting.

Regional Metamorphism and D2 - West-Vergent Thrusting

The thrust system on the eastern side of the range is characterized by very low-grade dynamothermal metamorphism, most progressed in the structurally higher plates containing the Bliss Canyon Formation. Here, slaty cleavage is sometimes weakly developed, particularly in the more argillitic lithologies in the axial areas of tight folds and right next to thrusts. No significant metamorphic replacement and little recrystallization was seen in thin section. The coarser clastics, though stylolitized and with suturing along grain-grain contacts, are not metamorphosed noticeably, and even the carbonates preserve depositional textures. In the Happy Creek Formation, some zeolites are present in the volcanic facies; these and the local slaty cleavage are the evidence for any metamorphism at all. This orogenic zone did not reach even greenschist grade, but is zeolite grade or lower.

Regional Metamorphism and D3 - East-Vergent Thrusting

The thrust belt along the west side of the range contains the highest metamorphic grades achieved in the range, particularly in the area around the mouth of Jackson Creek. The metamorphic grade declines steadily and rapidly to the south within the thrust packages, and the footwall to the thrust system is not metamorphosed (except where an earlier phase of contact metamorphism had already taken place, around the Harrison Grove and Parrot Peak stocks).

Within the northern nappes, the textures are phyllitic to highly schistose. The mafic lithologies from the Happy Creek have a well-developed foliated and lineated mineral fabric. The assemblage contains biotite (replacing chlorite) or muscovite, hornblende, an opaque phase, quartz and plagioclase. The latter is extensively recrystallized-relict deformed plagioclase phenocrysts with sericitized cores, and granulated albitic edges are present. Sillimanite is present in small amounts in one sample. The fabric is fresh and unstrained, and has experienced a very high degree of dynamothermal recrystallization - the only original features discernable are the inherited plagioclase phenocrysts. In places the hornblende randomly overprints the foliation, indicating static post-tectonic high temperatures.

The Bliss Canyon Formation lithologies in the northern parts of the thrust belt are equally metamorphosed. The tuffaceous meta-chert is highly recrystallized and crenulated with a weak axial planar cleavage, and muscovite and biotite flakes are parallel to this foliation. The carbonates are marble with calcite and diopside, and were slightly sheared following recrystallization. The siliceous meta-conglomerates are recrystallized to quartz, albite and muscovite. The meta-arenites have been replaced by actinolite, quartz, biotite/chlorite (intergrown along cleavage planes), opaques and muscovite, and are well-foliated and lineated. This facies also displayed a late static heating, with semi-random actinolite overgrowth. Chlorite seems to dominate over biotite as lower grades are seen more to the south. Where detrital features are preserved (such as pebble ghosts), they are highly stretched. All of these assemblages are in equilibrium.

Further south, within the Bliss Canyon and McGill Canyon Formations, the metamorphic grade is contrastingly very low. In the volcanic and epiclastic facies, plagioclase is sericitized but not recystallized, the matrix is replaced by chlorite, calcite, white mica and quartz, and the mafic phases have gone to chlorite, calcite, epidote and an opaque phase. The carbonates are not recrystallized. The finer-grained facies are argillite- to slate-grade.

In the northernmost part of the thrust zone, the metamorphism reaches the biotite zone of the greenschist facies. This is indicated by the presence of albite + biotite + chlorite $\pm$ quartz + hornblende or actinolite in the equilibrium assemblages. The sillimanite in a pelitic schist, and the diopside in the metacarbonate, indicate that locally amphibolite-grade conditions existed. To the south, grade decreases to chlorite zone (biotite gives way to chlorite). Further south yet, the grade is quite low, as indicated by the presence of chlorite, calcite, white mica, quartz and epidote. The lowest metamorphic grade achieved in the south is uncertain, but is probably either zeolite or prehnite-pumpellyite. The trend from north to south within the thrust belt is from moderate/high to very low temperatures, and low pressures throughout.

This metamorphic event was dynamothermal. The heat source is uncertain - the amount of shortening and telescoping in the thrust zone does not seem sufficient for the high temperatures observed. Also, the highest metamorphic grades are observed in areas with the least amount of stratigraphic uplift and deformation. The stratigraphically lowest unit, the McGill Canyon Formation, is also the southernmost, and the least metamorphosed. I conclude that the heating was largely caused by nearby batholithic intrusive activity, and that the thrusting took place soon after intrusion while the rocks were quite warm enough for the metamorphic assemblages observed. After deformation, they were still hot enough for static overprinting. The dynamothermal event is inferred to have taken place in the Aptian (or at least the late Early Cretaceous). This is based on sub-solidus deformation in a sill and soft-sediment deformation in lake deposits of that age. Though not observed in the Jackson Mountains, batholithic activity of an appropriate age is widespread in the Pine Forest Range to the west and north across the Black Rock Desert.

Basin-Range Normal Faulting - D4

The Basin and Range event is expressed in the Jackson Mountains as the reactivation of pre-existing thrusts as low-angle normal faults. In the northeastern part of the range in the Bottle Creek district, this activity was accompanied by hydrothermal activity along and near the reactivated range-front faults. These systems were not studied in any detail, but are inferred to have been epithermal. Mining of the associated hydrothermal quicksilver deposits has taken place to a considerable extent (Willden, 1963).

CHAPTER EIGHT:

STRUCTURAL GEOLOGY OF THE JACKSON MOUNTAINS, NW NEVADA

The Jackson Mountains are one of the north-south elongate ranges in the northwestern portion of the Basin and Range geomorphic province. To either side lie deep, alluvium-filled playa basins whose flat surfaces lie at about 4000 feet in elevation. The range has very abrupt flanks, and rises to over 9000 feet. The central part of the range is more subdued, so that there are two parallel ridges with a central depression that make up the topography; the eastern (and smaller) ridge is the Trout Creek Spur. This central depression is caused by the more recessive nature of the underlying King Lear Formation. The more resistant flanking ridges are composed primarily of the Happy Creek Formation and of stocks of the Early Mesozoic Intrusive Suite.

Detailed field mapping has revealed that the geologic evolution of the Jackson Mountains has been characterized by four separate and distinct structural deformation episodes (see plate I and II, and fig. 77, 78, 79 in chapter 9). Preceding these were various soft-sediment deformation events of limited extent in the different stratigraphic units (particularly in the McGill Canyon and Bliss Canyon Formations), and lumped together as phase D0. The first and earliest tectonic phase, D1, is characterized by activity on a family of dominantly east-west trending high-angle faults in the late Middle Jurassic. This system is best developed in the central depressed portion of the mountains in the Jackson Creek area, but overlaps to the east and west with domains dominated by phases D2 and D3 on the ridges to either side (particularly in the exposures of the McGill Canyon Formation).

Phase D2 consists dominantly of imbricate west-vergent thrust faults, with an average trend of 20° (azimuths are used throughout this chapter, and the dip is 90° clockwise of the azimuth). This generation is found dominantly along eastern flank of the range. Some early east-vergent thrusting around King Lear Peak on the western flank of the range is also assigned to this phase. During this event, some structures from D1 were reactivated as tear and oblique-slip faults. Two distinct stages may be present within phase D2, with distinct vergence directions (at about 240° and 300°).

Phase D3 is represented by a second episode of thrusting, both west and east-vergent and overlapping geographically with D2. These structures are present along both the western flank and the east side of the central depression. Faults from D1 were reactivated in the same fashion as for D2 during this later episode. In addition, the frontal portion of the D2 thrust system in the eastern thrust system was re-utilized, as was the back portion of the western thrust system. Both thrust systems had deformation during both D2 and D3, and locally truncate D1 phase structures.

Basin-Range normal faulting and tilting has caused the present topographic expression of the range. Features from both D2 and D3 (and possibly D1) were reactivated during this last phase, D4.

Phases D1, D2 and D3 have complex cross-cutting relationships with the Mesozoic strata of the Jackson Mountains (including the Happy Creek and King Lear Formations and the Early Mesozoic Intrusive Suite). Older units (the McGill Canyon and Bliss Canyon Formations) are also deformed by these phases. Synorogenic stratigraphic relationships are evident in these Mesozoic units, and synsedimentary activity on all three phases is also witnessed by the rock record. The complexly cross-cutting features between the sedimentary, volcanic, intrusive and deformational events enable the compilation of a detailed geologic history for the range. The timing of activity of the structural episodes can thus be determined with some accuracy. Much of this latter information has already been discussed in the preceding chapters on the relevant units, and so will only be summarized briefly here.

The basic structural architecture then consists of: (i) East-dipping thrusts making up the eastern range and imbricating dominantly the Bliss Canyon, Happy Creek and King Lear Formations; (ii) a homoclinally east-dipping and high-angle faulted section of King Lear and Happy Creek units in the central depression and east side of the western ridge; and (iii) west-dipping thrusts on the west side of the western range, stacking the McGill Canyon, Bliss Canyon, Happy Creek and King Lear Formations.

Because of the extent to which some structures have had several generations of activity on them and the corresponding difficulty of assigning a particular structural element to a certain phase, it is not practical to analyze the various phases in order of occurrence, from D1 to D4, as would be customary. Only the soft-sediment deformational events (D0) will be discussed in this independant fashion. Instead, as the individual structural domains are quite distinct in geometry, I will divide the structural geology discussion up into geometric domains dominated by a particular structural orientation, and discuss each domain independantly.

Thus the structural domains to be discussed are: (1) The high-angle fault family and related deformation in the generally east-dipping homocline of the central depression of the range (initiated during phase D1, and reutilized in D2 and D3); (2) the western-vergent thrust system along the eastern portion of the range (active during D2, with frontal portions reactivated in D3); (3) the easterly-vergent thrust system along the western flank of the range (with elements from D2 and D3 and also overprinting D1 generation features); and (4) Basin-Range normal faulting, on the extreme east and west flanks.

For each domain, the discussion will cover: (i) description of the characteristic structural elements, including fault geometry and orientation, sense of offset, style and orientation of folding, associated fabrics, dynamothermal metamorphism (if any), synsedimentary relationships, and cross-cutting relationships; (ii) interpretation of the kinematics, state of stress, timing and relationships of the various phases, and a general summary of structural development within each domain. The stereonet plotting was done using the Stereonet (version 3.6) program for the Macintosh by R. Allmendinger.

At the conclusion of the chapter, I will synthesize the history and character of deformation in the Jackson Mountains, the evolution and implications of the structural phases, and the influence exerted by these tectonic phases on the nature of the depositional environments and igneous activity.

The accompanying geologic map and cross-sections will provide details for this discussion. The structural terminology and classification schemes used in the chapter come from Billings (1972), Hobbs et al. (1976), and Ramsay (1983; 1987).

Soft-Sediment Deformation (D0)

Soft-sediment deformation is present in the finer-grained sedimentary facies of both the McGill Canyon and Bliss Canyon Formations. In the McGill Canyon, D0 is commonly associated with the emplacement of blocks of the olistostromal carbonate member, as already mentioned in chapter 2. In the Bliss Canyon, such mega-slide features are lacking. The soft-sediment deformation in the Bliss Canyon Formation is present as flame structures and as stratigraphically-limited zones of isoclinal to rootless ductile folds in the argillite successions. These features differ in style from the orogenic structures, are cut by the D2 cleavage, and have no systematic orientation or distribution. Similar characteristics are found in the McGill Canyon Formation independant of the slide blocks. Here, one finds broken formation (with no exotic blocks) of ductilely-sheared and folded argillite with brittly boudinaged fragments of arenite beds and volcanic flows. These zones also seem to be stratigraphically distinct. In both formations, these soft-sediment textures are inferred to be pre-tectonic and to have been caused by submarine mass-slumping of unconsolidated sediments soon after deposition.

Peperitic textures due to phreatic deformation are also found at several levels in the Happy Creek and King Lear Formations, related to the emplacement of volcanic flows and sills. In members B and F of the

Happy Creek and member I of the King Lear (and at the very top of the Bliss Canyon Formation), volcanic flows moving over wet sediments caused the ductile shearing to explosive intermixture of the flows and underlying sedimentary facies. The sills of member M of the King Lear also caused ductile intermixture and explosive peperitic textures in the sedimentary country rocks. See the appropriate chapters for details. These features are pre-tectonic, and are of an age with their sedimentary matrices.

High-Angle Faulting Domain

This system of faults is best expressed in the central depression in the middle portion of the range, within the Happy Creek and King Lear Formations. Elements are, in addition, found to the east and west within the younger thrust-dominated domains on either side.

Description

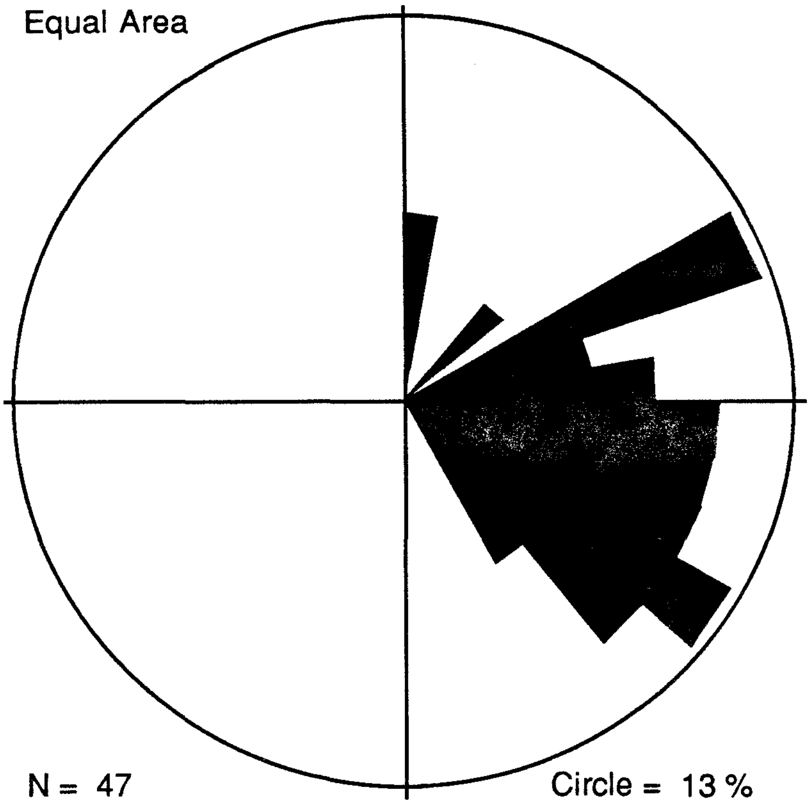
Geometry The high-angle faults are very steep and trend 45° on either side of an azimuth of 100°, with two conjugate concentrations at about 70° and 130° (see fig. 56). The latter orientation is dominant over the former. This distribution is not as clear on the rose diagram displaying fault trend azimuths as it is in the field, as some of the smaller and less significant structures have trends from 80° to 110° and are synthetic to the dominant set. Dips are between 70° and 90° to the north and south. Fault traces can be a few tens of meters to as much as ten kilometers in length. The faults are either overlain by younger King Lear Formation sediments, die out along strike, or are bounded by thrust faults. In the latter case, they can serve as lateral bounding surfaces to the thrusts, with differential movement and offset on either side, or can be cut and carried by thrust systems.

Along the high-angle faults themselves, cataclastic breccias are common in zones up to 1 m or so across. Slickensides with dip-slip orientations are present in some cases. The larger faults with the most offset (such as the Jackson Creek structure, or that in the Harrison Grove stock) have wider sheared zones up to several meters across, with a powdery fault gouge evolved. The smallest faults with the least offset are simply planar surfaces with some fracturing, and grade down through the shear cleavage into the spaced fracture cleavage.

<u>Offset</u> Offset is generally proportional to the fault length. A paucity of piercing points means that it is not always possible to distinguish dip-slip from strike-slip offset. Most faults have net displacement that is either south-side up and/or sinistral (southeast set), or north-side up and/or dextral (northeast set). Often it is not possible to tell the exact sense of offset during the initial phase on a particular strand, but where possible to tell, it is oblique. Some of the larger structures are seen to have both south-side up and sinistral in the initial phase, or, as in the case of the Jackson Creek fault, both north-side up and dextral offset. There does not appear to be any systematic relationship of offset (in the later phases) to fault orientation overall. Any fault trend is as likely to have net displacement south-side up or sinistral as it is to be north-side down or dextral because of the complex multiphase history, and adjacent and parallel faults may even have opposing senses.

However, where the sense of offset during the initial oblique-slip

FIG. 56



HIGH-ANGLE FAULT TRENDS

THE HIGH-ANGLE FAULTS ARE TYPICALLY NEAR VERTICAL. THE TREND OF THE FAULTS THEREFORE GIVE A GOOD PICTURE OF THE BEHAVIOUR OF THE FAULT SYSTEM. NOTE MAXIMA AT ABOUT 70 AND 125. THESE FAULTS HAVE BOTH DIP-SLIP AND STRIKE-SLIP HISTORY. THEY WERE INITIATED DURING PHASE D1 AND REACTIVATED AS TEAR FAULTS DURING BOTH D2 AND D3 PHASE THRUSTING.

phase of D1 is known, faults with azimuths of 100° or more (there are five) are consistently south side-down, and those trending less than 100° (two) are north-side down. The strike-slip sense during that early dominantly dip-slip phase is sinistral for the former set, and dextral for the latter.

Apparent strike-slip offset varies from a few meters to a few hundred meters typically, up to 2.6 km across the large fault in lower Jackson Creek and several km on the fault north of King Lear Peak. In the case of the Jackson Creek fault, the actual dip-slip offset was on the order of 1 km (south-side down, and seen in the stratigraphy of the King Lear). This was followed by later dextral strike-slip reactivation in the upper plate of a D3 thrust. In the headwaters of the north fork of Jackson Creek, the faults were dip-slip with several hundred meters of offset. Just north of King Lear Peak, the fault there had at least 100 m of dip-slip offset (south side down) and later very extensive sinistral strike-slip activity of a minimum of several kilometers during D2 thrusting. In the area just south of the Iron King Mine on the east side of the range, another fault had about 100 meters of dip-slip activity (south side up) but no later strike-slip history.

In most cases, offset is inferred to have been oblique-slip. Along some of the larger faults, early oblique-slip offset was followed later by strike-slip reactivation.

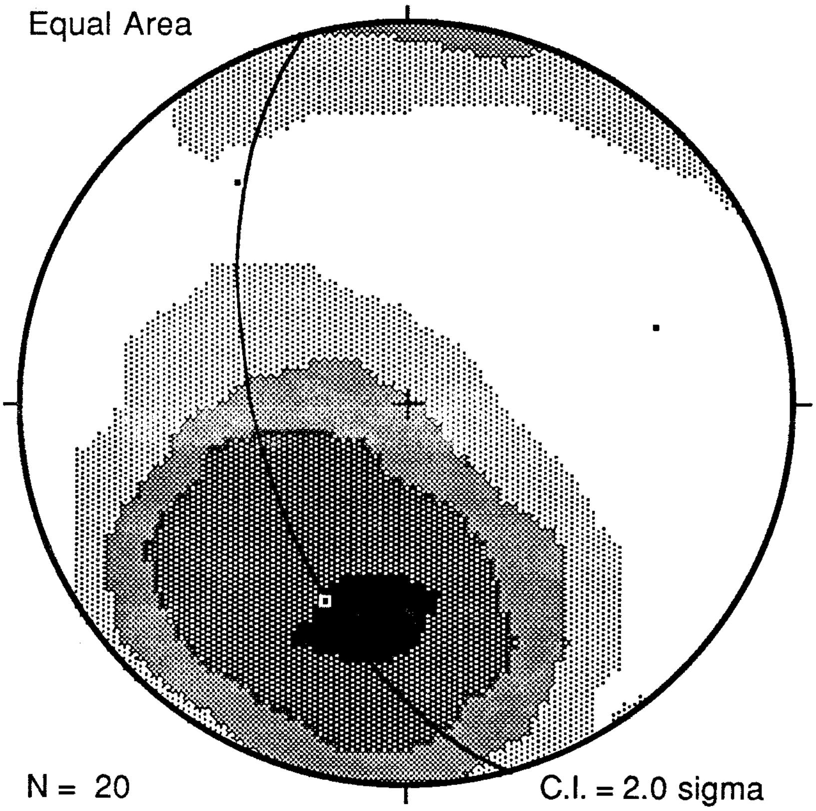
<u>Folding</u> Some steep warping and draping of bedding in the King Lear and Happy Creek Formations across the high-angle faults is observed on two of the structures, generally with south side up (see fig. 57). A monoclinal flexure at the surface during sedimentation (based on thickening across

the fault above the tilted strata) is inferred for the fault just south of the Iron King Mine. The axis is parallel to the high-angle faults, and the folds have amplitudes of a hundred meters or more.

Fabrics A pronounced spaced fracture cleavage to joint set (fig. 58) is present within a few hundred meters of some of the more prominent faults, affecting strata of the Bliss Canyon, Happy Creek, King Lear Formations (and, as will be mentioned in the discussion on the western thrust domain, the McGill Canyon Formation). The best expression of this system is in the drainage of Jackson Creek. Which phase of activity (dip-slip or strike-slip) these fabrics are associated with is not certain, though at least some development during the later strike-slip phase is indicated by the involvement of parts of the King Lear Formation post-dating the dip-slip phase. Some development of the foliation was also early, as dikes of the Early Mesozoic Intrusive Suite intruded along the joint set preferentially, and so postdate it. Other intrusives (including the Harrison Grove stock) are foliated themselves. The foliation is overprinted by the east-vergent thrusts. It is inferred that both initial oblique-slip activity during D1 and later strike slip reactivation during phase D2 contributed to the development of this fabric.

This foliation has a fairly well-defined average orientation of vertical and an azimuth of 67°, with some spread. This average is parallel to the minor of the two concentrations observed in the fault trends, and at a significant angle (30° to 50°) to the particular major faults that are likely to have caused the foliation to form. The foliation is a well-developed joint set to spaced fracture cleavage to

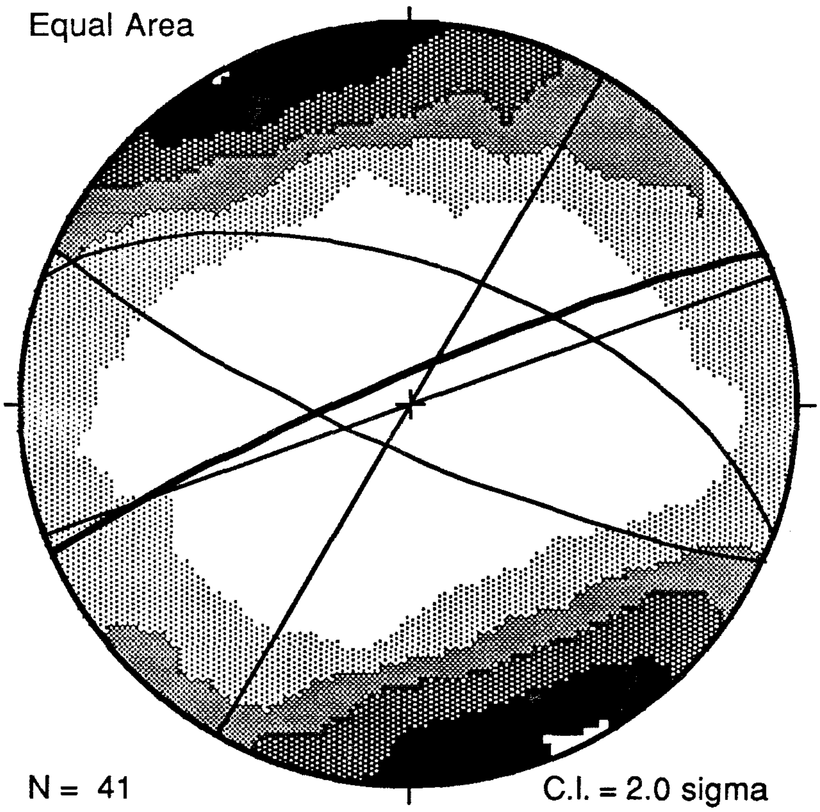
FIG.57



HIGH-ANGLE FAULT PROVINCE (D1) - BEDDING

CONTOURED DATA - POLES TO BEDDING
GREAT CIRCLE - BEST FIT CYLINDRICAL GIRDLE (164 58W)
BOX - AXIS TO GIRDLE (74 32)

FIG. 58



HIGH-ANGLE FAULT PROVINCE (D1) - CLEAVAGE AND FAULT PLANE MEASUREMENTS

CONTOURED DATA - POLES TO JOINTING AND SPACED FRACTURE CLEAVAGE
HEAVY GREAT CIRCLE - PLANE TO MAXIMA DEFINED BY POLES FOLIATION (247 85N)
GREAT CIRCLES - MEASURED FAULT PLANES

shear cleavage (Billings, 1972 - the latter term implying some slip along the cleavage planes, as was seen in several crenulated examples).

Alteration/Metamorphism No substantial metamorphism is related to this domain, with the exception of hydrothermal alteration along and near some of the high-angle faults. The alteration at high structural levels was characterized by replacement by Fe-minerals, and took place during the early oblique-slip phase. Extensive silicification and chalcedony formation also occurred. At deeper levels, such as near the Harrison Grove stock, Cu-minerals and extensive deuteric replacement is more characteristic (the stock may have provided the heat source). The textural and replacement features are characteristic of epithermal (0 to 1500 m and 50° to 200°C) to mesothermal (1200 to 4500 m and 200° to 300°C) deposits (Evans, 1980). More details are given in the appropriate chapters.

Stratigraphic Effects Field relations and detailed stratigraphy show clearly that the faults were mainly active in their oblique-slip phase during a limited period of time (e.g., they were acting as synsedimentary growth faults). This dip-slip period was during the deposition of upper member G and member H of the Happy Creek and member I and lower J of the King Lear Formation, dying out before the deposition of the Jackson Tuff (see appropriate discussions). Clastic debris from the hydrothermally altered fault zones is recognized in member I of the King Lear Formation. One exception to this timing is a pair of high-angle faults in Jackson Creek, which continued substantial growth fault activity into the deposition of member M of the King Lear Formation.

Cross-Cutting Relationships The high-angle faults cut the Harrison Grove stock (187 ± 2 m.y.) and the diorite sills intruding the upper Happy Creek Formation (170 - 175 m.y.). The faults are in turn truncated by the Happy Creek and Parrot Peak stocks (173 ± 14.3 m.y.) and are overlapped by the Jackson Tuff, as well as being pierced by the feeder dikes to that unit. Faults with only a dip-slip phase are cut and carried by thrust faults of phases D2 and D3. Those high-angle faults that also have a later strike-slip history are intimately related to these thrusting events as tear faults.

INTERPRETATION

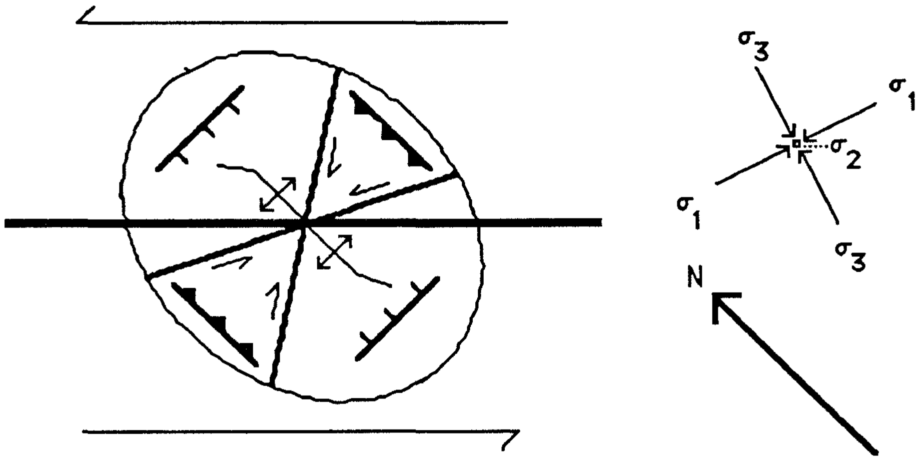
Kinematics Two discrete episodes of offset are apparent on the high-angle faults in this domain. The first episode involves predominantly dip-slip activity, and though strike-slip activity also occurred in this stage there are fewer opportunities to document such a nature. The dip-slip activity is discernable because of growth fault relationships in certain stratigraphic units. There is inferred to be a consistent sense of dip-slip offset. Azimuths of 100° or clockwise (the southeast conjugate set) are south side-down, and those counterclockwise are north side-down (the northeast set). This episode is assigned to and characterizes phase D1.

The second episode features strike-slip activity on some of the pre-existing oblique-slip structures, with the sense of strike-slip activity independant of the earlier sense of dip-slip offset. In this episode, the reactivated faults are tear structures intimately associated with both phase D3 and phase D2 thrusting. Different amounts of shortening and folding on either side of such a fault is observed in the east-vergent D2 thrusts southeast of King Lear Peak. With D3 phase thrusts (both senses of vergence) the thrust surfaces are often offset along the laterally-bounding high-angle tear faults, with different fold and thrust elements expressed on either side.

State of Stress Phase D1 is much more characteristic of wrench faulting than of an area undergoing rifting and extension (see fig. 59). The latter would be characterized by parallel fault planes dipping 60° or less, and with normal offsets. Within a wrench zone, however, two dominant conjugate orientations of high-angle faults at about 60° to one another form; these are called Riedel shears. The major and more developed synthetic set will be about 15° from Ú$_1$, while the minor and less continuous antithetic set will be about 75° from Ú$_1$ (Wilcox et al., 1973; Sanderson and Marchini, 1984; Ramsay, 1987). Such is the case here, with Ú$_1$ at about 100° (and hence Ú$_2$ vertical and Ú$_3$ at 10°) and the wrenching inferred to be left-lateral. The two conjugate systems are at 130° (dominant southeast set) and 70° (subordinate northeast set). Many of the smaller faults, trending in between, are themselves synthetic structures to the dominant set. This arrangement is consistent with the inferred sense of dip-slip offset on the two sets, as the upthrown side of a particular fault is that nearest Ú$_1$ (Harding and Lowell, 1979). It is also consistent with the sense of strike-slip offset on the faults-sinistral on the synthetic set and dextral on the antithetic set.

The evolution of a major throughgoing strike-slip fault is sometimes part of the later development of such a wrench zone, if shear continues. This is not observed to have happened within the Jackson Mountains

FIG. 59



GEOMETRIC ARCHITECTURE AND STRESS STATE FOR A SINISTRAL WRENCH ZONE (AFTER HARDING, 1974). THE NORTH ARROW SHOWS THE INFERRED ORIENTATION OF THAT SYSTEM WITHIN THE JACKSON MOUNTAINS DURING PHASE D1. DURING PHASE D2, $\sigma 2$ AND $\sigma 3$ SWITCHED, AND THRUSTING TOOK PLACE ALONG THE INDICATED ORIENTATIONS, REACTIVATING THE HIGH-ANGLE FAULTS FROM D1 AS TEAR (SLIP TRANSFER) FAULTS.

(though such a master structure might have existed nearby, and evidence for wrench features of the right age are abundant - see chapter 11). This particular wrench style is most likely to belong to a slightly transtensional retgime (e.g., one with a slight component of extension as well as transform activity). Wrench systems are notoriously complex, and this could be an oversimplified and inaccurate analysis.

The later strike-slip phase will be discussed again in the following sections, with the thrust systems that initiated it.

<u>Timing</u> The dip-slip (D1) phase of activity began at or soon after 175 m.y. (as dated by the intrusives it truncates). Most dip-slip activity of that phase ceased by about 165 m.y. (as thrusts cutting a high-angle fault with only dip-slip offset are themselves pierced by a 162 m.y. stock), and the high-angle faults are overlapped by Upper Jurassic and Lower Cretaceous sediments of the King Lear Formation. Several faults in Jackson Creek continued growth-fault activity into the Early Cretaceous. The timing of the later strike-slip episode(s) will also be discussed later.

<u>Summary</u> A left-lateral wrench setting in the last half of the Middle Jurassic caused the formation of two conjugate high-angle fault systems, with both dip-slip and strike-slip activity. The two dominant directions are at about 70° and 130°, with $\acute{U}_1$ at about 100°, and the faults were active during sedimentation of the uppermost Happy Creek and lower to middle King Lear Formations. No master strike-slip faulting was observed. Some of the faults were reactivated during both the later phase of D2 and during phase D3 thrusting as strike-slip tear faults.

Eastern Thrust Domain

Thrust-faulting characterizes the whole eastern ridge (the Trout Creek Spur) of the Jackson Mountains. This thrusting is west directed, and extends as far as the west edge of the central depression. Two distinct stages are inferred to have taken place in an earlier, Middle to Late Jurassic phase. There was also a much later renewal of thrusting in the domain in the late Early Cretaceous.

Description

Geometry The thrusts are defined in the field by abrupt changes in lithologic units, associated deformation, the preferential development of shear foliations and brecciation, presence of epidote veining and slickensides, and by their topographic expression.

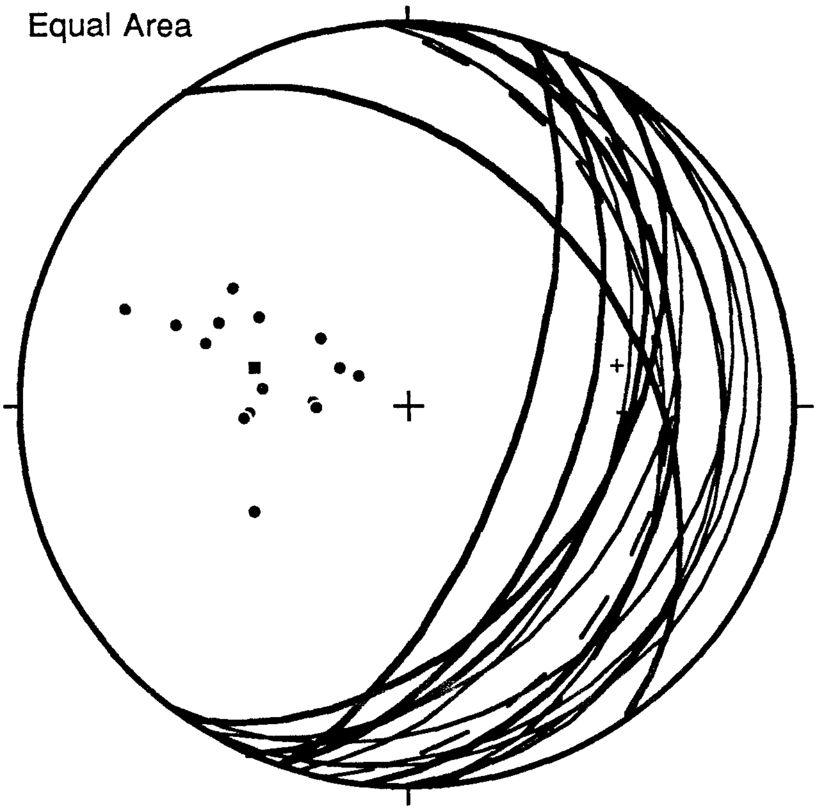
In the Trout Creek Spur, the fault planes strike between 0° and 40° and dip up to 45° to the east (see fig. 60). The average is 13°, 32°E. Those thrusts in the frontal zone (towards the west) that cut up into the King Lear Formation and especially those that reached the paleosurface are much flatter, with dips of 10° to 20° or even a horizontal orientation. Individual fault planes can be seen to curve and flatten as they reach this area, as klippen in front of the main thrust system. The frontal imbricate zone is locally breached by erosion to form windows. The thrusts further back (to the east) in the structural stack tend to be steeper, from 30° to 45°. It should be remembered that Tertiary tilting of about 10° to the east has steepened the thrust planes from their original attitude, so that when formed these thrusts dipped from about 35° to the east to 10° or so to the west.

The thrust system forms an anastamosing network both laterally and down-dip. Thrusts join and diverge along their map traces to form mega-phacoids hundreds of meters in thickness and with lengths of 3 to 13 km. The result is pile of about 16 exposed thin imbricate nappes, about ten deep at any one point. The average structural spacing of the thrust planes is about 400 to 500 m, with the maximum at 900 m. The average exposed length is about 8 to 9 km. These thin, imbricate and anastamosing thrusts are typical of shallow level, brittle deformation. To the north, the forwardmost (and most western) three or four thrusts die out into a fold belt, becoming blind thrusts.

Such thrust belts are commonly underlain by a very low-angle main thrust surface, and splay off of it. This surface was not exposed but would have to lie within or below the McGill Canyon Formation, as that unit is involved in the imbricate stack. The low-angle frontal thrusts are (in the terminology of Ramsay, 1987) termination splays. The other thrust splays have isolated, diverging (split into two), rejoining (coming off of a plane and merging back) and connecting (from one main splay to another) geometries. The root zone is not present in the Jackson Mountains, and there are no back thrusts. A duplex nature (where the fault planes curve towards a horizontal roof thrust at their tops) is quite lacking, as is visible ramping. A ramp and flat architecture is inferred, however, as the stratigraphic packages within the thrust sheets are usually parallel to the bounding faults (especially in the Happy Creek Formations). The eastern thrust domain is thus a simple imbricate fault stack.

Along the thrust surfaces themselves (where exposed), brecciation and

FIG. 60



THRUST FAULTS - EASTERN THRUST PROVINCE

DOTS - POLES TO FAULT PLANES
FILLED BOX - MEAN VECTOR TO THOSE POLES
DASHED GREAT CIRCLE AT 13 33 - PLANE TO THAT VECTOR
HEAVY GREAT CIRCLES - MEASURED FAULT PLANES
GREAT CIRCLES - CALCULATED FAULT PLANES
SECOND DASHED GREAT CIRCLE AT 355 35 - FLASER
FOLIATION IN KLFm SILL
CROSS - STRETCHING LINEATION IN THRUST ZONE

THRUSTING TOOK PLACE IN BOTH STAGES OF PHASE D2
AND DURING PHASE D3.

a pronounced phacoidal to planar shear fabric are developed.

Offset The sense of offset is clearly visible in a number of sites in this domain. In all cases, a compressional nature with older over younger was documented (except where later Basin-Range normal reactivation has taken place). The general architecture within the thrust belt has much deeper stratigraphic levels being exposed in the easternmost part. Offset along individual faults is always compatible with this, and there has been extensive duplication of stratigraphy. The accompanying fold sets agree with this evaluation of vergence. Fold axial planes are upright to east-dipping, and the sense of shear deduced from the fold profiles is always headwall over footwall. Asymmetric drag synclines are present in the King Lear Formation directly under the frontal faults, and further back in the stack anticlinal nappes are common in the Bliss Canyon Formation.

On a mesoscopic scale, sigmoidal tension gashes and slickenside steps in the Trout Creek Spur indicate westward vergence (see fig. 61). Flaser fabrics in a shear zone in a sill of member M of the King Lear Formation gave a like sense, as did small soft-sediment thrust faults in the upper lacustrine portion of the same member. Small-scale thrust and ramp features are present in the argillite of the Bliss Canyon Formation.

The slickensides give a slightly more complex picture (which will be supported later by bedding, fold axis and cleavage data). Only the azimuths were plotted, as the plunges were highly variable and did not contain signficant information (a result of the phacoidal fabric within the thrust zones). Two quite distinct concentrations are seen in a rose diagram plot of the slickenside azimuth data. One is at about 70°, the other (somewhat stronger) at about 125°. From field relations the former set is inferred to be the earlier one. The former set is also oblique to the thrusts, while the latter indicates dip-slip offset.

Offsets on the thrust system seem to be greatest to the south, where strata as deep as the McGill Canyon Formation are exposed on the same thrust plate that has the King Lear Formation far to the north. The alternative is that the whole system is tilted up from south to north. In either case, this occurred during or after D2 but before D3. The favored interpretation is that thrusts originally trending about 350° during the first stage of D2 were rotated during the second stage. The part of the stack in the middle and southern portion of Trout Creek Spur (with deeper levels brought up to the south, and thrusts trending more northeasterly than in the north) progressively underwent more shortening and rotation than those to the north.

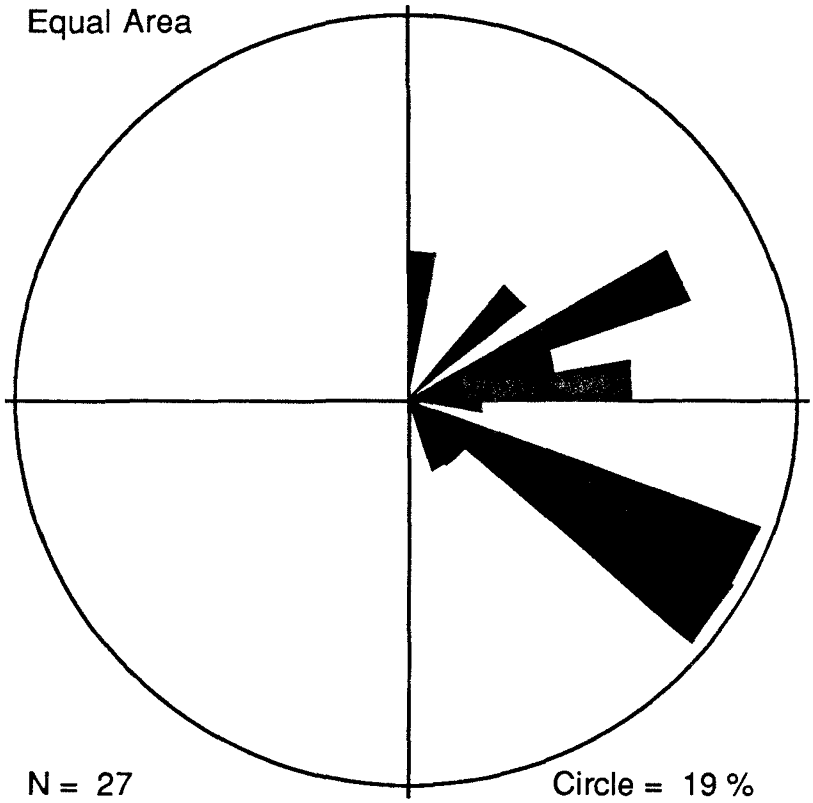
Folding The thrust sheets, where stratigraphy is internally discernable, are often but not always macroscopically folded. Folding is much more likely to occur in the well-bedded and structurally weaker Bliss Canyon and King Lear Formations than in the more massive and homogenous Happy Creek Formation.

Around Bottle Creek, the macroscopic folds are large and open to tight, with amplitudes of several hundreds of meters and axial planes vertical or inclined to the east; overturning of the fold limbs does occur. The limbs are planar, and the fold hinges angular, in chevron style. The fold axes plunge gently northeast or southwest (see fig. 62, 63, 64). An axial planar cleavage exists to these folds (see next section).

In the north of the range, two generations of macroscopic folds are visible - the eastern and older (in the Bliss Canyon and lower Happy Creek Formations) trending about 355° and located just south of Buff Peak, and a second and younger (deforming the upper Happy Creek and King Lear Formations) trending about 45° in Happy Creek. They are separated by a thrust fault of phase D3, which truncates the older set at a substantial angle but is genetically related to and parallels the younger set. In fact, the younger set is located at the point where thrusting jumps from the structurally higher fault cutting member L around Buff Peak to the north, to the frontal (and therefore further forward in the imbricate stack) thrust plane to the south in Jackson Creek. The fold set formed as a consequence of this geometry, after a fashion similar to a lateral ramp between the two planes, transfers slip from the northern and higher fault to the southern and lower fault plane.

Bedding plane slip with extensive slickensided surfaces is very common in the younger fold set in Happy Creek. The interlimb angles range from 60° to 90°. These later folds are large-scale chevron to box folds, with very sharp and brecciated fold hinges and moderate plunges to the northeast, and with cylindrical to conical forms. The axial traces also hint at dextral drag. This geometry agrees with the interpretations of the system as a slip transfer system, dying out gently to the north and abruptly to the south. The axial planes are vertical to slightly east dipping, and the amplitudes are 500 to 600 m. The folds turn into thrusts (along the axial planes of the synclines) to the south, but the thrusts (predominantly D2 features) have only minor offset in D3 during the folding, and are pinned further the south during this time frame.

FIG. 61

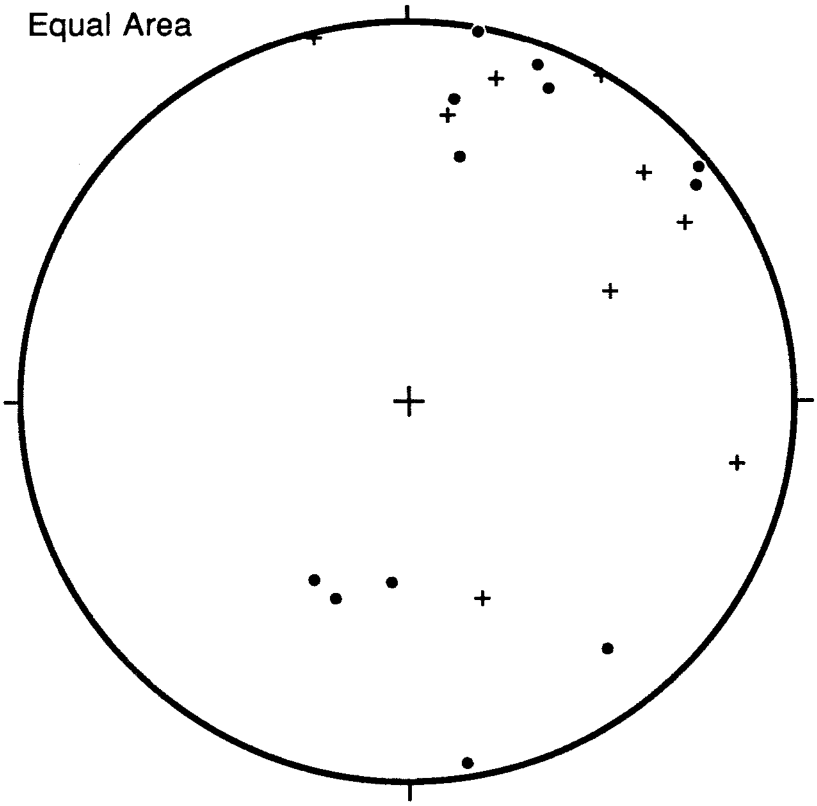


SLICKENSIDES FROM THE EASTERN THRUST SYSTEM

NOTE TWO DISTINCT MAXIMA AT APPROXIMATELY 65 AND 120. THE FORMER IS FROM THE EARLY STAGE OF PHASE D2, WHILE THE LATTER IS FROM THE LATER STAGE OF D2 AND FROM D3.

ONLY THE AZIMUTHS ARE PLOTTED, AS THE PLUNGE VARIED GREATLY AND DID NOT APPEAR TO CONTAIN SIGNIFICANCE.

FIG.62

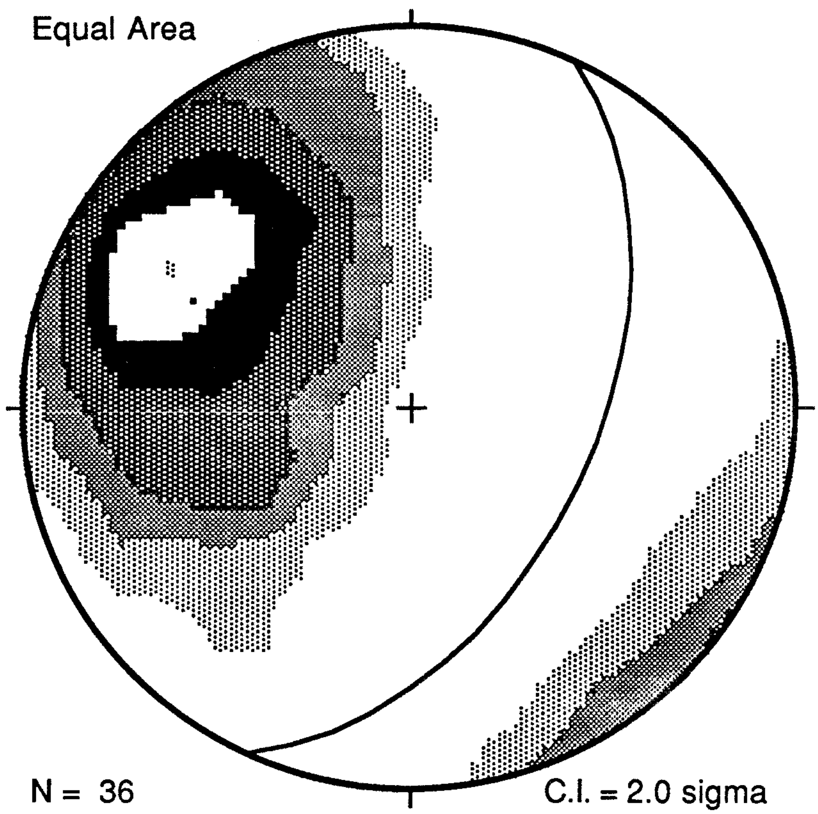


EASTERN THRUST SYSTEM - FOLD AXES

DOTS - MACROSCOPIC FOLD AXES CALCULATED FROM BEDDING MEASUREMENTS
CROSSES - MESOSCOPIC FOLD AXES MEASURED IN THE FIELD

THE FOLDS FORMED DURING BOTH STAGES OF PHASE D2, AND DURING PHASE D3. REFOLDING OF EARLIER STRUCTURES BY LATER EVENTS PROBABLY TOOK PLACE, AND ACCOUNTS FOR THE SCATTER IN DISTRIBUTION.

FIG. 63

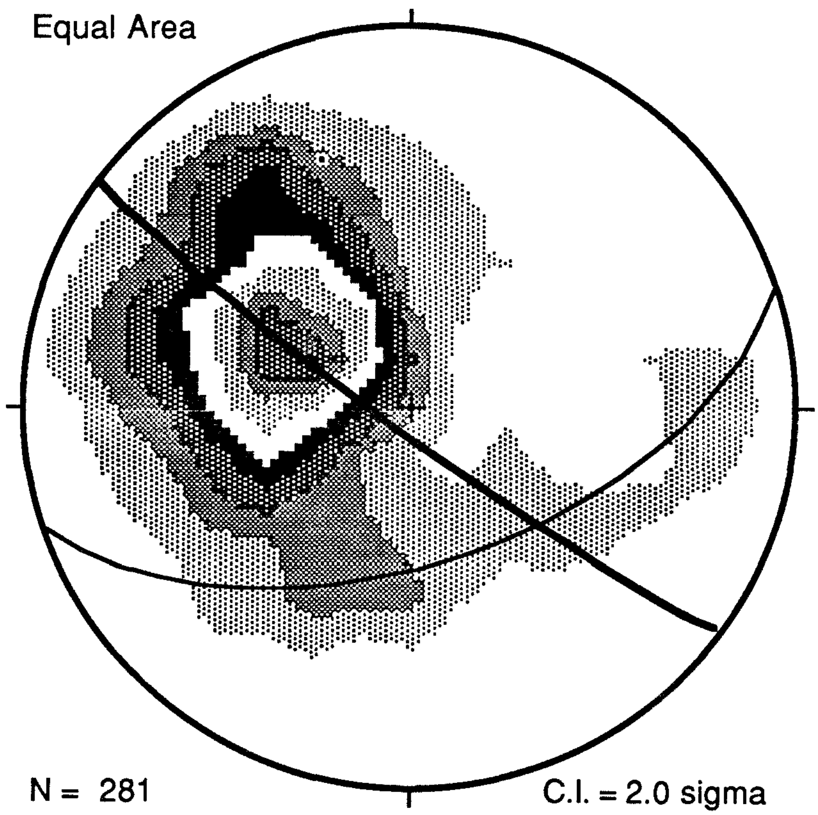


EASTERN THRUST SYSTEM - CLEAVAGE

CONTOURED DATA - POLES TO CLEAVAGE
BOX - MEAN VECTOR FOR POLES (295 39)
GREAT CIRCLE - BEST FIT GIRDLE TO MEAN VECTOR (25 51E)

CLEAVAGE FORMED IN BOTH PHASES OF D2 AND, IN THE WESTERN-MOST THRUSTS, D3 AS WELL.

FIG. 64



EASTERN THRUST SYSTEM - BEDDING

CONTOURED DATA - POLES TO BEDDING
HEAVY GREAT CIRCLE - MAJOR GIRDLE FIT TO POLES (128 85S)
GREAT CIRCLE - MINOR GIRDLE FIT TO POLES (70 57S)
BOXES - POLES TO THOSE GIRDLES

MAJOR GIRDLE - LATER PHASE OF D2, AND D3
MINOR GIRDLE - EARLY PHASE OF D2

The thrusts might have been overlapped by member K (the folded lithology) and then reactivated slightly.

The older set of folds south of Buff Peak are alike in geometry and style to the younger set, other than their trend and truncation. A second deformation is indicated in the area of these older folds by complex mesoscopic bedding patterns, though whether younger or older is not clear in the field. I infer it to be younger cross-folding from the second stage of D2, from the macroscopic relationships.

Immediately beneath the frontal thrusts (D3 phase) in a number of localities, the upper King Lear Formation (members J to M) is folded into a discontinuous syncline. The syncline may have formed just west of and in front of the frontal thrust faults as they were just reaching the paleosurface during King Lear sedimentation. The structure is open and upright to tight and reclined, asymmetric to the west with the axial surface east-dipping. The upper limbs in some folds are overturned, with minor thrusting along the axial plane of the syncline. The axial traces trend parallel to the strike of the superimposed thrusts. Amplitudes are on the order of 20 to several hundred m. Further away from the this zone, scattered open low-amplitude folds are present in the King Lear Formation, but generally the unit is homoclinally east dipping.

The mesoscopic folds in this domain (present mainly on the east, where Bliss Canyon rocks are involved) are more variable in orientation. Some are inferred (because of their orientations) to have been refolded. These folds are similar in geometry and style to their larger equivalents. They have planar limbs and angular hinges, are chevron to boxfold to polyclinal, and are parallel (the limbs maintain constant thickness). These features are distinct from the earlier soft-sediment folds.

Fabrics Axial planar cleavages were formed during both stages of D2 and during D3. The foliation from the early stage of D2 is oblique to the other two generations, and is only seen in the Bliss Canyon and lower Happy Creek Formations in the eastern and most uplifted portion of the thrust domain. A plot of poles to cleavage for the entire domain indicates that cleavage is generally steeper than thrusting by about 20°, but parallel in strike. This observation was also made in the field.

The D2 foliations in the Bliss Canyon Formation are expressed as crenulations on bedding (slip cleavage), fissility in the argillite (slaty cleavage), or a spaced (solution) cleavage in coarser-grained lithologies. In the Happy Creek Formation, the D2 foliation is a spaced fracture cleavage to (near the thrust zones) a phacoidal to parallel shear cleavage. The volcanic rocks are generally resistant to expressing a well-developed foliation. The closer to the thrust, the better developed and more continuous the cleavage, and the more likely it is to have epidote seaming and slickenside formation.

Foliation in the King Lear is generally not so well developed - there is a very weak D3 axial planar slaty cleavage in the tighter folds and a spaced fracture cleavage in the few meters just underneath the thrust plates, but the foliation dies out rapidly away from the thrusts. The thrust-generated cleavage of phase D3 in the King Lear and Happy Creek in the western part of the domain is quite distinct in orientation and expression from the spaced fracture cleavage and joints formed during event D1. Caution is required in the interpretation of these foliations, as different types have different origins. For example, the axial planar slaty cleavage forms from flow and mineral realignment parallel to Ú$_1$, while the spaced fracture cleavages form by shear failure at a distinct angle (about 30°) from Ú$_1$, and the solution cleavage forms by preferential solution along stylolites in the plane of maximum compression (Billings, 1972).

Some conglomerates in the thrust zones are deformed. One example from the uppermost Bliss Canyon Formation exhibited an aspect ratio of 5:2:1, defining a down-dip shear lineation and some flattening in the plane of cleavage.

Separation of Phases The plot of bedding poles (fig. 64) in the domain shows that the dominant factor was southeast-northwest compression, with vergence to the northwest (about 300°). However, there is considerable scatter and I infer a second, weaker element due to compression with vergence to the southwest (about 255°). This weaker element matches the weaker of the two concentrations in slickenside orientations, and is earlier than the dominant fabric. The dominant element matches quite well with the major slickenside orientation. As these fabrics both occur in thrust sheets plugged by a 162 m.y. pluton, they cannot be part of D3, (which, as mentioned later, postdates the pluton). Instead, there are two stages to D2 deformation -an earlier, more oblique (dextral) stage verging about 255°, and a later, more dip-slip stage verging 45° away, towards 300°. This multiphase history is also seen in the cleavage. The later stage overprints the older. D3 is coaxial with the second stage of D2, but is distinctly later and occurred on the frontal few thrusts only.

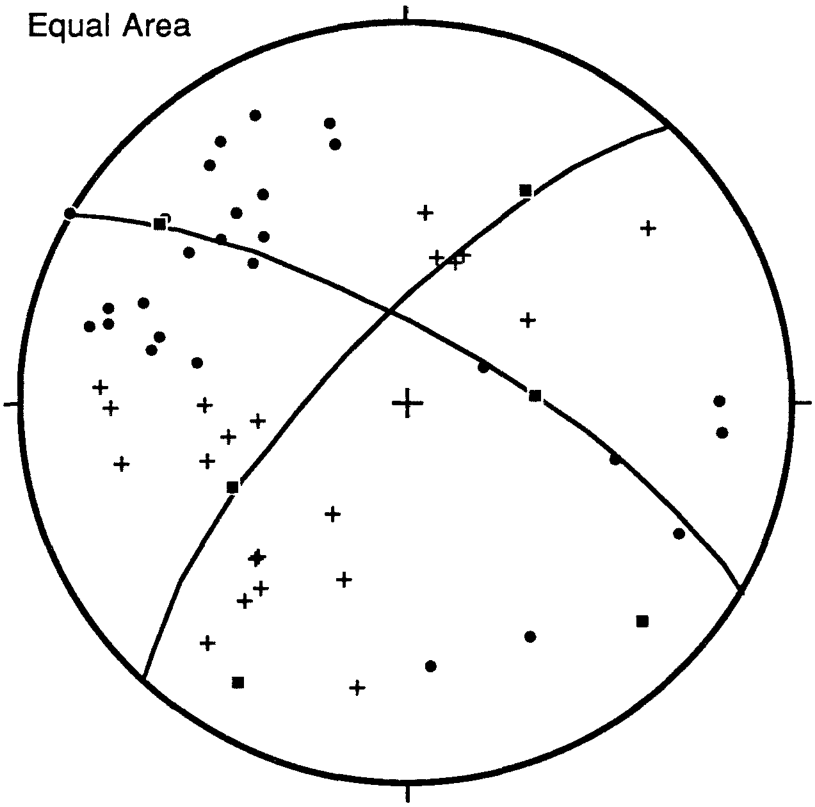
The slickenside rose diagrams and field relations also indicate a multiphase deformation history.

Plots of bedding, cleavage and fold axes from the eastern (and pre-D3) portion of the thrust belt only help distinguish the two stages within D2 (fig. 65, 66, 67). Once bedding in the macroscopic folds (most of which are definitely coaxial with the thrusts, and were formed during the later stage of D2) is removed, a second girdle indicating northeast-southwest oblique compression is indicated. Cleavage similarly indicates that two events took place here - an early and overprinted phase of northeast-southwest oblique compression, and later phase of east-west to eastsoutheast-westnorthwest compression. Both occurred on the same system of thrust faults, and belong to phase D2. Plotting the fold axes from this area does not clarify the situation much, as there is too little data.

Metamorphism Regional metamorphism in the thrust belt in the eastern portion of the range is very low grade. No diagnostic metamorphic minerals were seen, and the sedimentary fabrics are not overprinted by any dynamothermal textures. The argillites are locally slaty but not phyllitic.

Stratigraphic Effects The first pulse of thrusting can be seen in the stratigraphy of the King Lear Formation, with the sudden appearance of cherty/lithic detritus from the McGill Canyon and Bliss Canyon Formations in member J. The earliest phases of thrusting might have contributed to the upper part of member I. This pulse of orogenic molasse by phase D2, possibly the first stage. The interfingering of lower members J and K may reflect ongoing D2 phase (the later stage?) uplift and erosional stripping, which must then have stopped in the Late Jurassic.

FIG. 65



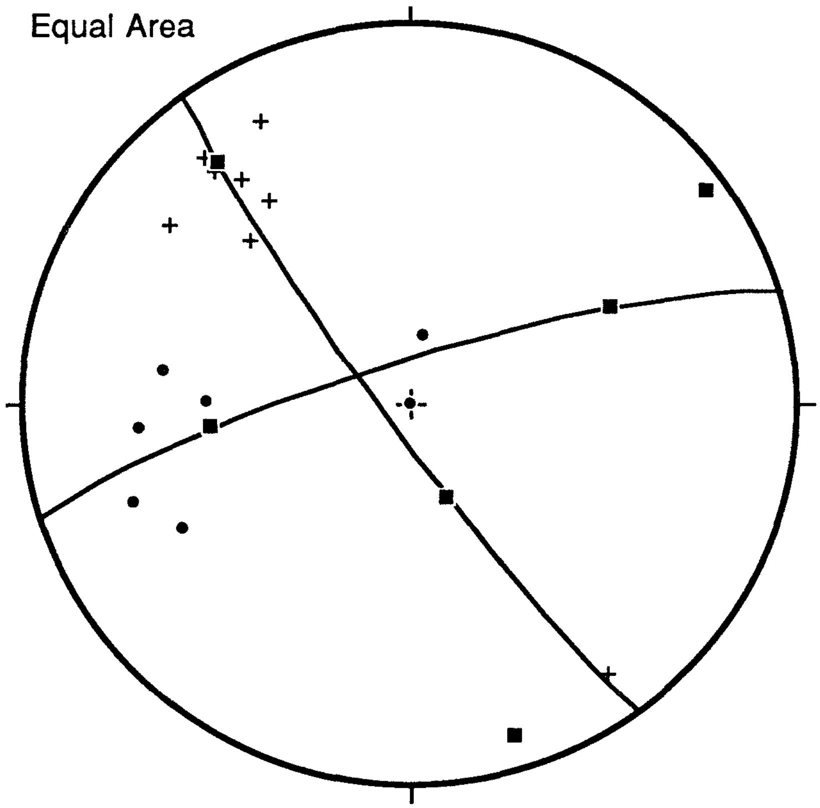
BEDDING FROM THE BLISS CANYON AND LOWER HAPPY CREEK FORMATIONS, EASTERN PART OF THE EASTERN THRUST SYSTEM

DOTS - POLES TO BEDDING, INFERRED TO BELONG TO LATER PHASE OF D2 WITH A GIRDLE AT 300 74N AND A POLE TO THE GIRDLE (FILLED BOX) AT 210 16

CROSSES - POLES TO BEDDING, INFERRED TO BE FROM EARLIER PHASE OF D2 WITH A GIRDLE AT 223 74N AND A POLE TO THE GIRDLE (FILLED BOX) AT 133 16

THE SPREAD IN DISTRIBUTION MAY BE CAUSED BY REFOLDING DURING THE LATER PHASE OPF D2, OR EVEN DURING D3.

FIG. 66

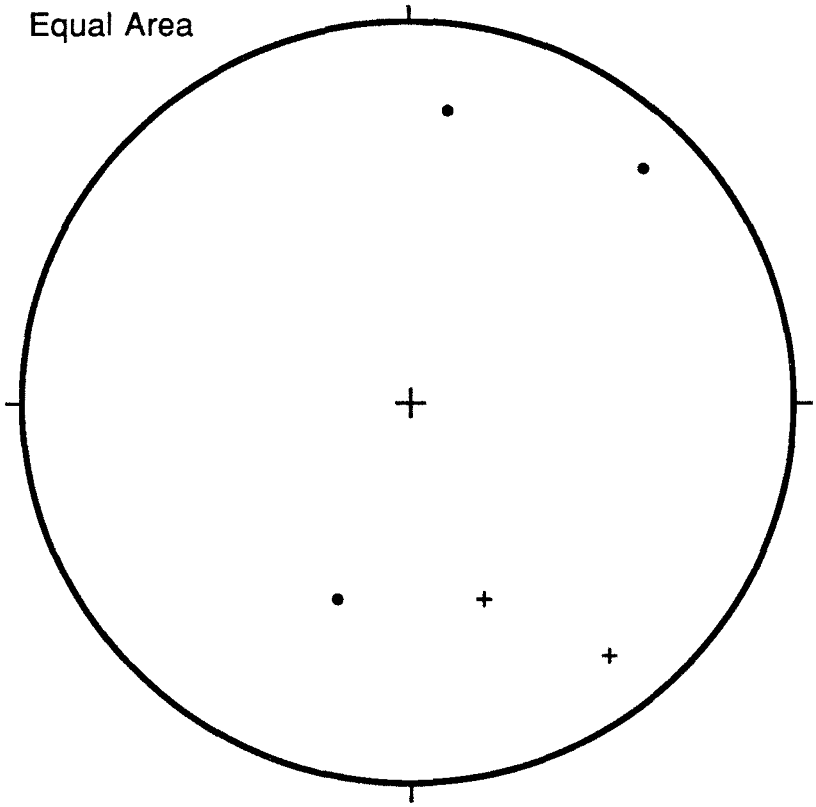


CLEAVAGE FROM THE BLISS CANYON AND LOWER HAPPY CREEK FORMATIONS, EASTERN PART OF THE EASTERN THRUST SYTEM

DOTS - EARLIER PHASE OF D2, WITH GIRDLE AT 253 80N AND THE POLE TO THE GIRDLE (FILLED BOX) AT 163 10
CROSSES - LATER PHASE OF D2, WITH GIRDLE AT 144 84W AND THE POLE TO THE GIRDLE (FILLED BOX) AT 54 6

FIG.67

Equal Area



FOLD AXES FROM THE BLISS CANYON AND LOWER HAPPY CREEK FORMATIONS, EASTERN PART OF THE EASTERN THRUST SYSTEM

DOTS - FOLD AXES INFERRED TO BE FROM LATER PHASE OF D2
CROSSES - FOLD AXES INFERRED TO FROM EARLIER PHASE OF D2

BOTH MESOSCOPIC (MEASURED IN FIELD) AND MACROSCOPIC (CALCULATED FROM BEDDING) FOLD AXES ARE INCLUDED, BUT DATA IS INSUFFICIENT TO ANALYZE FURTHER.

At some undetermined point in these members, and probably after a considerable hiatus, D3 phase uplift and erosion became a factor. Member N, the sedimentary fault scarp breccia, is intimately related to the frontal D3 phase faulting. These deposits were shed from the advancing frontal scarps once the thrust faults had surfaced, and were then deformed and over-ridden.

East of Buff Peak, the eastern part of the thrust system is overlain unconformably by member J of the King Lear Formation. The upper section of the thrust belt was thus active during D2 only (these faults are also pinned to the south by the 162 m.y. stock). The King Lear here is para-autochthonous, as the whole area was then in the upper plate to a D3 thrust just to the west.

From the distribution of member N of the King Lear at four different structural levels, thrusting on the frontal faults must have been out of sequence. For this distribution of N, the underlying thrust plate would have to have been uplifted tectonically in a thrust sheet, then itself overlapped by debris and overthrusted from behind by a fault further back in the stack. Some of the higher occurrences also have very thin and early King Lear sections underlying N, and so might have been active during D2 instead of D3 (N is then of two distinct generations).

Cross-Cutting Relationships The thrusts attributed to D2 are truncated by a 162 m.y. stock (the Trout Creek Spur monzodiorite), and in turn cut 170 - 175 m.y. diorite sills. They are also pierced by the tonalite dikes and sills of the Bottle Creek suite, and by the rhyolite plug and dikes of member L of the King Lear Formation. The latter silicic volcanic complex is in turn thrusted by phase D3, which may affect the

Trout Creek Spur stock. The thrust and fold belt of phase D2 south and east of King Lear Peak is also plugged by the Navajo Peak stock (phase D3 thrusts of the western thrust domain cut this body). In addition, member M has compressional soft-sediment deformation indicating nearby thrusting while the top of the unit was still wet and unconsolidated. One dacite sill of M (dated at 115 m.y.) similarly has flaser fabrics, indicating thrusting while still warm. The laccolith complex in member M also indicates a compressional stress state during intrusion (D3), the same as the gabbro and diorite laccolith and sills of the Early Mesozoic Intrusive Suite (phase D2) (Williams and MacBirney, 1979).

Interpretation

Kinematics Three separate compressional events are recorded in the eastern thrust domain. The first two are lumped together as two stages in D2 because of their proximity in time and their use of the same family of thrust faults. The third is phase D3, which took place considerably later and along only the frontal part of the imbricate thrust stack formed by D2.

The first stage of D2 involved oblique convergence, with an inferred dextral sense. Oblique-slip along the thrusts (which dip east and trend around 15°) is indicated by slickensides, cleavage, and folding. Vergence was towards 255°. The second stage followed on the heels of the first in the same area, and involved pure dip-slip compression, as also documented by slickensides, cleavage, and folding patterns. The vergence was to an azimuth of 300°. The second stage overprints the first one, and is the dominant factor forming the architecture of the eastern thrust domain. A long period of quiescence was followed by renewed D3 compression, also verging 300°. The frontal faults from the second stage of D2 were reactivated, and several new thrusts splayed off in the front of the imbricate stack. The bulk of the stack was not reactivated, as it is pinned by several stocks and dikes and is locally overlapped by the King Lear Formation.

During D3, some of the pre-existing high-angle faults with trends close to perpendicular to the thrusting were reactivated as tear faults, transferring slip from one thrust sheet to another. In several cases, thrust and fold traces end against the reactivated tear faults, or are differentially advanced along them. Some of the frontal thrust plates of D3 became gravity slides, dipping west (when Tertiary tilting is restored) and cutting back down section into the King Lear Formation as they slid west.

Shortening is difficult to reliably estimate because of the paucity of good markers in the thrust zone, but was on the order of 20 to 25 km in the exposed part of the eastern thrust domain during both stages of D2. D3 thrusting was on the order of 5 to 10 km, with the folds in the Happy Creek area taking up 2 to 3 km of the compressional shortening in that area as the thrusting decreased and slip was transferred back in the stack to the reactivated Buff Peak thrust fault.

State of Stress During the second stage of D2 and during D3, Ú$_1$ was horizontal with a trend of about 110° (assuming that folding, axial planar cleavage formation and faulting were perpendicular, and slickensides parallel, to Ú$_1$). As thrusting was taking place, Ú$_3$ was vertical, and Ú$_2$ horizontal and trending 20°. The stress state for the second stage of D2 and for D3 is compatible with that for D1 and the high-angle faulting, with $\acute{U}_2$ and $\acute{U}_3$ exchanged. The stress state during the first stage of D2 thrusting is not as clear to me because of the oblique-slip nature, but likely represents the intermediate stage between D1 and the second stage of D2 (which are constrained to have been very close to one another temporally, in the period from 170 - 175 to 162 m.y.). As the two stress axes switched, the thrusts (which probably already existed in incipient form, as the stress state during intrusion of the sills was compressional) were in a suitable orientation for oblique compression with a dextral sense.

Timing Both phases of D2 are constrained to have occurred after or just overlapping the high-angle faults of phase D1, which cut sills dated at 170 - 175 m.y. The thrusted stratigraphy is as young as 169 $\pm$ 10 m.y. D2 thrusts are pierced by a 162 m.y. monzodiorite stock and so thrusting had ceased by that point in time. D2 faults in the area of Buff Peak are also unconformably overlain by member J of the King Lear, which elsewhere is Late Jurassic to Lower Cretaceous.

D3 occurred in the Aptian (and perhaps somewhat earlier). The late Early Cretaceous lacustrine sediments of member M, and a dacite sill of the same member, both are contemporaneous with thrusting. A gap in time is also indicated by brittle deformation (D3?) in the Trout Creek Spur stock, which plugs the D2 thrusts.

Summary The first stage of D2 involved thrusting with a dextral oblique nature, and occurred immediately after or even overlapped in time with the high-angle wrench-related faults of phase D1, probably in the period 165 to 170 m.y. The thrusts trended just west of north-south, and vergence was towards an azimuth of 255°. This evolved, also quickly (and ending by 162 m.y.), into the second stage, which continued the thrusting but in a more dip-slip fashion. The thrusts were now trending more east of north, and vergence was to an azimuth of 300°. Continued imbrication, overprinting of the first stage features, rotation of the thrusts, and differential shortening (greater in the north than south) established the present architecture of the imbricate stack. Both phases (D2 and D3) are west-vergent.

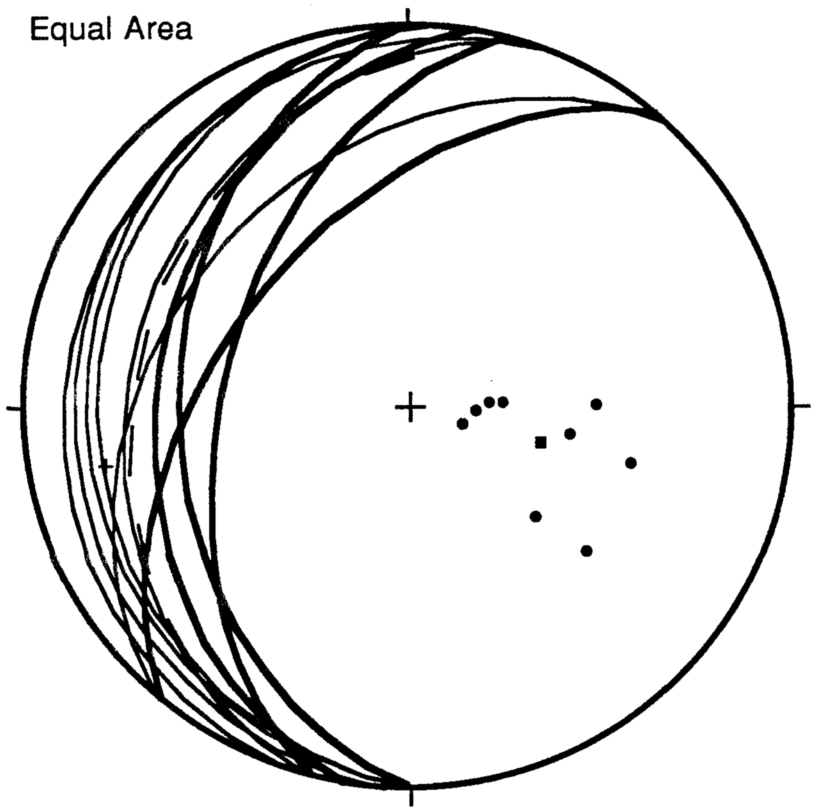
The stress axes for D1 and D2 are coaxial, with the interchanging of the $\acute{U}_2$ and $\acute{U}_3$ stress axes. The two phases (D1 and D2) represent evolution within the same (though complex) tectonic environment (see fig. 59). Phase D3 is very similar in nature and orientation, though much later in time, to the second stage of D2. The frontal imbricate thrusts from D2 were reactivated, utilizing some of the high-angle faults of D1 as transfer faults, and several new imbricate frontal splays formed. Phase D3 deforms Lower Cretaceous sediments and overlaps in time with a sill dated at 115 m.y.

Western Thrust Domain

This system is located along the west flank of the western ridge within the range, from the mouth of Mary Sloan Creek to south border of the field area. It is geographically separate from the eastern thrust domain and abuts, along its eastern margin, the central depression containing the high-angle fault domain.

Geometry The average thrust fault plane orientation in this domain is 198°, 29°W (see fig. 68). The range is from about 180° to 220° in trend,

FIG. 68



THRUST FAULTS - WESTERN THRUST PROVINCE

DOTS - POLES TO FAULT PLANES
FILLED BOX - MEAN VECTOR TO POLES
DASHED GREAT CIRCLE - PLANE TO THAT MEAN VECTOR
HEAVY GREAT CIRCLE - MEASURED FAULT PLANES
GREAT CIRCLES - CALCULATED FAULT PLANES
CROSS - TECTONITE MINERAL LINEATION IN THRUST ZONE

THRUSTING TOOK PLACE DURING PHASE D3 ONLY, EXCEPT FOR ONE THRUST SOUTH AND EAST OF KING LEAR PEAK FROM D2.

and 15° to 50° in dip (all to the west). The exposed part of this system consists of no more than three anastamosing fault strands in any one area. In the north, near the mouth of Jackson Creek, there are two distinct and rather low-angle strands from phase D3. These eventually join as two steeper and converging splays in the area of Bliss Canyon. There is a shallower diverging splay coming off of the frontal thrust around Alaska Canyon, which dies out northwards as a blind thrust and anticline to the north near Hobo Canyon. Below the single moderately-dipping D3 strand north of McGill Canyon there abruptly appears a second, steeper and more eastern D2 thrust, bounded on the north by a major high-angle fault (trending 120°), which juxtaposes it against the Happy Creek Formation. This high-angle fault is present, though slightly sinistrally offset, in the higher D3 thrust plate as it continues down from the north. The high-angle fault here juxtaposes the McGill Canyon and Bliss Canyon Formations. Another high angle-fault occurs in the domain, offsetting the uppermost thrust between Hobo and Alaska Canyons. The thrust planes observed in the McGill Canyon area are exposed as far south as Navajo Peak, at the southern end of the western flank of the range, where poorly exposed argillites (assigned to the McGill Canyon Formation) are put over the Navajo Peak stock and the Happy Creek Formation by a D3 thrust. D2 faults occur around King Lear Peak, with McGill Canyon over Happy Creek and that over King Lear Formation. The latter thrust is quite shallow, dipping only 15°.

Along the D3 fault planes near Jackson Creek, a tectonic mega-breccia was observed at one site, with blocks of both upper and lower plates involved. In other sites in this area, a powdery black fault gouge is present along the actual slip surface. Near that surface is a few meter wide zone of phacoidal shearing, with extreme ductile boudinage of lithologies caught up in the zone. Further south, fabrics along the D2 and D3 thrust are less highly developed and are characterized by a narrow zone of more intense fracturing and well-developed cleavage parallel to the fault, with the fault zone itself reddish and altered.

Offset The sense of offset on these north to northeast trending, west-dipping faults is abundantly clear in this western domain. The sense of stratigraphic offset across the fault planes is always hanging-wall up (except for the D3 thrust fault west of King Lear Peak, where there is net normal offset due to Tertiary reactivation). The hanging-wall to D3 also has a much greater metamorphic grade developed, particularly in the north. Drag folds underneath the divergent splay (D2 or D3?) between Alaska and Hobo Canyons indicate the same sense of shear. Macroscopic fold asymmetry in the Bliss Canyon and lower Happy Creek Formations indicates eastward vergence during D3 - looking north, the macroscopic folds are Z-shaped in profile, their axial planes are west-dipping, and some fold limbs are overturned to the east. Mesoscopic fold asymmetry is the same in the D2 generation, and is quite evident in the area east and southeast of King Lear Peak. Small mesoscopic thrust faults are present in some outcrops of the McGill Canyon Formation, with westerly dips and the same sense of simple shear as the macroscopic faults.

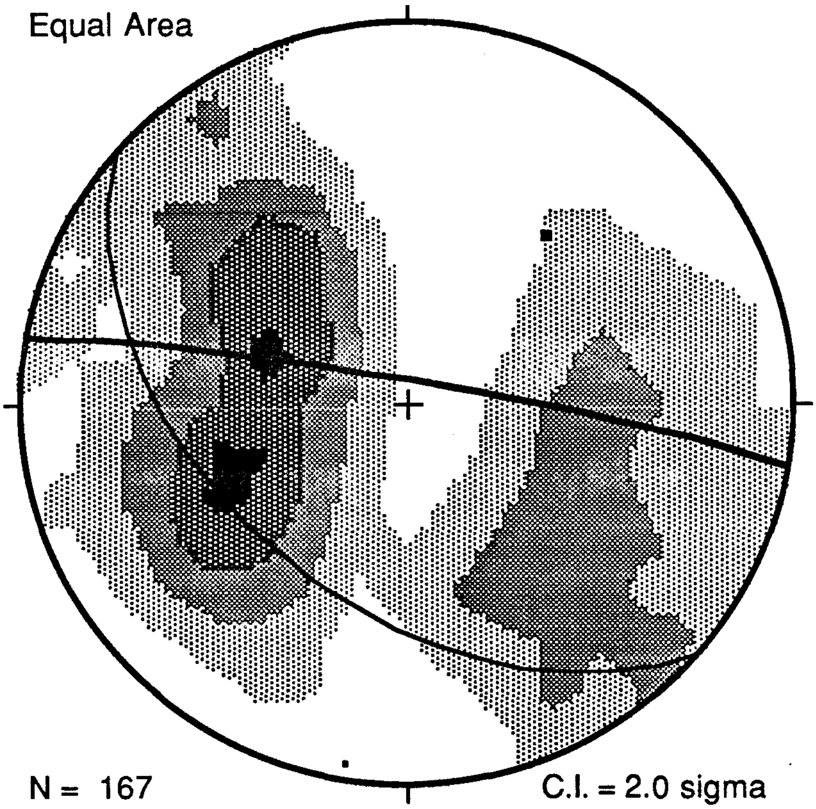
A slightly oblique nature to the D3 thrusting is indicated by the sinistral offset of the high-angle fault (and by folding - see following paragraph). On the other hand, the tectonite mineral lineation around the mouth of Jackson Creek indicated more of a dextral oblique vergence.

Offset on the high-angle fault north of King Lear Peak is sinistral, and it acted as a slip transfer feature for the D2 thrusts before being offset in D3. This large structure between Bliss and McGill Canyons had an early D1 phase of minor dip-slip activity, and at the same time substantial amounts of sinistral strike-slip offset occurred - on the order of 4 km, judging by the general dip of the sequence and the extent of the stratigraphic discordance. The tear fault nature is indicated by the fact that the offset on the upper plate where the high-angle fault was reactivated as a tear fault (to the frontal D2 thrusts) is less than on the lower plate, and is inferred to be about 2 km. The structure crosses the range, and has 3 km of offset on the far side (seen in the offset of the King Lear/Happy Creek contact); the offset increases to the west because folding on the south side during D2 took up additional displacement during tear fault activity. Within the McGill Canyon Formation near the fault, an early phase of vertical folding and strike-slip microfaults support these conclusions (see section on phase separation).

Folding In the middle and northern part, all of the the Bliss Canyon plus the lower Happy Creek Formation is deformed into tight to isoclinal, horizontal and moderately to gently inclined D3 folds. The axial planes here are almost parallel to the thrust planes bounding the nappe (see fig. 69, 70, 71). The folds in detail have a consistent en echelon pattern, with the fold traces trending slightly clockwise of the thrusts. The pattern established by these synformal and antiformal structures indicates somewhat oblique slip, with a sinistral sense. This disagrees with the vergence implied by the tectonite lineation observed in the same area but agrees with the sense of offset of the high-angle fault planes along the thrusts. A strong axial planar schistosity is present in the D3 thrust slices in the north.

The folds in the Bliss Canyon Formation in the upper plate (D3), in the area of Alaska and Bliss Canyons, are thinned on the limbs, with much thicker and less sheared hinge areas and a marked sense of asymmetry. They are also en echelon by around 15° to the strike of the thrusts in the identical fashion as to the north. The overturned limbs are more thinned than the upright limbs, and are also more contorted into second order parasitic folds. Such "fish-hook" structures are not easily classified into any of the normal niches (e.g., class 1, 2 or 3) but combine aspects of class 1 A and C (Ramsay, 1987). They are inferred by Ramsay to form during compressional buckling with rotation of the principal incremental shortening direction in a vertical plane. In order for this geometry to form, the most competent unit (the middle member of the Bliss Canyon Formation, expressing the folds) must also have strata above it (the coarser-grained upper member) that are slightly more competent than below it (the argillitic lower member). These folds have a weakly to moderately developed axial planar cleavage, and also porpoise along strike (almost sheath-style), with gentle to moderate plunges on the fold axes and the development of domes, basins and saddles. Mesoscopic folds associated with these larger features (particularly in the argillitic horizons) are tight to isoclinal, with Z-profiles (looking north), and can be rootless and transposed. The axial planes are at a definite angle (steeper by about 30° to 45°) to the thrusts in this area, instead of parallel as to the north and in the lower plate. All this is

FIG.69

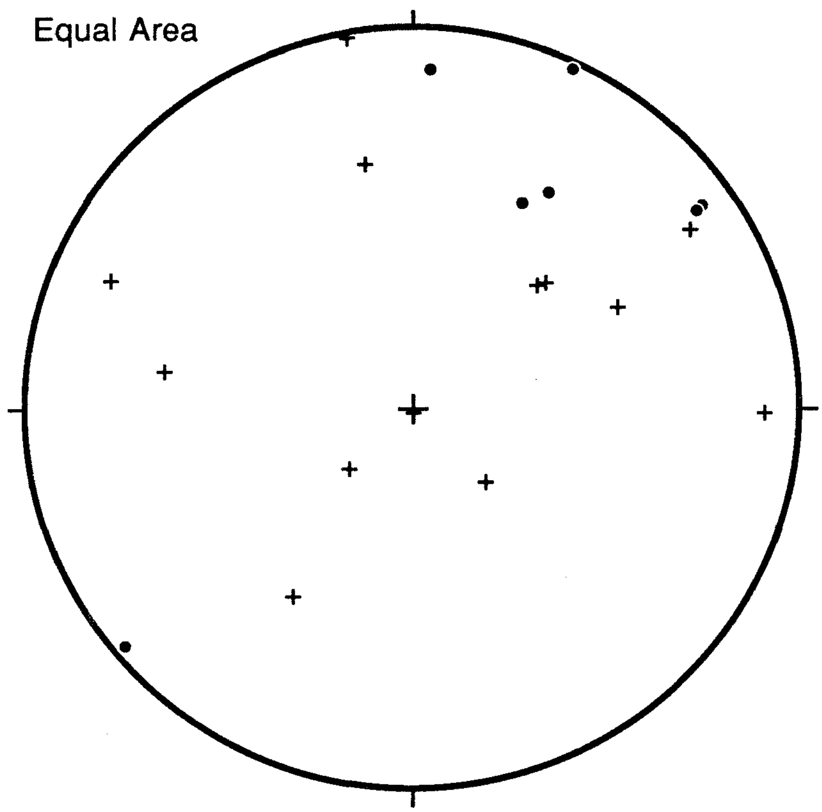


BEDDING - WESTERN THRUST SYSTEM

CONTOURED DATA - POLES TO BEDDING
HEAVY GREAT CIRCLE - MAJOR GIRDLE TO POLES TO BEDDING (280 85N) WITH A POLE (BOX) AT 190 5
GREAT CIRCLE - MINOR GIRDLE TO POLES TO BEDDING (131 47S) WITH A POLE (BOX) AT 41 43

THE MAJOR GIRDLE WAS CAUSED BY DEFORMATION DURING PHASE D3 AND THE LATER STAGE OF D2, AND THE MINOR GIRDLE IS INFERRED TO HAVE BEEN FORMED DURING D1.

FIG.70

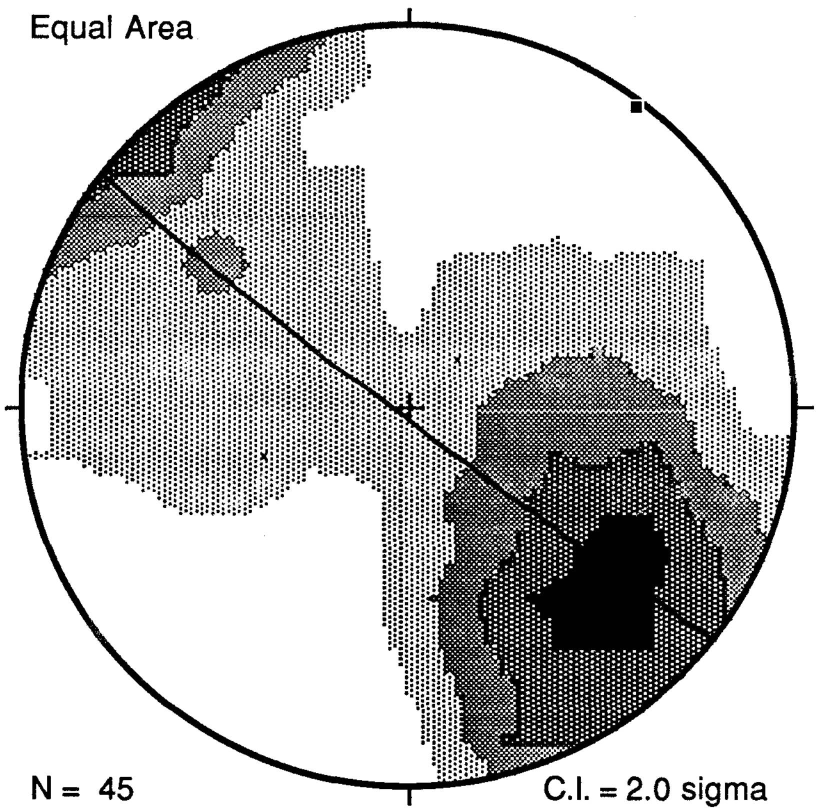


FOLD AXES - WESTERN THRUST SYSTEM

DOTS - MACROSCOPIC FOLD AXES CALCULATED FROM BEDDING MEASUREMENTS
CROSSES - MESOCOPIC FOLD AXES MEASURED IN THE FIELD

MACROCOPIC FOLD AXES ARE FROM D3 AND POSSIBLY D2. THE MESOSCOPIC FOLD AXES ARE FROM D3 AND, IN SOME AREAS, D1.

FIG. 71



WESTERN THRUST SYSTEM - CLEAVAGE

CONTOURED DATA - POLES TO CLEAVAGE
GREAT CIRCLE - BEST FIT GIRDLE TO CLEAVAGE POLES (128 89W)
FILLED BOX - AXIS TO GIRDLE (38 1)

CLEAVAGE FORMED DURING PHASE D3 (AND IN SOME AREAS D1)

D3 phase deformation.

In the area east and southeast of King Lear Peak the King Lear/Happy Creek contact is folded into large upright open folds with planar limbs and fairly angular hinge areas. No overall sense of vergence is evident, as the folds are symmetric and the axial planes vertical. The amplitudes are several hundred meters.

In the McGill Canyon area, the steep southeasterly dips of the macroscopically homoclinal McGill Canyon Formation indicate that this portion of the thrust package is where the thrust ramped up through that formation, so that as thrust displacement continued over the next flat, McGill Canyon rocks were rotated to a steep dip. These are D2 features.

In front of and to the east of the thrust belt, the lower Happy Creek Formation north of Alaska Canyon is folded into upright, open and angular parallel folds with amplitudes of a few hundred meters. These folds (a syncline-anticline pair) phase southwards into that divergent splay, and die out to the north. To the east, the last fold limb becomes the east-dipping homocline of the central high-angle fault domain. They could belong to either D2 or D3.

A stereonet plot of fold axes indicates that the macroscopic folds, with horizontal or gently plunging fold axes and gently to moderately westward-inclined axial planes, are consistent and easily attributed to thrusting. The mesoscopic fold axes are much more complex, with an almost random azimuthal distribution and notably steeper fold axes. The inferrance I draw is that some of these mesoscopic structures represent an earlier and partially overprinted generation (see paragraph on separation of phases).

Fabrics The area around the mouth of Jackson Creek has well-developed tectonite fabrics developed in the Bliss Canyon and Happy Creek Formations, associated with the D3 thrusting. Schistosity parallels the thrusts, and consists of very well-aligned biotite and chlorite plates and amphibole laths, though mineral segregation is not present. Remnant sedimentary or volcanic textures are rare (relict plagioclase grains and ghosts of conglomerate clasts), and overprinting is extreme with transposition of bedding. In the microfabrics, quartz and plagioclase grains are in ribbons, deformed in a manner best attributed to pure shear. In some places the amphibole forms a very good tectonite lineation, generally with a steeply down-dip rake. To the south, the grade decreases rapidly and the foliation becomes an axial planar slaty cleavage, significantly steeper than the thrusts. This foliation often crenulates bedding, and is a slip cleavage. Further south, the axial planar cleavage is only locally present, and an earlier and angularly discordant cleavage and fold generation has been overprinted (see paragraph on separation of phases). In this area, reduction spots in a slate in the McGill Canyon Formation are flattened. However, crinoid columnals in calcarenite beds have undergone both pure and simple shear, with the result that the columnals are both flattened and asymmetrically deformed. This fabric postdates the olistostromal emplacement of the blocks. A stereonet plot of poles to foliation in the western thrust domain indicates that the foliation by-and-large was formed by D3 thrusting, as the poles define a girdle at 128° and vertical - almost normal to the thrusts - and a strong concentration centered about 129° 27°. This latter means that these axial planar cleavages are consistently steeper than thrusting by about 30°. In detail, the overall trend of the cleavage, as with the folds, is at an angle to the trend of the thrusts (about 20° clockwise), supporting the conclusion of sinistral oblique-slip during the D3 phase.

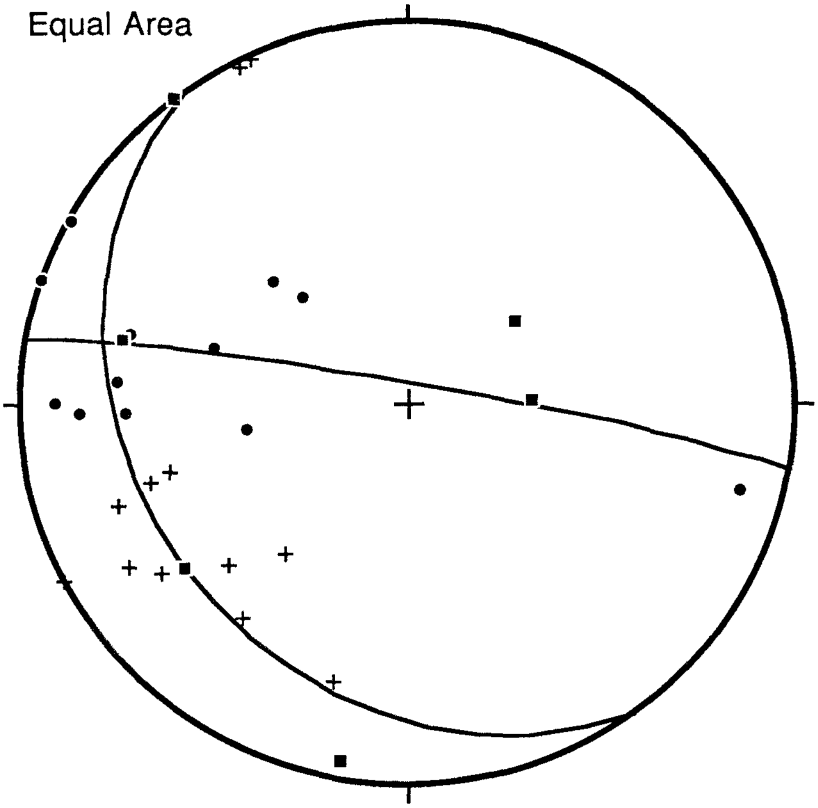
Metamorphism As discussed in the chapter on metamorphism, the hanging wall within the thrust domain is, in the north, extensively metamorphosed to biotite schist and locally to amphibolite grade. This heating event was not genetically related to D3 thrusting; instead when thrusting took place the rocks had been extensively heated by nearby igneous intrusion (regional contact metamorphism) and so were deformed while very hot, resulting in the tectonite fabrics. Some static overprinting of the schistosity indicates cooling continued after thrusting had ceased, though no retrograde metamorphism or dynamic overprinting occurred. The fabrics in these metamorphic tectonites were formed during but not solely because of the compressional orogeny. This heating event and the resultant fabrics die out rapidly to the south through chlorite grade rocks, and the McGill Canyon Formation is of very low grade (zeolite to prehnite-pumpellyite).

Separation of Phases In the region of McGill Canyon Formation exposure, south of the large high-angle fault, an earlier generation of deformation is visible. In the field it is characterized by steeply plunging to vertical open to tight folds with steeply inclined to vertical axial planes, and with a local axial planar cleavage. Both of the latter are often northeast trending. Large areas of vertical pencil argillite in the lower member of McGill Canyon Formation are due to these fabrics. These folds are refolded by the horizontal (fold axis) and moderately to gently inclined (axial plane) folds, already mentioned, and cut by its axial planar cleavage. This early generation of folds is related to the family of generally east-west trending sinistral strike-slip microfaults.

Plots of fold axes and poles to bedding and cleavage (fig. 72, 73, 74) in the McGill Canyon Formation help clarify the situation greatly. For both bedding and cleavage poles, the D2 thrust generation fabrics are clearly defined, but a second well-defined fabric exists. In both bedding and cleavage, removal of the D2 girdle at 190° to 200° and near vertical leaves a distinct earlier generation. This D1 girdle is shallow in both cleavage and bedding quite shallow (164 22W and 144 29W, respectively), and implies a very steep fold axis plunging to the northeast (74 68 and 54 61, respectively). A plot of fold axes in the same area indicates that this population of very steeply northeast plunging folds has been refolded during thrusting about an axis trending northerly and subhorizontal. This is probably D2 and not D3 phase thrusting, in the McGill Canyon area, for reasons discussed later.

These very steep northeast-plunging fold axes agree with the sense of slip on the high-angle fault (oblique, with both sinistral and south-side down offset). Field relations indicate they formed before the principal thrusting (D2) took place nearby. The sense of dip-slip component recorded in the stratigraphy of member I of the King Lear Formation for this fault is south-side down, and predates either reactivation event. If the average girdle axes for bedding and cleavage are rotated back about bedding in the McGill Canyon Formation, they plunge steeply to the northwest (288, 70; 317, 67) and fall directly on the trend of the fault (something the unrotated orientations did not do - see fig. 75). This

FIG.72

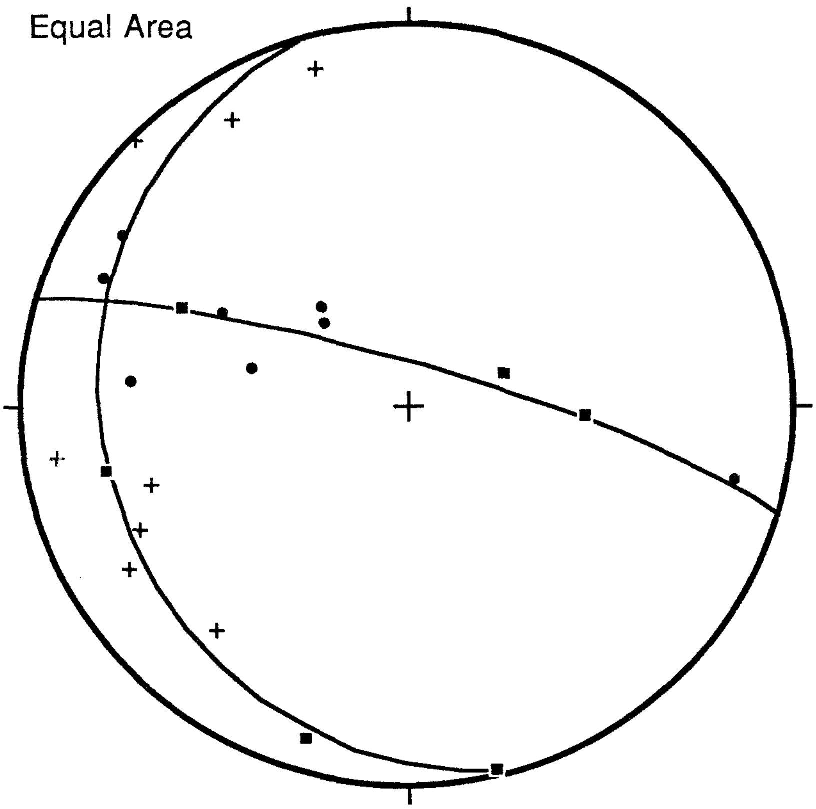


MCGILL CANYON FORMATION - BEDDING

DOTS - POLES TO BEDDING IN D3 PHASE STRUCTURES (GIRDLE AT 280 85N, POLE TO GIRDLE AT 190 5)
CROSSES - POLES TO BEDDING IN D1 PHASE STRUCTURES (GIRDLE AT 144 29W, POLE TO GIRDLE AT 54 61)

LOCALLY TIGHT VERTICAL FOLDS OF PHASE D1 ARE REFOLDED BY HORIZONTAL NE-TRENDING OPEN FOLDS OF D3.

FIG. 73

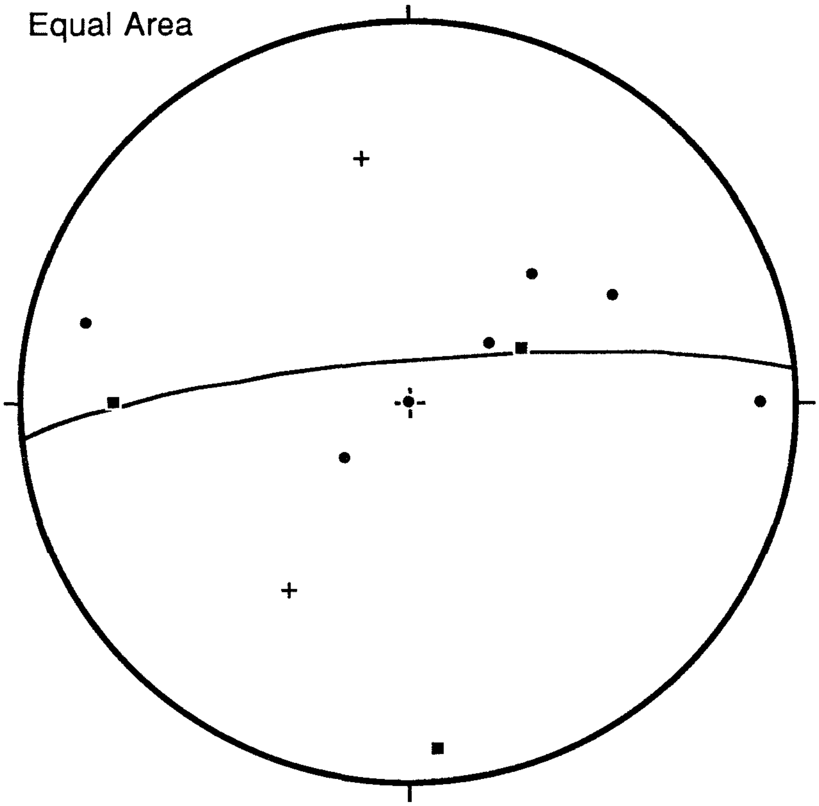


CLEAVAGE FROM THE MCGILL CANYON FORMATION

DOTS - POLES TO CLEAVAGE ATTRIBUTED TO D3 (GIRDLE AT 286 81N, POLE TO GIRDLE AT 196 9)
CROSSES - POLES TO CLEAVAGE FROM EARLIER PHASE (D1?) (GIRDLE AT 164 22W, POLE TO GIRDLE AT 74 68)

THE STEEP D1 PHASE CLEAVAGE IS AXIAL PLANAR TO VERTICAL FOLDS THAT ARE CUT BY THE D3 PHASE CLEAVAGE. THE D1 CLEAVAGE HAS ALSO LOCALLY BEEN FOLDED BY D3.

FIG. 74

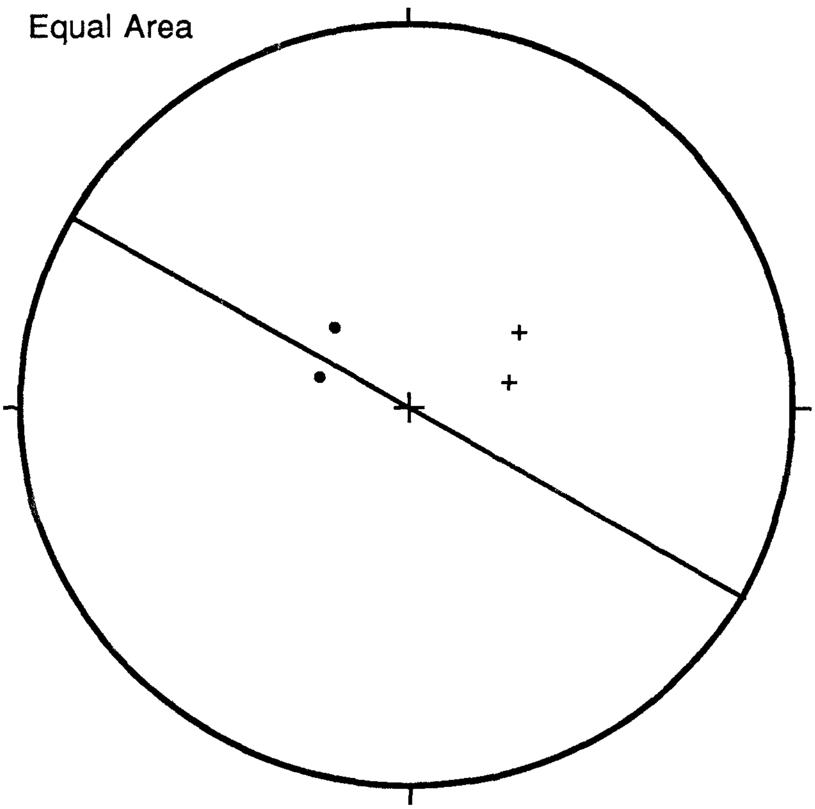


FOLD AXES FROM THE MCGILL CANYON FORMATION

DOTS - FOLD AXES FROM PHASE D1 - ORIGINALLY VERTICAL AND TIGHT, FORMING A GIRDLE (265, 81N) WITH A POLE AT 175 9
CROSSES - FOLD AXES ATTIBUTED TO PHASE D3 (MORE OPEN, NE N TRENDING, AND HORIZONTAL)

THE D1 FOLD AXES HAVE BEEN REFOLDED ABOUT A N-S, HORIZONTAL AXIS DURING PHASE D3.

FIG. 75



EARLY WRENCH-RELATED STRUCTURES IN THE MCGILL CANYON FORMATION

CROSSES - THE OVERALL FOLD AXES (DEFINED BY THE POLES TO THE GIRDLES) FROM CLEAVAGE (74 68) AND BEDDING (54 61)
DOTS - THOSE FOLD AXES ROTATED BACK ABOUT THE OVERALL HOMOCLINAL BEDDING IN THE MCGILL CANYON FORMATION, TO 288 70 (CLEAVAGE) AND 317 67 (BEDDING)
GREAT CIRCLE - THE HIGH-ANGLE FAULT SURFACE, VERTICAL AND TRENDING 120.

THIS EARLY GENERATION OF WRENCH RELATED DEFORMATION IS ATTRIBUTED TO D1.

also indicates sinistral and south-side down offset. As other D1 high-angle faults in this orientation in the range exhibit the same sinistral strike-slip and south side-down dip-slip during their inital phase, it is quite likely that the early phase of deformation in the McGill Canyon Formation discussed here actually took place during D1 wrenching. It would also be odd that, if the steeply plunging fabric was related to slip transfer during thrusting, the fabrics should so clearly predate the thrusting-related elements.

This then is the preferred explanation for this generation, that it evolved early, during D1 phase wrenching. The high-angle fault was reactivated as sinistral tear during D2 east-directed thrusting at a later time, but the fabrics predate this event. During D3, the high-angle fault was not reactivated but was offset. The frontal thrusts here (putting McGill Canyon Formation on Happy Creek, and that on King Lear) are assigned to D2. The reason that it would be rather odd if the high-angle fault (definitely a tear to those thrusts) had been active during phase D3, is that the observed offset along the high-angle fault is sinistral, but given the vergence in the D3 portion of the thrust belt a dextral component would have been predicted. Also, the high-angle/tear fault is offset by a D3 thrust and is not inferred to have been active during D3. The D2 thrust that put McGill Canyon Formation on Happy Creek is inferred to be cut out by a D3 thrust to the south.

Stratigraphic Effects The deposition of the exposed levels of the King Lear Formation postdates both D2 and D3 thrusting in the western domain, with no visible stratigraphic effects.

Cross-Cutting Relationships All intrusive and stratigraphic units in this domain are truncated by the thrusting except the Navajo Peak stock. This latter body intrudes a D2 thrust (the one which puts Happy Creek over King Lear Formation near King Lear Peak), but is in turn truncated on the west by a D3 thrust. The units cut include the McGill Canyon through Happy Creek Formations and the bodies of the Early Mesozoic Intrusive Suite (the Parrot Peak diorite, Harrison Grove granodiorite, the Bliss Canyon tonalite, and all the dikes).

Interpretation

Kinematics Good evidence for D1 phase wrenching exists, particularly in the McGill Canyon Formation (whose argillitic facies express deformations better than other lithologies in other units). Judging by the fabrics developed, this phase had substantial but unknown amounts of sinistral strike-slip activity along a fault trending 120°, with the south side-down on the order of a hundred meters.

Good but indirect evidence for D2 phase thrusting was found in the western thrust domain, in that it is the best explanation for the discontinuous frontal thrust to the lower sheet containing the McGill Canyon and Happy Creek Formations. A more eastern east-vergent shallow thrust also attributed to that phase is present near King Lear Peak. This is probably a frontal splay to the higher D2 thrust fault. Both only occur south of the high-angle fault north of King Lear Peak, which was used as a sinistral tear. Thrusting in D2 was dip-slip and locally directed towards an azimuth of about 110°.

The western domain is dominated by D3 east-directed thrusting along a fault zone trending 210°. The thrusting was oblique with a sinistral sense, with vergence at about 135° (at 15° to the trend of the thrusts). The fault planes were fairly steep, verging on reverse faults with dips between 30° to 65°, when Tertiary tilting is removed. The event probably did not take long, as rocks heated regionally by unseen intrusions were still hot enough to statically overprint themselves after thrusting had ceased. It is quite likely that the D3 thrusting reactivated and completely overprinted D2 thrusts. This would explain why oblique slip took place.

The shortening during D2 was 6 km or greater around King Lear Peak (50%), diminishing to only 500 m just south of Jackson Creek and dying out north of that. In the D3 portion of the western thrust domain, shortening is difficult to estimate because of the lack of sufficient marker beds, but near Alaska Canyon it is a minimum of 3 km (and 200%) as determined from the folding within the Bliss Canyon middle member. Total shortening is probably on the order of 6 to 8 km total in the exposed portion of the belt.

<u>State of Stress</u> For the D1 phase wrenching, the state of stress inferred is the same as that discussed earlier for the high-angle fault domain as a whole. Why thrusting during D3 was oblique is not clear; probably earlier structures from D2 were reactivated during a slightly different state of stress throughout the system. Using the fold axes as the best indicator of overall slip, it would appear that $\acute{U}_1$ during D3 was at about 135° and horizontal, $\acute{U}_2$ at 45° and horizontal, and $\acute{U}_3$ vertical. This is similar to that inferred for D2, though D3 is rotated clockwise approximately 25°.

<u>Timing</u> D1 wrenching probably overlapped in time with the deposition of

member I of the King Lear Formation, as it has affected sedimentation in that unit. D2 postdated member I and predated the Navajo Peak stock. From geochronologic information elsewhere in the range, D2 is late Middle Jurassic to Late Jurassic in age. D3 is not easy to pin down - it crosscuts everything in the western domain. It therefore must be Early Cretacous or later. However, from the western thrust and high-angle fault domains, relationships in the King Lear Formation discussed earlier do pin down the D3 event in the eastern domain as occurring in the late Early Cretaceous (synsedimentary thrusting in strata of that age), and more specifically to overlap the date 115 m.y. (from flaser deformation of a dacite sill of that age).

<u>Summary</u> D1 phase wrenching in the Middle Jurassic left an imprint on the McGill Canyon Formation, and affected member I of the King Lear Formation. This wrenching was sinistral with the south-side down, along high-angle fault trending 120°. Limited east-vergent D2 folding and thrusting (in the late Middle Jurassic) reactivated the high-angle fault as a sinistral tear fault, placed the McGill Canyon Formation over the Happy Creek, and the latter unit over the King Lear. Extensive oblique thrusting in D3 (overlapping with an area of intrusively heated wall rock in the north) formed the main framework for the western thrust domain, with reactivation of earlier D2 thrusts with a slight sinistral oblique-slip character.

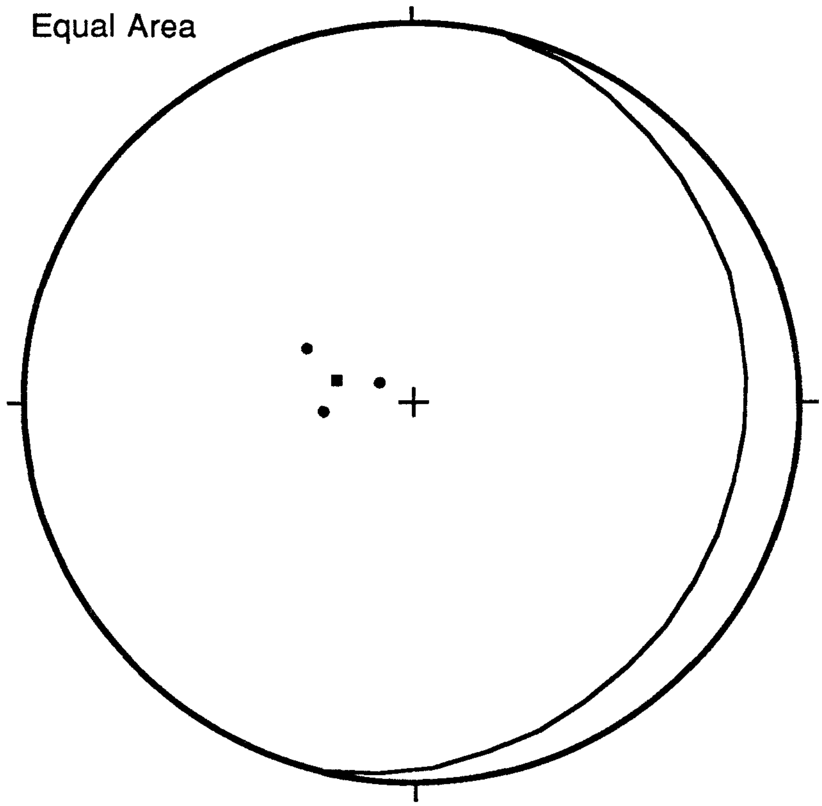
<u>Basin - Range Deformation</u>

In two places - between Bottle Creek and Buff Peak on the extreme eastern flank of the range, and along the extreme western flank of the range due west of King Lear Peak - the thrust planes from D2 and D3 were reactivated in the Tertiary. In both cases, dip-slip normal offset is present, of only modest amounts on the order of a few hundred meters. These are not major range-bounding faults. The D3 fault west of King Lear Peak has been active in the Holocene, as a visible and not highly degraded scarp indicates. The D2 fault near Bottle Creek offsets the Tertiary basalt and rhyolite succession, and connects to a large (dextral?) tear fault, which cuts across the northern extension of the Trout Creek Spur just north of the Niebuhr mine. This fault separates members J and M of the King Lear from a thick Tertiary basalt and rhyolite interval. South of this tear throughout the range, the Tertiary interval unconformably overlaps the Mesozoic units. Evidently the range existed with considerable topography when the flows were deposited, as they project updip directly into the range - the strata were therefore laid down along a buttress unconformity of considerable relief. North of the Tertiary tear fault in an area not mapped in any detail, the Tertiary strata are warped into gentle, north-south trending folds, probably related in some fashion to that tear fault. In addition, the fact that the entire range has been gently tilted to the east after the deposition of the lava flows (which are middle Miocene to early Pliocene in age-Langenheim and Larson, 1973) implies a low-angle listric normal fault underneath the range, probably west-dipping (see fig. 76).

Summary of the Structural Evolution of the Jackson Mountains

No evidence exists for an orogenic event taking place between the deposition of the McGill Canyon Formation and the Bliss Canyon Formation

FIG. 76



DOTS - POLES TO BEDDING, UNDEFORMED TERTIARY STRATA
AVERAGE TERTIARY BEDDING ABOUT 20, 15 E (GREAT CIRCLE AND POLE)

(i.e., Late Permian to early Middle Triassic), as no event is expressed in the older unit that is not present and accounted for elsewhere in the range. Soft-sediment deformation related to submarine slumping took place in both units (plus the emplacement of olistostromal blocks in the McGill Canyon), but is limited, unsystematic and not regionally signficant in terms of structural evolution.

The first significant phase of deformation is the wrench-related phenomena of D1. This consists of two sets of conjugate high-angle faults, oriented at about 70° and 125°. The northeasterly set are subordinate, and have oblique slip with the north-side down and dextral offset. The southeasterly set is dominant, and also has oblique slip with the south-side down and a sinistral sense. The subordinate set is antithetic and the dominant one synthetic in an inferred wrench system, with $\acute{U}_1$ at about 100°. If a through-going master strike-slip was ever established in the area (it is certainly not exposed in the Jackson Mountains), it would trend about 140°. The system had pronounced synsedimentary affects, strongly influencing sedimentation in member H of the Happy Creek and members I and J of the King Lear in several fault-bounded basins. In places a related spaced fracture cleavage is well-developed. The McGill Canyon Formation in particular was deformed along vertical axes during this phase in the western domain. The event is constrained to have occurred in the time period of approximately 175 m.y. to 165 m.y., based on cross-cutting relationships with dated igneous units.

D1 was followed immediately by, and even overlapped in time with (though separated geographically at any instant), an episode of thrust faulting - D2. In the eastern domain, the thrusts trend 15° and dip gently to moderately east. Thrust planes anastamose with one another, both laterally and down-dip. A simple imbricate stack of thin plates resulted, with a basal decollement inferred but not exposed. The first stage of thrusting was slightly oblique with a dextral sense, and the second and dominant stage was pure dip-slip west-directed thrusting. Substantial folding within the thrust packages also took place, with axial planar cleavage developed. There is also related east-directed thrusting in the western domain in the McGill Canyon area and south and east of King Lear Peak. In several areas, the high-angle faults were reactivated as slip transfer surfaces between thrusts, with a sense independant of their earlier history. No signficant regional metamorphism is associated with the D2 event in this area.

The appearance and evolution of the eastern part of this domain can be traced in the stratigraphy of the King Lear Formation, as the McGill Canyon and Bliss Canyon Formations were tectonically exposed, eroded and witnessed as the cherty/lithic detritus in member J (Jurassic in age). The uplift and reworking of the Happy Creek is evident in upper member I and member K. The thrusts are overlapped by member J, and are also plugged by a 162 m.y. stock. They cut sills in the Happy Creek dated at 170 - 175 m.y., and volcanic flows (member G) dated at 169 $\pm$ 10 m.y. Their period of activity was therefore between about 170 and 162 m.y. in the eastern part, though the thrusts near King Lear Peak in the south of the range could have been active later. The transition from D1 to D2 was caused by the switching of the Ú$_2$ and Ú$_3$ stress axes - the early stage of D2 may represent the transition period.

After a substantial hiatus in time, there was a second phase of thrusting. This event, D3, reactivated the frontal portion of the imbricate thrust system established during D2 in the eastern thrust domain, and broke several new splays. These shallowly dipping frontal thrusts breached the depositional paleosurface of the King Lear Formation, depositing and overriding their own sedimentary fault scarp breccias. They also caused soft-sediment deformation in the uppermost King Lear (member M), and flaser deformation in dacite sills. Near Buff Peak, the active slip surface jumped back in the stack to reactivate an older thrust, truncating some D2 elements. In this slip transfer area, the King Lear was extensively deformed into northeast plunging folds. Elsewhere, some of the high-angle faults were reactivated as tear faults once again. In the western thrust domain, D2 thrusts were reactivated extensively by phase D3, with oblique thrusting with a sinistral sense. As the area had very recently been heated by extensive nearby intrusive activity, the thrusting took place in the northern part of the domain under very hot conditions, resulting in biotite and hornblende schists with tectonite fabrics. The rocks were still hot when heating ceased, as there is a mild late static overprint. In the western domain, these D3 thrusts truncate the other geological elements. The phase was active very late in the sedimentation of the King Lear (the late Early Cretaceous), and overlaps the age 115 m.y. (the age of that warmly-deformed sill with flaser textures). The stress axes are apparently rotated slightly clockwise from those controlling D2 and D1.

Tertiary normal faulting was minor, and reactivated some of the outward dipping, flanking thrusts on either side of the range, as well as tilting the range as a whole to the east.

CHAPTER NINE:

SUMMARY GEOLOGIC HISTORY OF THE JACKSON MOUNTAINS, NW NEVADA

The oldest stratigraphic unit exposed in the Jackson Mountains is the McGill Canyon Formation, divided into lower, upper and olistostromal carbonate members. The lower member is Mississippian to Leonardian (late Early Permian), and consists of a 2000 m thick hemipelagic basin plain to slope offlap sequence, containing a submarine turbidite fan complex. The fan is of the radial, sandy type. The provenance is recycled orogen, lacking a substantial volcanic arc component. The transition to the 1000 m thick upper member (late Leonardian to early Late Permian) is marked by the appearance of andesitic volcanism, the increasing abundance of calcarenite tempestites, shoaling of facies to distal shelf environments (above storm wave base), and the disappearance of the turbidite fan except for feeder channel facies. The upper member is also characterized by the inclusions of sedimentary slide blocks of the olistostromal carbonate member. This latter unit is the lateral equivalent of the lower part of the upper member, and consists of Early Permian shallow carbonate platform deposits and cross-cutting andesite dikes. The basin represented by the McGill Canyon Formation was restricted and anoxic, normally- to over-supplied, stable and long-lived. It was situated next to a complex margin with both carbonate platform and volcanic arc features, as well as being a source for orogenic debris. All of these features taken together are indicative of a back-arc basin setting. There is no evidence for an orogenic event between the end of deposition of the McGill Canyon Formation and the beginning of sedimentation in the

Bliss Canyon Formation.

The Bliss Canyon Formation overlies the McGill Canyon with possible disconformity; the basal contact is not well-exposed in the Jacksons. This formation is divided into lower, middle and upper members. The lower member is a little less than 1 km thick, and of Ladinian to middle Norian age. The member is composed of hemipelagic basin plain and submarine clastic turbidite fan facies, with a component from a nearby carbonate bank as well. The provenance for the clastic turbidites was a recycled orogen. The member passes laterally and upwards into the middle and upper members, or into the Happy Creek Formation. The middle member is a very shallow marine carbonate platform facies, early to middle Norian in age and up to several hundred meters thick. The upper member consists of shallow shelf and lagoonal to littoral facies, of middle to late Norian age and also several hundred meters thick. The source was the same as for the lower member clastic turbidites. The facies architecture is of a southeast-facing offlapping margin, with a carbonate bank prograding southeastward over shallow, but anoxic, basinal deposits and followed in turn by a clastic shoreline. The basin nature was similar to that of the underlying McGill Canyon Formation, though more restricted. No volcanic contribution is seen except for at the very top, at the conformable and gradational but sudden transition into the Happy Creek Formation.

The Happy Creek Formation very abruptly overlaps the facies architecture of the Bliss Canyon, along a contact that is regionally characterized by the presence of about 30 m of volcanic epiclastic detritus. In places this narrow zone contains interbedded Bliss Canyon lower or upper member deposits. The Happy Creek is a thick subaerial construct of basaltic andesite to dacite volcanic flows and epiclastics. It has been subdivided into members A to H (oldest to youngest), which total between 2 and 4 km in thickness. Members A, E, F, and G are basaltic andesite, andesite, andesite to dacite, and andesite flows respectively, the latter three with limited interbeds of alluvial epiclastic facies. The formation is dominantly andesitic. Member B is made up of very shallow marine to intertidal deposits (carbonate and hyaloclastic), interfingering with A in the north of the range. Member A is conspicuously augite-phyric. Members C and D are laterally extensive alluvial fan and channel facies overlying member A, and overlain in turn by E. Member F overlies E with some local erosion, weathering and unconformity. Member G in particular is composed of several distinct volcanic constructs, with paleoslope angles around 10°. The volcanic pile of the Happy Creek as a whole thins to the east (exclusive of sill injection) and especially to the south. Member H (andesite flows and fault scarp breccias) was deposited in active basins bounded by a conjugate set of high-angle oblique-slip east-west wrench faults that developed late in the history of member G. This wrench event is phase D1. The Happy Creek Formation extends from the late Norian (204.7 $\pm$ 13.6 m.y. for member A) to Bathonian (169 $\pm$ 10 m.y., from G). It is intruded by stocks dated at 187 $\pm$ 2 (the Harrison Grove stock intruding A, C and E) and 173 $\pm$ 14.3 m.y (the Happy Creek and Parrot Peak bodies, intruding A, B, D, E, F and G), and 170 - 175 m.y. sills (diorite, intruding G). Members G and H also contain xenoliths of these intrusives. D1 structures cut the Harrison Grove stock and the diorite sills, but are cut by the Happy Creek and Parrot Peak stocks. D1 faulting therefore lasted from about 175 to 165 m.y.

The intrusive bodies are part of the Early Mesozoic Intrusive Suite in the range. Compositionally the stocks range from gabbro (Navajo Peak) to diorite (Parrot Peak) to diorite, monzodiorite and quartz diorite (Trout Creek Spur) to quartz monzodiorite and granodiorite (Harrison Grove) to granodiorite and granite (Happy Creek) to granite (Deer Creek Peak) and to tonalite (Bliss Canyon and Trout Creek). Related dioritic, granitic to granodioritic, and tonalitic dike families are present, as are very extensive dioritic and gabbroic sills and laccoliths. The stocks and sills are comagmatic with the volcanic flows of the Happy Creek Formation and the limited silicic volcanic rocks in the King Lear Formation. Limited contact metamorphism to albite-epidote and hornblende hornfels conditions exists around several of the more mafic bodies. The igneous geochemistry is highly indicative of an island arc to slightly back-arc setting (my own unpublished data, not reported here).

Large-scale magmatic trends are present in the igneous rocks of the Jackson Mountains: (1) From mafic to silicic and rather alkalic (members A, E and F and the Harrison Grove, Happy Creek and Deer Creek Peak stocks) from 210 m.y. to about 180 m.y. (2) From intermediate to mafic and then to silicic, but less alkalic overall (member G, the Parrot Peak stock and the diorite sills to member H and lower J and the gabbro sills to upper member I and the Trout Creek Spur and Trout Creek stocks), over the period about 175 m.y. to 160 m.y. (3) Dominantly silicic, and K-poor (the Jackson Tuff, members L and M). The Navajo Peak body is inferred to be from the initial, more mafic stage of this third phase.

The second magmatic episode apparently ended about the same time as or just after phase D2, though the more mafic part of this second episode overlaps in time with and may be related to D1. The third magmatic episode finished at or just before D3. The compressional deformational phases appear to punctuate between the magmatic episodes or to overlap with their most silicic stages.

The King Lear Formation is Callovian to Aptian in age, and is divided into members I through N (in order of first stratigraphic appearance). Member I (volcanic epiclastic) and lower J were deposited while D1 phase wrenching and basin formation was still active. The D1 faults are then overlapped by the Jackson Tuff. Member J records the first influx of orogenically-produced cherty/lithic molasse from the D2 thrust system, being initiated to the east and south at this time (and therefore overlapping in time with D1). D2 thrusts are intruded by the 162 $\pm$ 1 m.y. Trout Creek Spur stock, by tonalite dikes, by the undated Navajo Peak and Trout Creek stocks, and are locally overlapped by member J (Upper Jurassic to Lower Cretaceous). D2 therefore took place between about 160 m.y and 170 m.y. Interdigitation of upper member J and member K (also epiclastic) record activity in the thrust belt, though whether due to D2 or D3 generation uplift is not clear. Three separate volcanic centers are present in the King Lear. Two are rhyolitic domes of member L with flows, epiclastics, and pyroclastics, near Buff Peak and the lowlands of Jackson Creek and placed in member L. The third is a dacitic laccolith complex of member M north of King Lear Peak, with cogenetic epiclastics, lahars, pyroclastics, and flows and is defined as member M. The laccolith is 115 $\pm$ 1 m.y. A second phase of thrusting (D3-

reactivating D2 structures) is seen in the stratigraphic record as well, particularly in member N (sedimentary fault scarp breccias overridden by the D3 thrust front along the paleosurface). This system was active during the upper part of member M, in the Aptian (and perhaps earlier as well).

The conjugate high-angle faults and structural basins of D1 are present down the center and west side of the range. The subordinate northeast fault set was dextral and north-side down; the dominant southeast set sinistral and south-side down. This phase is consistent with left-lateral wrenching along a southeast-northwest trend, with $\acute{U}_1$ at about 110° and $\acute{U}_2$ vertical. Locally D1 deformation (in this case at deeper stratigraphic levels, in the Mcgill Canyon Formation) is extensive with vertical folding (strike-slip dominated). Elsewhere there is monoclinal warping and development of a spaced fracture cleavage, around Jackson Creek and nearer the paleosurface (dip-slip dominated).

On the east side of the range is a west-directed imbricate thrust stack, built in D2 and reactivated in D3 (only the frontal, western portion) with out-of-sequence thrusting. D2 had an early dextral oblique thrust stage and a later dip-slip thrust stage. The early stage might be related to the rapid transition from a D1 wrench to a D2 thrust stress state as $\acute{U}_2$ and $\acute{U}_3$ switched axes. A fold belt formed during D3 in front of and under the frontal thrust, including syntectonic member N. Some D1 faults were reactivated as tear (slip transfer) faults during both D2 and D3. In the north of the range, major D3 offset jumped southward from a reactivated D2 thrust further east and higher in the stack to a new D3 splay to the west and more frontally located. Around that transfer area, a major fold belt formed in the upper Happy Creek and King Lear and acted as a slip transfer zone.

On the west side of the range is an east-vergent fold and simple thrust belt, also initiated in D2 but dominated by D3 phase thrusting. During D3 the rocks in the north of the belt were very warm from adjacent intrusive heating, and formed high temperature tectonite fabrics and assemblages in the hanging-wall, overprinting earlier textures. The D3 fabrics are also slightly statically overprinted, as cooling continued after thrusting ceased. Again, D1 faults were re-utilized as tears during D2, but had had a substantial earlier wrench history. D2 thrusts were preserved in front of the D3 thrusts, in their footwall.

Miocene basalt and then rhyolite volcanics overlapped the Mesozoic units and structural patterns. Considerable topography was still present in the range and the Tertiary strata were deposited along a buttress unconformity of significant and irregular relief.

Basin-Range normal faulting reactivated an east-dipping D2 thrust along the northeast flank of the range, offsetting the Tertiary beds as well as the older assemblage. Similarly, a west-dipping D3 thrust was reactivated along the southwest flank of the range, with Quaternary offset evidenced by a well-preserved fault scarp. The entire range was tilted around 15° to the east, implying perhaps a west-dipping listric normal detachment surface under the range.

Lake Lahontan shoreline features are abundantly present along an elevation of about 4300 feet at the foot of the western flank of the Jackson Mountains. Within the range, Quaternary fluvial and landslide deposits are present within most of the stream valleys.

CHAPTER TEN:

TECTONIC INTERPRETATION OF THE GEOLOGY OF THE JACKSON MOUNTAINS

A great deal can be deduced about the large-scale tectonic setting and evolution of the Jackson Mountains from the details of this study, which is the purpose of this chapter. In order to reach this goal, I will treat the macroscopic stratigraphic settings and trends presented by the formations cropping out in the range, together with the phases and nature of structural and magmatic evolution, and will synthesize the tectonic environment and history (see fig. 77, 78, 79). A detailed summary of the geologic development within the range was given in the previous chapter. Regional tectonics and correlation will be covered in chapter 11.

On a large scale, the Mississippian to early Late Permian McGill Canyon Formation records the infilling of and margin offlap into a marine basinal terrain. The basin was rather restricted and anoxic, with hemipelagic sedimentation. The sedimentation rate was typical of a eugeocline (arc-related basin - Schwab, 1976). This and the offlap facies geometry with accompanying shoaling and regression indicate the basin was normally- to oversupplied with sediment (Brown and Fisher, 1977). An unusually high degree of stability in the palogeography is inferred from the long life and fixed position of the small submarine turbidite fan. Such basins are typically small, located in tectonically active areas, and partly confined. Andesitic flows and crystal tuffs in

FIG.77

KEY TO LITHOLOGIC SYMBOLS AND EXPLANATION

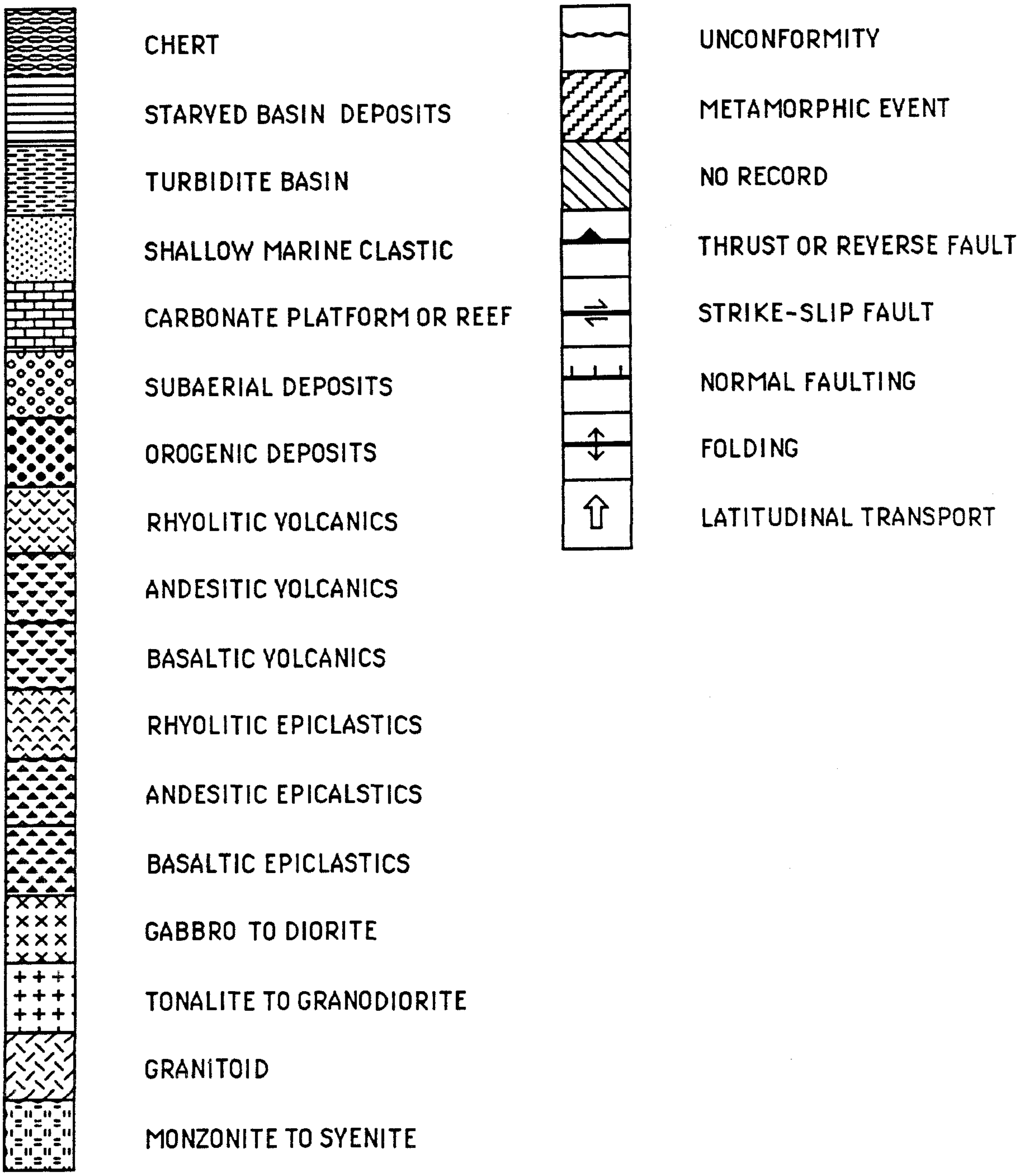
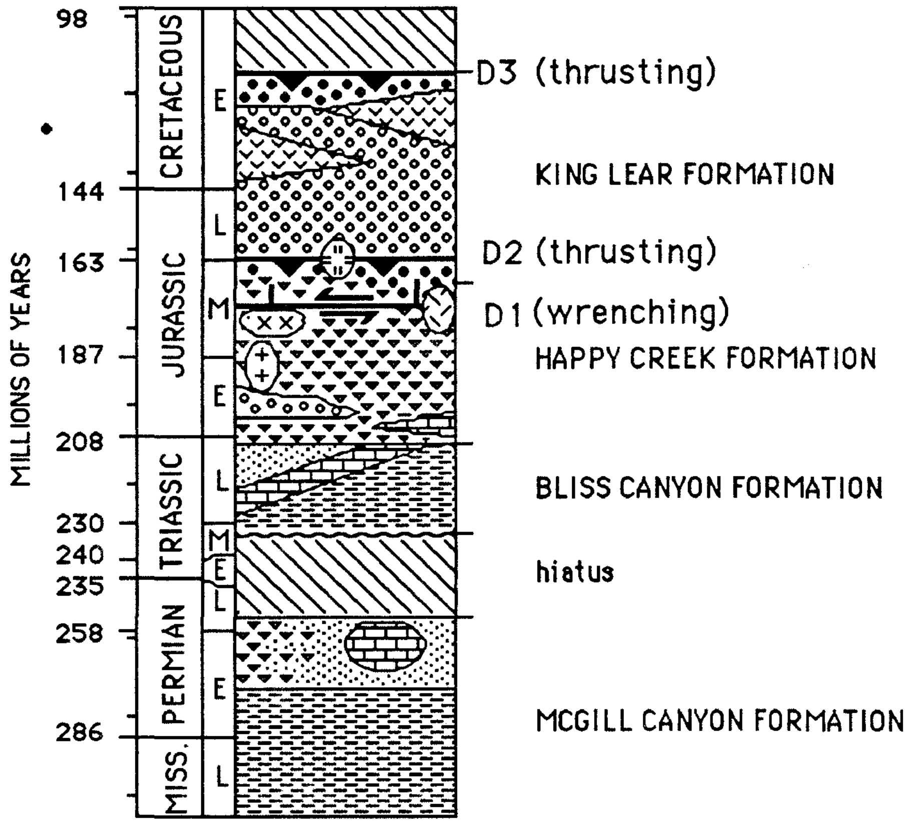


FIG.78

TECTONOSTRATIGAPHIC SUMMARY OF THE JACKSON MOUNTAINS



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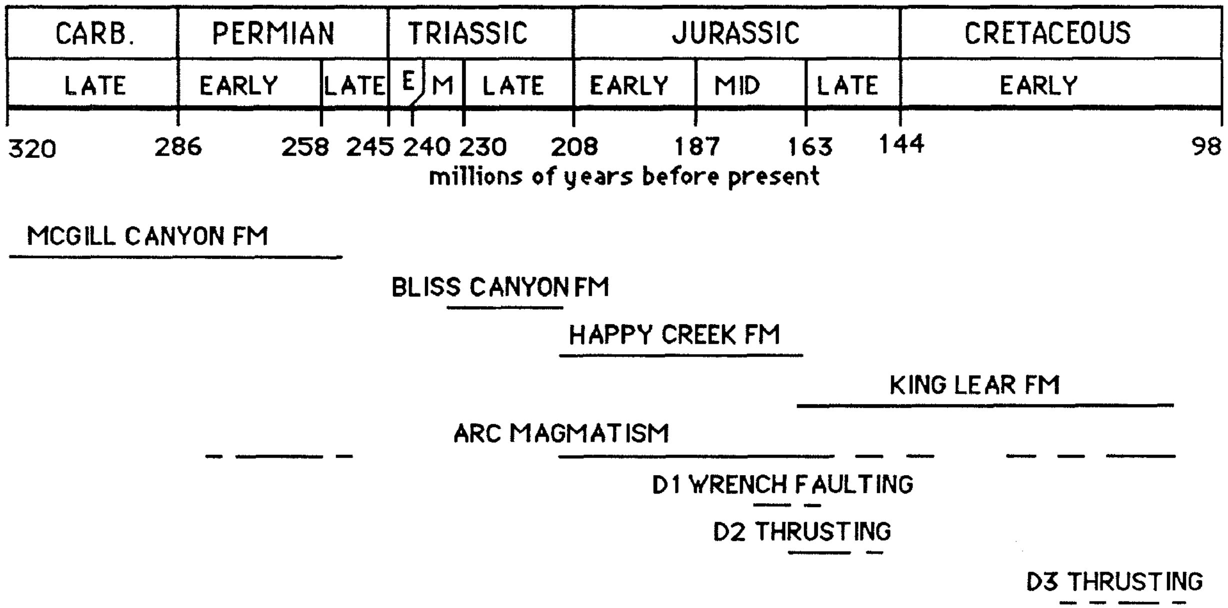
MAHER AND SALEEBY (1986, 1988)
MAHER AND KIRSCHVINK (1987)
MAHER (THIS WORK)
RUSSELL (1981, 1984)
WILLDEN (1963)

FIG. 79

TIMING OF PHASES IN THE GEOLOGIC EVOLUTION OF THE JACKSON MOUNTAINS, NW NEVADA



the section indicate proximity of the basin to an active volcanic terrain as well, particularly in the late Early Permian, but the sedimentary provenance in the clastics is dominated by a recycled orogen containing rocks of Middle Paleozoic age. The nearby development of an extensive carbonate platform is also evident in the section. The overall tectonic setting is inferred to be a long-lived back-arc marginal basin, on the continent side of the marginal arc, away from the subduction zone and the main axis of volcanism. The long life is evidenced by the persistence and lack of change in the basin through time and by the age of the orogenic source terrane within the arc. This is a somewhat unusual characteristic for such an arc-related setting, but this fringing arc was an old and complex tectonic feature (Miller, 1987). That the margin was east-facing is also inferred from the regional facies patterns in rocks of this age within the Black Rock Desert surrounding the Jackson Mountains, particularly the Pine Forest Range (see chapter 11). The size of the basin cannot be evaluated from the Jackson Mountains alone.

The Ladinian to middle Norian Bliss Canyon Formation is inferred to rest on the McGill Canyon Formation with conformity to disconformity in the one area of possible exposure at Trout Creek Spur (as is the case just to the north at Quinn River Crossing - Jones et al., 1988). No significant orogenic event separates the two, and the Bliss Canyon was deposited in a setting very similar indeed to the McGill Canyon. The Bliss Canyon Formation records once again regional offlap, passing upwards on the west side of the range from a restricted and anoxic basin to carbonate fore-reef slope to very shallow carbonate platform to back-platform lagoonal and then to littoral and perhaps even fluvial facies.

The shoreline and carbonate platform were prograding to the east and south, and did not reach the eastern half of the range before the basin system was disrupted by Happy Creek volcanism. The Bliss Canyon Formation has lower sedimentation rates than the McGill Canyon, more representative of a continental shelf setting, and lacks evidence of syndepositional volcanic activity anywhere in the neighborhood.

The basin was also quite shallow (a few hundred meters deep in the uppermost lower member, from the size of the amygdules in the basal Happy Creek flows and from above-storm-wave-base sedimentary structures). It is inferred that the arc evident in the McGill Canyon strata had shut itself off before the deposition of the Bliss Canyon Formation (perhaps during the distant Sonoma Orogeny), and was not being actively uplifted and stripped. This left the basin to be slowly filled instead by erosion of the passive arc terrane to the west (and perhaps with a source in the North American continent to the east, but this contribution is not evident in the Jackson Mountains). The provenance is the same for the McGill Canyon and Bliss Canyon Formations. Definitely, there was continuity of the physiographic basin and of the sourceland between the two formations, even though deposition in the basin might have stopped or the strata that were laid down in between been removed. Some minor subsidence within the basin might also separate the deposition of the two units. The Bliss Canyon thus represents the east-facing back-arc margin of a passive fringing or island arc complex. Extreme stability for an active margin setting is evident once again.

An abrupt change in the tectonic setting is documented by the late Norian to Bathonian Happy Creek Formation. Everywhere, this unit overlies the complex facies geometry of the Bliss Canyon abruptly but conformably, in both basinal and arc-margin environments. A major pulse of mafic volcanism covered the entire area. The area was also shoaling so that even locations situated in the basin plain during Bliss Canyon sedimentation rapidly became subaerial shortly after. Regional uplift and progradation is therefore indicated. Only at the very base is the Happy Creek inferred to be marine. Even the lower Happy Creek flows are interbedded with tidalites and littoral hyaloclastite deposits, and those higher in the section have paleosols and alluvial lenses. Several distinct subaerial andesitic stratovolcanic centers and constructional piles are visible in the stratigraphy of the formation. From this time on, the area of the Jackson Mountains has been emergent.

Subvolcanic stocks and laccoliths and cogenetic feeder dike swarms are also present, in the Early Mesozoic Intrusive Suite. The axis of arc activity had established itself near the Jackson Mountains at this time. Several distinct episodes of arc igneous activity took place, generally trending from more mafic to more silicic, and declining in K_2O-content with each episode. These episodes covered the intervals of approximately 210 to 185 m.y., 180 to 160 m.y., and from somewhere in the Late Jurassic or early Early Cretaceous to 115 m.y. Geochemical analysis (not treated in this thesis in any detail) definitely indicates an origin with a subduction-related arc system for all Mesozoic igneous lithologies (my own data; Russell, 1981), which is fully compatible with the petrology (see chapter six).

Intra-arc deformation was also a significant characteristic of the Middle Jurassic to Early Cretaceous tectonic setting. The stability evident before in the Bliss Canyon and McGill Canyon Formations had been replaced. The first phase (D1) involved wrench faulting with a probable sinistral sense from 175 to 165 m.y., as concluded from the formation of local fault-bounded wrench basins along a conjugate system (sinistral synthetic and dextral antithetic shears) (Harding, 1974; Reading, 1980). A transtensional setting within the arc is concluded; overall compression is inferred to have been slightly south of east and the wrench zone perhaps northwest-southeast. This wrench event overlapped in time with Happy Creek volcanism (which became more mafic during it, and then largely ceased).

The wrenching continued into the deposition of the basal King Lear Formation with growth fault relationships. These growth faults were overlapped by molasse shed from the Bliss Canyon and McGill Canyon Formations as they were exposed in a west-directed thrust belt to the east and south (D2). Some synchroneity of high-angle faulting on the west and thrusting on the east is indicated by the deposition of the first pulse of molasse while the high-angle faults bounding the basins were still active, and one of the major high-angle faults in Jackson Creek continued activity throughout D2 thrusting and King Lear deposition. In addition, sills and laccoliths in the eastern range also indicate a compressional stress state in that area during intrusion (Williams and MacBirney, 1979), although they are themselves cut by the high-angle faults. Also, at the same time in the western part of the range, only stocks and dikes were intruded - characteristic of a neutral to extensional stress regime. The geographic and structural complexity of deformation and the abundance of volcanic activity are indicative of an intra-arc setting (Cross and Pilger, 1982), locally wrench-dominated. It should be made clear that the development of a wrench system may or may not imply the development of a through-going strike-slip system.

The first stage in this episode of thrusting (D2) was noticeably oblique (dextral), and may represent a transition between local transtension- and transpression-dominated phases within the intra-arc wrench system. From the calcalkaline to slightly alkaline geochemistry, this is inferred to have been located to the continental side of the main axis of volcanism. The transtensional and transpressional regimes were coaxial in stress, with the minimum stress axis becoming vertical in the later phase and the direction of maximum compression oriented slightly south of east throughout (present coordinates). In many ways this block-faulted geometry is similar to that developed in the more landward areas of the California borderlands in the late Mesozoic and Cenozoic (Howell et al., 1980). The exceptions are that the California borderlands are a forearc and not intra- or back-arc phenomenon, and also that in the Jackson Mountains, wrenching preceded instead of followed compression. Intra-arc wrenching in areas of oblique convergence is also a common occurrence within the complex southeast Pacific region (Huchon and Le Pichon, 1984; Kimura, 1986; Maung, 1987). This area has been suggested as a suitable tectonic model for the Mesozoic North America Cordillera (Saleeby, 1983; Silver and Smith, 1983).

The revived arc was long-lived, with several discrete pulses of magmatic activity, and was apparently located on the North American continental margin in its later phases (after D2). The paleogeographic basin separating the arc and the continent in this area in the Late

Paleozoic and Triassic had disappeared by the Middle Jurassic through infilling, D2 compressional shortening and uplift, and the sourceland now lays to the south and east within that new orogenic province, instead of to the west as earlier.

Arc volcanism in the area of the Jacksons waned after the end of deposition of the Happy Creek Formation, but continued at low levels throughout the King Lear Formation, increasing again in the upper part of the formation with the appearance of several small silicic volcanic centers. Some thrusting may also have persisted (D3), because the sedimentary record in the King Lear indicates the continued existence of the orogen as a detrital sourceland. D3 phase compression was most active around 115 m.y., and there is no evidence for a wrench-related origin. The stress axes may also have rotated slightly from those of D2, to a more southeast-northwest orientation for the maximum compressive stress. D3 is inferred to have been caused by compression within or behind the continental arc.

CHAPTER ELEVEN:
REGIONAL CORRELATION AND TECTONIC EVOLUTION FROM THE VIEWPOINT OF THE JACKSON MOUNTAINS

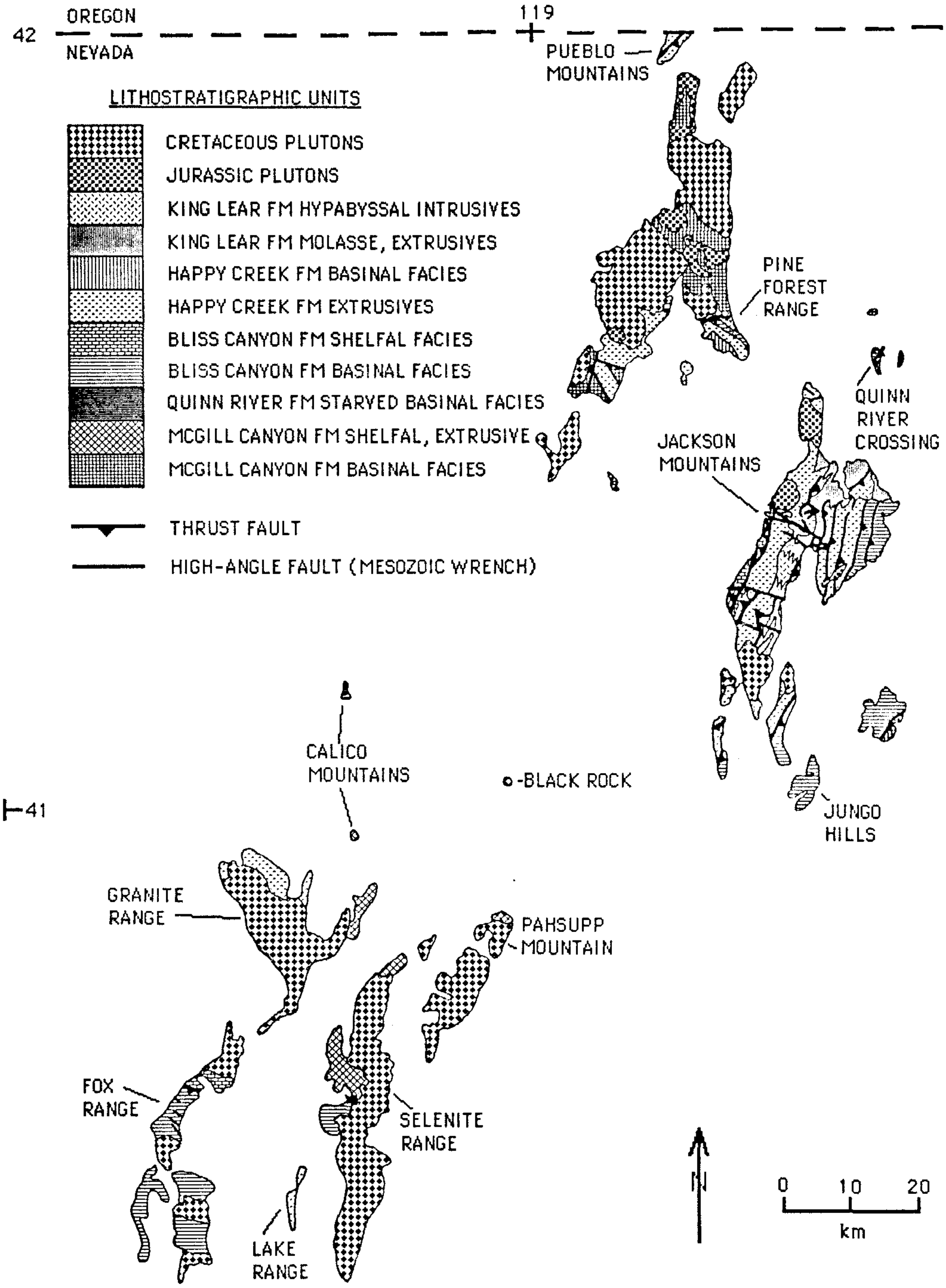
In this chapter, I will analyze the implications of the geology of the Jackson Mountains for regional correlation of stratigraphic units and tectonic phases, and for Cordilleran geologic evolution. In order to do this it is necessary to briefly discuss the geology and development of the nearby related terranes, including the Black Rock Desert (containing the Jackson Mountains, Pine Forest Range, Quinn River Crossing, Black Rock and several other areas), the eastern Klamath Mountains, the pendants of the Sierra Nevada, and various areas in western and central Nevada. An exotic origin is not implied by the use of the term terrane here, but instead a fault-bounded lithostratigraphic assemblage of genetically-related rocks. Following the terrane summary will be a synthesis of regional geologic evolution in the Late Paleozoic and Mesozoic. The geologic and tectonic evolution of the Jackson Mountains has already been summarized in chapters nine and ten.

The Black Rock Desert

The Black Rock Desert area (see regional map, fig. 80) lies in the northwest corner of Nevada, and contains a number of geologically related assemblages. The history of this area will be summarized after discussions of the individual ranges. It is considered a single terrane, with an internally consistent history distinct from neighboring areas,

FIG.80

PRE-CENOZOIC GEOLOGY OF THE BLACK ROCK DESERT AREA, NW NEVADA



from which it is separated by Mesozoic thrust and perhaps wrench faults.

Pine Forest Range During the summer of 1987, a preliminary map of the southeast corner of the Pine Forest Range was compiled (see fig. 81). Previously, the area had been reconaissance mapped by Smith (1973), but the section had been inverted in his work (Silberling, written communication, 1986), following presumably after Willden (1963). The geology shares many features with that of the Jackson Mountains, and individual members can in many cases be correlated between the two areas (see fig. 82). The discussion of the Pine Forest Range will be in considerable detail, to include my own work and because a satisfactory account is not published anywhere else. The area is being mapped as a Ph.D. project by Sondra Wyld of Stanford.

The McGill Canyon Formation as mapped in the range contains units Trp, Pm and Phm of Smith (1973). The lower member (his Trp) contains thin-bedded grey to green-grey argillite and quartzose arenite. The member coarsens and thickens upward toward the top of the section with the argillite disappearing and cross-bedded, very coarse-grained quartzose thin- to thick-bedded arenites and conglomerates appearing. These coarser-grained facies are in clast-support, are well-rounded, and well-sorted. The provenance also shifts from quartzose/cherty/lithic to more arkosic and angular near the top. Limestone lenses tens of meters thick and hundreds of meters long are also discontinuosly present in the upper part of the unit. The current exposed thickness in the mapped area is estimated at 960 m; much more is exposed to the north. The member is very similar to the lower member of the McGill Canyon Formation in the

FIG. 81

PRELIMINARY GEOLOGIC MAP OF THE SOUTHEAST PINE FOREST RANGE, NW NEVADA

FIELD WORK IN 1987 WITH JAY LIESKE

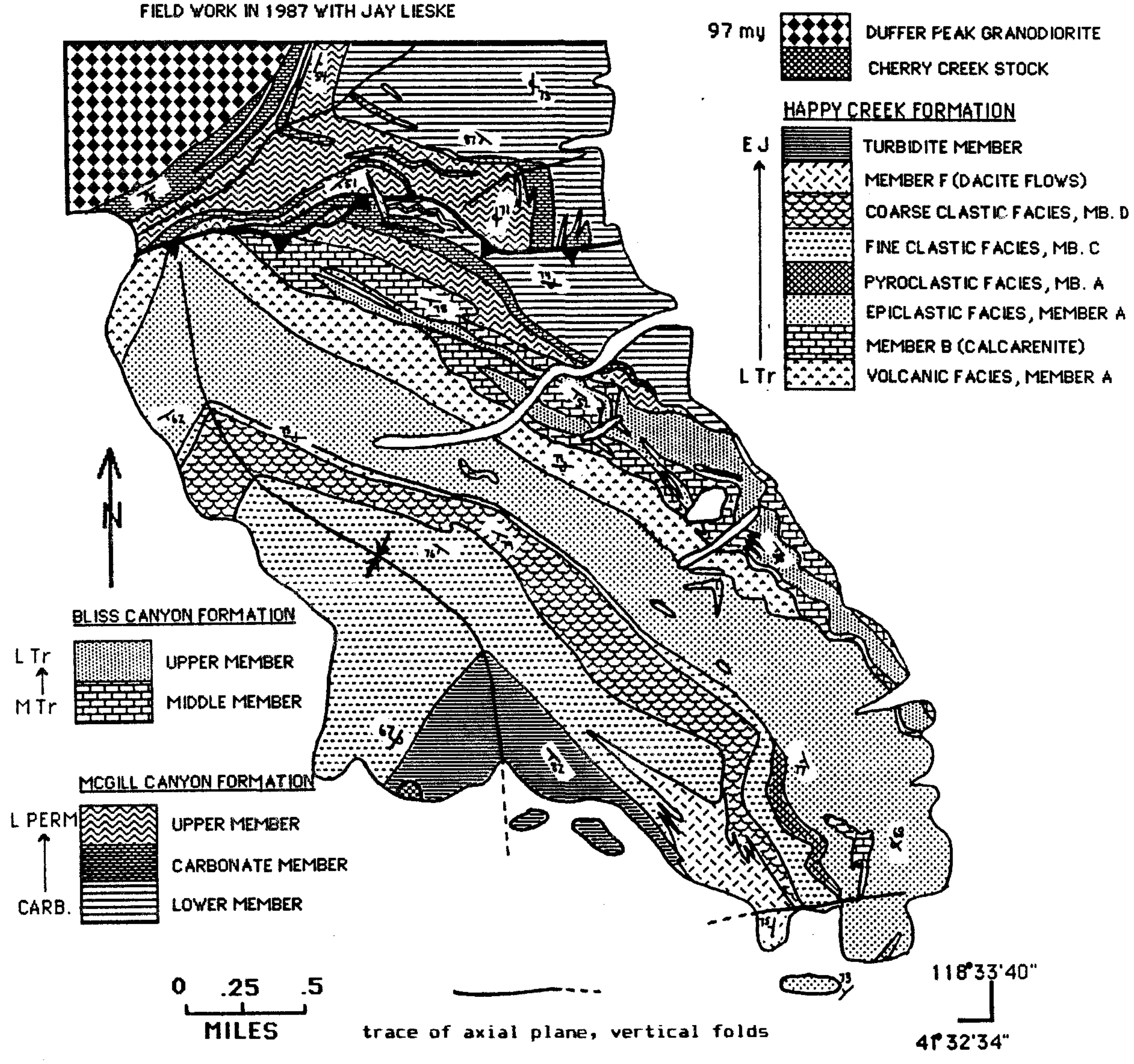
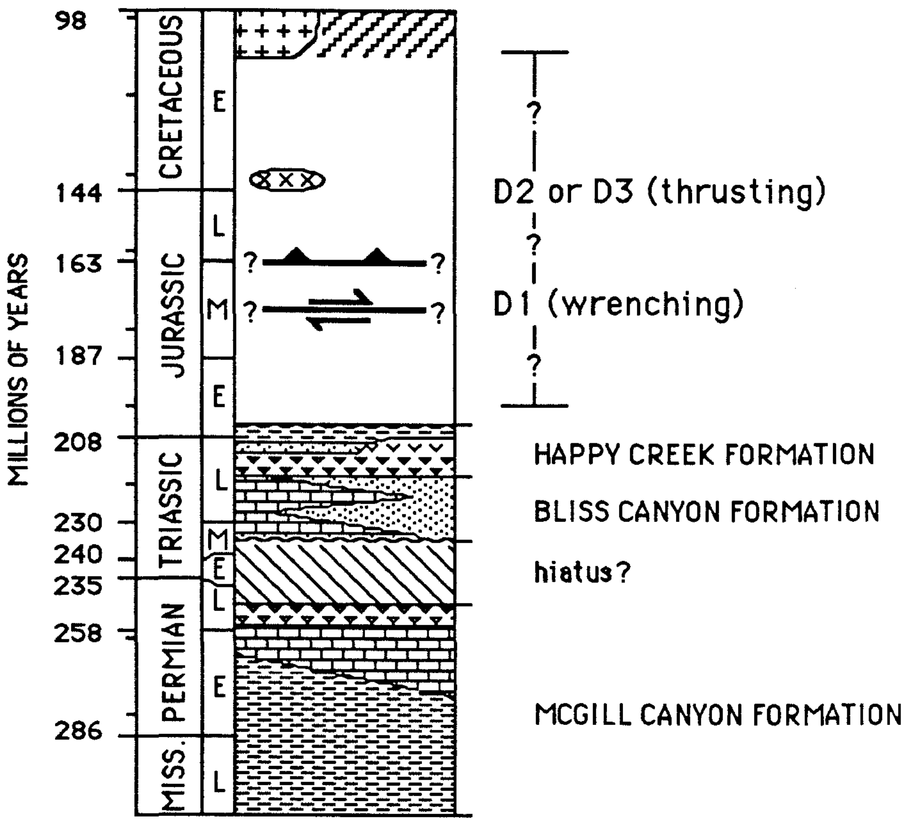


FIG.82

TECTONOSTRATIGAPHIC SUMMARY OF THE SE PINE FOREST RANGE



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N. SILBERLING (UNPUBLISHED DATA)
WILLDEN (1964)
SMITH (1973)
MCDANIEL (1982)
SILBERLING AND OTHERS (1987)

Jackson Mountains. Basinal turbidite conditions shoaling upward to shelfal above-storm-wave-base environments with calcarenite mounds are concluded for the member.

Overlying the lower member is a 760 m thick (structural thickness) sequence of interbedded epiclastic and carbonate lenses - the upper and carbonate members (Phm and Pm of Smith). The carbonate member consists of lenses and very thick beds of a megacrinoid calcarenite (a biosparrudite or grainstone) where depositional textures are preserved. Commonly it is sheared and recrystallized to a fine- to medium-grained foliated light grey marble with graphitic laminae and floating deformed crinoid stems. The crinoids have undergone simple shear, rotated into the foliation plane with slip and rotation between the individual disks. Locally the limestone is very black and rich in organic material, with fish teeth, bone and scale material - almost a phosphatic bone-bed (conodont separation was attempted and failed - D. Clark, written communication, 1988). Carboniferous to Permian corals and crinoid columnals are present in this unit (Silberling, written communication, 1986).

The upper member of the McGill Canyon is somewhat more variable. It is characterized by containing volcanic and epiclastic facies, particularly in the lower part. These facies include green hornblende and/or plagioclase phyric and green aphyric andesite flows (containing crinoid molds), white silicic ignimbritic and crystal tuffs, arkosic volcanic arenites (some cross-bedded), thin limestone and grey chert (tuff?) beds. The volcanic facies become less common in the upper part of the member, where thin limestone beds in a matrix of quartzite and phyllite predominate. Very high in the section and just beneath the thrust the upper member contains a thin bed of brachiopod biosparrudite, almost undeformed (Thamnosia depressa, of an early Wordian, earliest Guadalupian, early Late Permian age - B.R. Wardlaw, written communication, 1988). The valves are closely packed, unbroken, paired, and in growth position. Another similar bed of solitary rugose corals is present slightly lower in the section (Paleozoic, but otherwise unidentifiable - C. Stevens, written communications, 1988). Both of these beds are inferred to contain preserved shallow marine communities, and the megacrinoidal limestone of the carbonate member lenses are also interpreted as shallow marine banks.

The tectonic setting of the McGill Canyon Formation is inferred to be within or on the back-arc margin of a sporadically-active island arc edifice. The detailed stratigraphy, age and provenance is surprisingly close to that of the unit in the Jackson Mountains, though perhaps slightly shallower and more proximal to the volcanic arc source.

The Bliss Canyon Formation (including Trl, Trc and Trls of Smith) has the upper (Trc) and middle (Trl and Trls) members but not the lower member present, and is about 670 m thick. It overlies the McGill Canyon Formation with apparent conformity, but the upper member of the McGill Canyon may disappear underneath it to the south (due either to unrecognized deformation, facies pinch-out, or erosion). The upper and middle members of the Bliss Canyon interfinger and are present as lenses in one another to an extent not seen in the Jackson Mountains; the labels are not entirely appropriate here but the correlation to the units in the Jackson Mountains is clear. The middle member is characterized by thick-bedded light grey crinoidal calcarenitic limestones (biosparrite or grainstone). The crinoids are often distinct pentagonal to star-shaped forms, dominantly Early Mesozoic (H. Lowenstam, oral communication, 1987) and in any case are considerably smaller than those in the Paleozoic limestones. The limestones contain skeletal fragments, including the Upper Triassic clam Minetrigonia (from the top of the formation), bryozoa, and other Upper Triassic shelly fauna and early Norian conodonts (N. Silberling, written communication, 1988). Intraformational flat pebble conglomerates (rip-ups of micrite and calcarenite) are found in most of the less-deformed limestone exposures (often at the top-shoaling sequences?), and are diagnostic of very shallow marine settings, especially intertidal or supratidal. The intraclasts are flat and rounded, .5 to 1.5 cm in size and very well-sorted. Lenses (fluvial distributary channel deposits?) of conglomerate with a calcarenitic matrix and argillite and quartz clasts are found in the limestone lenses as well. I interpret the limestones as shallow marine mound accumulations, but N. Silberling (written communication, 1988) has proposed that they are instead allodapic in origin (basinal and turbiditic).

The upper member facies is dominantly thinly interbedded to laminated medium to dark grey argillite and fine-grained quartz arenite. The thinner beds near the base are sometimes normally graded, and there is some bioturbation. Intraformational flat-pebble conglomerates (with argillite, quartzite and limestone rip-ups) are also present, in clast-support, and indicate an origin above storm-wave-base. Imbrication in the conglomerates dips westward. Exotic clasts (vein quartz, plutonic and especially extraformational chert) are rare but present. The conglomerates also are rounded and in clast-support or are diamictites with wacke matrix-support (proximal debris flows); they typically are present as discontinuous lenses or thin beds. Some siltstsone beds have scoured bases, and are cross-laminated. The environmental interpretation of this member is not unambiguous, but in conjunction with the middle member with which it is intimately interbedded I favor a shallow marine, shelfal setting. More particularly, a clastic shelf above storm wave base, with tempestite deposits, scattered bioclastic mounds and fluvial distributary channels. The alternative, that the sequence is basinal, and that both the limestones and coarser-grained clastics are turbidites and the finer-grained clastics hemipelagic, cannot be ruled out but I consider significantly less likely. As in the Jackson Mountains, no volcanism is expressed in the Bliss Canyon Section.

The Happy Creek Formation overlies the Bliss Canyon with a sharp but conformable contact; some interbedding takes place at the contact. The Happy Creek as expressed in the Pine Forest Range is clearly correlatable but also clearly distinct from the exposures in the Jackson Mountains, especially in the upper portions. Member A is the basal unit, and consists of several distinct and mappable facies. The lowest facies of A is the identical green augite-phyric basaltic andesite seen throughout the Jackson Mountains, and has been dated by Rb/Sr geochronology at 204.7 $\pm$ 13.6 m.y. (R. Kistler, written communication, 1988). Coarsely vesicular and pillowed flows and flow breccias with a glassy or limestone matrix are abundant in the facies, with minor cross-laminated volcanic arenite beds (hyaloclastites?). The pillows are elliptical, .5 to 1.5 m and have glassy rinds and carbonate in the vesicles and interstices. The augite andesite facies is up to 430 m thick, and pinches out to the south.

Upward in the member, an andesitic epiclastic facies become predominant over the flow facies, and also contains clasts of Bliss Canyon Formation limestone, quartzite and argillite, and a diorite phase. The epiclastic facies includes sedimentary breccias, conglomerates and volcanic arenites, and is up to 770 m thick. Augite disappears upward as a phenocryt phase, and plagioclase, quartz and biotite appear in the flows, which are still pillowed. Crystal tuffs and tuff breccias are present throughout A, and in several locations vent facies with lapilli tuffs containing flattened scoriaceous bombs are mapped as a separate pyroclastic facies within the epiclastic facies of A. As well as interfingering at the base with the limestones of the Bliss Canyon Formation, scattered small lenses of identical calcarenitic crinoidal limestone (member B) are present throughout member A.

An extensive clastic unit, lacking volcanic flows, abruptly overlies member A and is defined as members C and D, because of the epiclastic nature and stratigraphic position. However, the members differ substantially in facies and provenance from their presumed equivalents in the Jackson Mountains, and it is at this point that substantial departures in stratigraphic development begin between the two areas. A 20 to 170 m thick lower fine-grained clastic interval abruptly overlies A, and contains early late Norian ammonites, including Rhabdoceras (N. Silberling, written communication, 1986). Thin-bedded grey argillite, quartz arenite and less common interbeds of crinoidal limestone and silicic crystal tuff characterize the unit, placed in C. Cross-to planar lamination, current ripples and lenticular bedding are present and indicate distal shelf marine conditions, as opposed to the subaerial alluvial fan and braided channel deposits in this member in the Jackson Mountains.

This succession is overlain by a coarser-grained interval (placed in member D), and composed of very poorly sorted and thickly bedded graded breccias to conglomerates with angular clasts of intraformational volcanics, silicified argillite, and late Karnian - early Norian limestone slabs (this detritus has an origin in the Happy Creek and Bliss Canyon Formations). The succession is 140 to 530 m thick (and thickens to the northwest). Texturally there is clast-support of the larger slabs, but the smaller clasts can be in matrix support; imbrication is present and the matrix is andesitic epiclastics. Soft-sediment deformation is associated with these beds, particulary in the underlying beds. Some of these slabs are up to seven meters in size. These massive conglomerates are interbedded with the same thin-bedded argillite and quartz arenite facies seen just below it. They are also interbedded in the south of the map area at the top with the dacitic volcanic flows of member F. A submarine origin, probably still on the distal shelf, is inferred for the finer-grained facies. The megaconglomerate facies definitely originated as submarine proximal debris flows into a shelfal setting, with some area of significant relief as a source.

Some major event not seen in the Jacksons caused the development of substantial local tectonic relief, exposure of rocks as low as the Bliss Canyon formation, and the shedding of debris flows into this proximal marine area; the sedimentary facies relationships in this interval are rather like those associated with a wrench-related borderlands tectonic style, such as the San Onofre Breccia, or the Violin Breccia of the Ridge Basin (Link and Osborne, 1978).

The facies in D are overlain by a 600 m thick interval of member C very similar to that immediately underlying it, with the addition of common dacite flows and tuffs and rare crinoidal limestones. The tuffs are graded to cross-laminated and water-lain, and occur in the base of the overlying turbidite member. Cross-lamination, oscillation and current ripples, thin graded beds (tempestites) and very thin planar bedding are present in the lower portion, which fines and exhibits progressively lower-energy sedimentary structures upward, with the addition of thick volcanic arenite beds with scoured bases. A late Late Norian to Early Jurassic(?) ammonite fauna is present in this unit and in the turbidite member at the top of the section (Silberling, written communication, 1986 and 1988). A setting in the distal shelf, above wave-base but deepening rapidly upward in the section to a position below storm-wave-base and on the slope, is inferred.

To the south upper member C is interfingered with and then lenses out completely into a dacite flow unit, put in member F with some uncertainity (some hints of silicic volcanic activity were seen in member C in the Jackson Mountains - this could be equivalent to that, not the stratigraphically higher and perhaps younger member F). The flow nature is clear, as they are seen flowing over and pressing their basal breccia into unconsolidated argillite and quartz arenite, and are present as clasts in the sediments. Individual thin flows can be traced out for hundreds of meters within the stratigraphy, thickening and joining to the south to a maximum composite thickness of 500 m. The flows are light grey and fine-grained, with scattered hornblende and/or plagioclase phenocrysts in a very fine-grained matrix, and with associated flow breccias. This member represents an isolated submarine dacitic volcanic center, in a slope setting. The associated tuffs may indicate subaerial exposure of the volcano nearby.

F and upper C are overlain by a classical turbidite sequence, which has no equivalent in the Jackson Mountains and is placed in its own member. The interval is made up of thin-bedded grey argillite and cross-laminated, normally graded greywacke with rare ripples and flame structures. Thick channels of wacke and arenite with scoured bases and shale chips are interspersed throughout. This is definitely a basinal clastic environment, in the mid- to outer fan, and continues the subsidence started in member C. The interval is at least 820 m thick, with no upper contact. Pronounced and rapid subsidence took place across the Triassic-Jurassic boundary in the Pine Forest Range, as opposed to the uplift experienced by the Jackson Mountains in the same time period. Intra-arc unrest and possibly borderlands-style tectonics might account for this.

Several intrusives - the Duffer Peak granodiorite batholith (96.2 $\pm$ 3.5 m.y - Smith, 1973) and an unnamed small granodioritic stock in Cherry Creek intrude the map area, and there are a number of other intrusive bodies in the Pine Forest Range on the west and north. The metamorphic grade also increases dramatically from south to north. The argillite and quartz arenites of the turbidite member are essentially unmetamorphosed, on the extreme south and in the upper plate. Going north, especially north of Bishop Canyon (the large steam valley shown on the map in the middle), and in the lower plate to the thrust fault, the rocks are phyllitic, then weakly schistose. Some of the pelitic schists in the lower member of the McGill Canyon Formation have apparently developed kyanite and biotite and the limestones tend to be highly recrystallized to marble and calcsilicates; the coarser-grained clastic facies are commonly stretched. Adjacent to the pluton (and to the north of the map area around several other large stocks), the country rock (McGill Canyon Formation) is ductilely deformed into late tight vertical folds with limbs parallel to the pluton margins. The geometry and setting is highly reminiscent of structures seen in the Lake Isabella area of the Sierra Nevada batholith, where Mesozoic sediments have been deformed by vertical flow and stretching during downward movement around rising Cretaceous plutons (Saleeby and Busby-Spera, 1986). I believe the regional metamorphic gradient in the Pine Forest Range and the deformation around the latest Early Cretaceous Duffer Peak granodioritic batholith (which extends to the west and north, and perhaps underneath) is also due to contact metamorphism and intrusive deformation by that body, in a similar fashion to the pendants in the Sierra Nevada.

The structural geology of the mapped portion of the Pine Forest Range is also very similar in important respects to that of the Jackson Mountains, outside the later pluton-related generation. A low-angle thrust fault cuts across the northern part of the map area, with McGill Canyon Formation in the lower plate and the entire sequence in the upper. The McGill Canyon Formation in the lower plate is deformed into several large vertical folds by the later batholith intrusion phase, as described in the previous paragraph. The upper plate (essentially unmetamorphosed, especially to the south) is thrown into a huge vertical synform, opening to the south with an interlimb angle of about 60°, an amplitude of 6 km and an angular fold hinge. Bedding on the limbs is very steep and often vertical to overturned, and the small mesoscopic folds are also typically steeply plunging. This fold is distinct in style from the generation caused by the intrusions. Only along the eastern range front, at low elevations, does bedding become moderate to shallow (and consistently upright). This is inferred to be caused by large scale drag along the cross-cutting underlying thrust fault, with eastward vergence inferred for that fault, which agrees with the sense of offset along it (though offset is rather minor, probably only 1 to 2 km).

The geometry associated with the large vertical synform is very similar to the D1 phase deformation seen in the western thrust domain of the Jackson Mountains, and they clearly are the same event. The later thrust fault, which now dips gently southeast, could be either a D2 or D3 phase thrust. In either case, it has since been uplifted and tilted by the intrusion of the Duffer Peak batholith. The two events in the Pine Forest Range are post-earliest Jurassic (the youngest strata deformed by D1) and pre-97 m.y. (the age of the cross-cutting pluton, which has dikes pinning the thrust). D1 is well-constrained in the Jackson Mountains to the period 165-175 m.y., but the thrust in the Pine Forest Range cannot be assigned to either phase D2 or D3 of the Jackson Mountains with much confidence. I prefer D2, as some of the plutons associated with the intrusive deformation and contact metamorphism to the north of the map area are of latest Jurassic age, and are apparently undeformed (Smith, 1973).

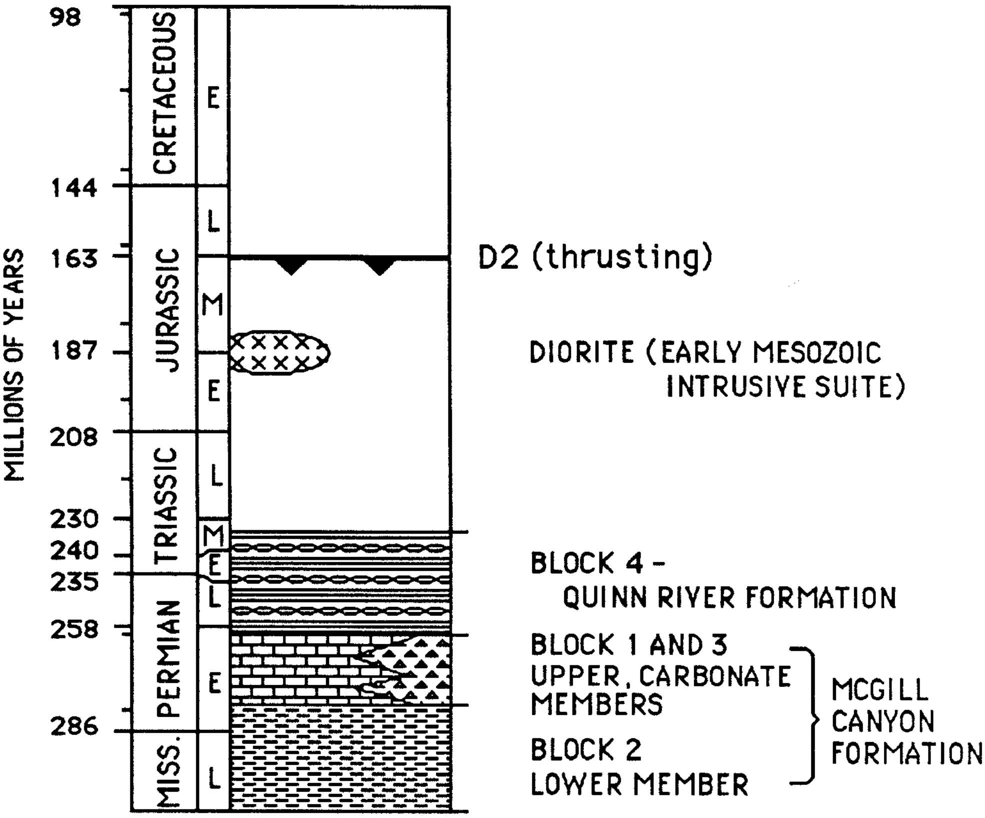
Quinn River Crossing Lying due north of the Jackson Mountains across the Quinn River are some low-lying exposures in the southern Bilk Creek Range, referred to as Quinn River Crossing. Structurally I infer them to lie within the D2 portion of the eastern thrust domain of the Jackson Mountains, as they are directly on strike with that belt. Metamorphism is of low-grade; the conodont alteration index is only 3, as opposed to 5 to 8 in most ranges in this part of Nevada (Ketner and Wardlaw, 1981; D. Clark, written communication, 1987).

The stratigraphy and structure are discussed by Willden (1964), Skinner and Wilde (1966), Ketner and Wardlaw (1981), McDaniel (1982), Silberling and Jones (1982), Silberling et al. (1987), and Jones et al. (1988). There are four thrust-bounded structural blocks, and the thrusts were west-directed.

Block 1 contains cherty and quartzose to volcanogenic and tuffaceous arenite, andesite and basalt extrusives, bedded chert (some of it phosphatic, radiolaria-bearing, and spiculitic), and fossiliferous limestones of middle(?)-late Wolfcampian to early Lenoardian age (see fig. 83). The fauna in these limestones contains rugose corals, brachiopods and schwagerinid fusulinids of McCloud-affinity. The limestones have undergone some slumping, and are clearly shallow marine in origin (on both facies and faunal considerations). The interbedded clastic sediments must also be at least partly shallow marine, though they are described by Ketner and Wardlaw as turbiditic. Limestones are

FIG.83

TECTONOSTRATIGAPHIC SUMMARY OF QUINN RIVER CROSSING



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WILLDEN, 1964

SKINNER AND WILDE, 1966

KETNER AND WARDLAW, 1981

MCDANIEL, 1982

SILBERLING AND JONES, 1982

JONES ET. AL., 1988

more abundant and thicker in the upper part of the block. The block has a structural thickness of 2 km. The clasts of chert contain Early Paleozoic to post Middle Devonian radiolaria, and so the beds are younger than that. The block is intruded by a coarse-grained hornblende diorite with a K-Ar age of 187 $\pm$ 5 m.y., predating the thrusts.

Block 2 is composed of cherty and quartzose arenite and conglomerate (with some metamorphic rock fragments and detrital grains) and is commonly graded and contains coarse-grained channel deposits. The age is uncertain.

Block 3 is made up of bioclastic, thick-bedded fossiliferous limestone more than 500 m thick, with chert nodules commonly developed. The fauna contains McCloud-affinity fusulinds, crinoid columnal, rugose corals and brachiopods of Wolfcampian to late Leonardian age. Conodonts from the block are shallow marine forms, which agrees with the facies interpretation.

Block 4 has an inferred disconformable but concordant contact between Upper Permian tuffaceous radiolarian and spiculitic chert, limestone and dolomite and Lower Triassic tuffaceous and radiolarian phosphatic chert, argillite and arenite. Primitive radiolaria occur at the base of the Triassic, as well as a Spathian ammonite, and there are Guadalupian radiolaria near the base and late Middle Triassic shelly fossils near the top. This is the Quinn River Formation, and it probably has about 200 m of stratigraphy exposed. The Permo-Triassic boundary might be preserved here, which is an uncommon event. Siliceous tuffs also appear in the Middle Triassic portion. No evidence exists in the section for diastrophism associated with the Sonoma Orogeny.

Rocks of comparable age in the Jackson Mountains and Quinn River Crossing are also of very similar lithologies and facies; the stratigraphies are remarkably similar (this includes the Pine Forest Range). The faunal affinites in the Permian fossils also indicate depositional linkage of these areas in that time at least. As the Paleozoic strata at Quinn River Crossing have no formal formation designation, I correlate them directly to the Late Paleozoic units I have defined in the Jackson Mountains.

To be more specific, on this basis I place block 2 in the lower member of the McGill Canyon Formation (although age data on the sequence is missing, its lithologic and facies similarities and probable stratigraphic position are convincing). Block 3 and the carbonate facies in block 1 are inferred to be the carbonate member of the McGill Canyon Formation; this area and/or the Pine Forest Range may have been the source for the olistostromal slide blocks in the Jackson Mountains, especially as some sliding of those limestones is also observed at Quinn River Crossing in block 1 - the section is in place overall, however. The volcanogenic portion of block 1 belongs to the upper member.

The Quinn River Formation of block 4 has no clear correlative in the Jacksons, except perhaps in the Trout Creek Spur exposure (and it also appears to be missing in the Pine Forest Range). The upper part of the formation is just older than (though similar in characteristics to) any of the Bliss Canyon Formation strata. Possibly, strata of this age were never deposited in the shallower marine settings in the other areas (although Silberling et al., 1987, report that there is thin radiolarian chert and argillite unit between the lowest Bliss Canyon limestone and the highest limestone and terrigenous clastic beds of the McGill Canyon Formation, in the Pine Forest Range).

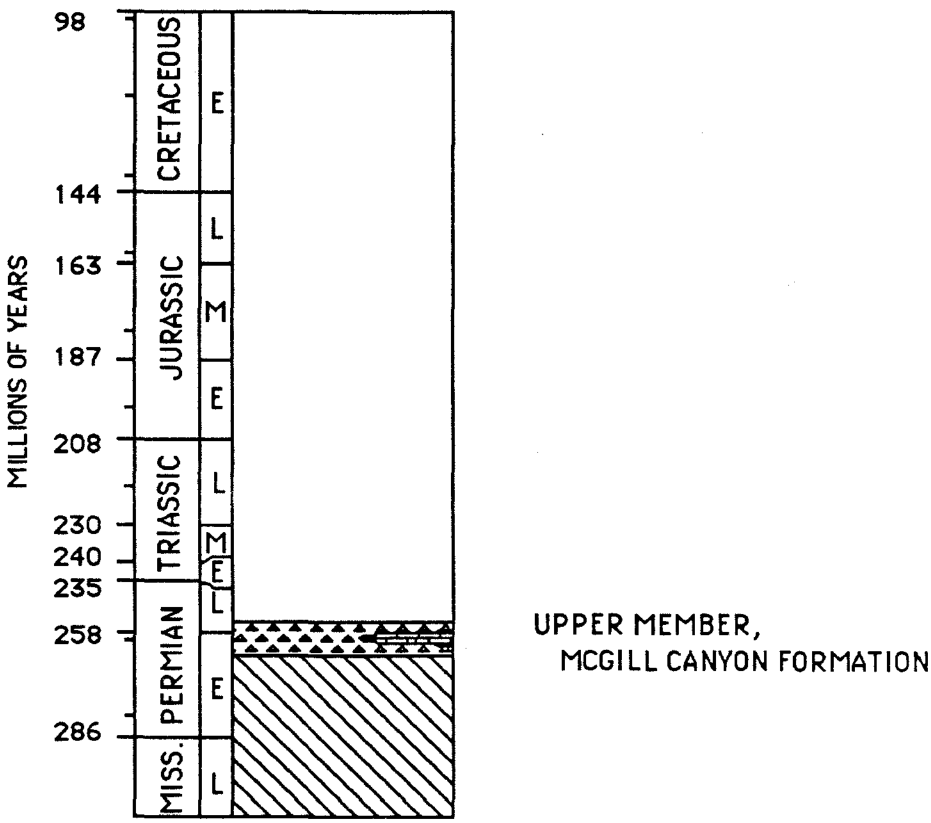
Black Rock This small range in the middle of the Black Rock Desert west of the Jacksons has been described briefly by Gianella and Larson (1960) and Silberling et al. (1987). The range contains about 600 m of fossiliferous, shallow marine epiclastics and limestone with late(?) Permian brachiopods. The section is dominantly shelly andesitic volcanic arenite and breccia, with a 30 m thick limestone horizon (fig. 84).

I place these strata within the upper member of the McGill Canyon Formation, based on their age, provenance and facies. The interval is distinctively coarser-grained and more volcanic/epiclastic-dominated than the exposures of the member in the Jackson Mountains (and presumably more proximal to the arc volcanic source), but is also clearly a closely related lateral equivalent.

Pueblo Range This range lies to the north of the Pine Forest Range, straddling the Oregon-Nevada border, and has been discussed by Roback (1988). The western flank of range contains a thick Late Jurassic sequence of dominantly andesitic flows, and volumetrically subordinate more silicic flow, pyroclastic and epiclastic facies. Comagmatic sub-volcanic stocks intrude the pile. The geochemistry indicates an arc derivation (Roback, oral communication, 1988). The western portion of the range is also largely undeformed. In the eastern portion, the sequence is more distal (less flow and more pyroclastic and epiclastic deposits) and has been metamorphosed to phyllite and schist (greenschist

FIG.84

TECTONOSTRATIGAPHIC SUMMARY OF BLACK ROCK

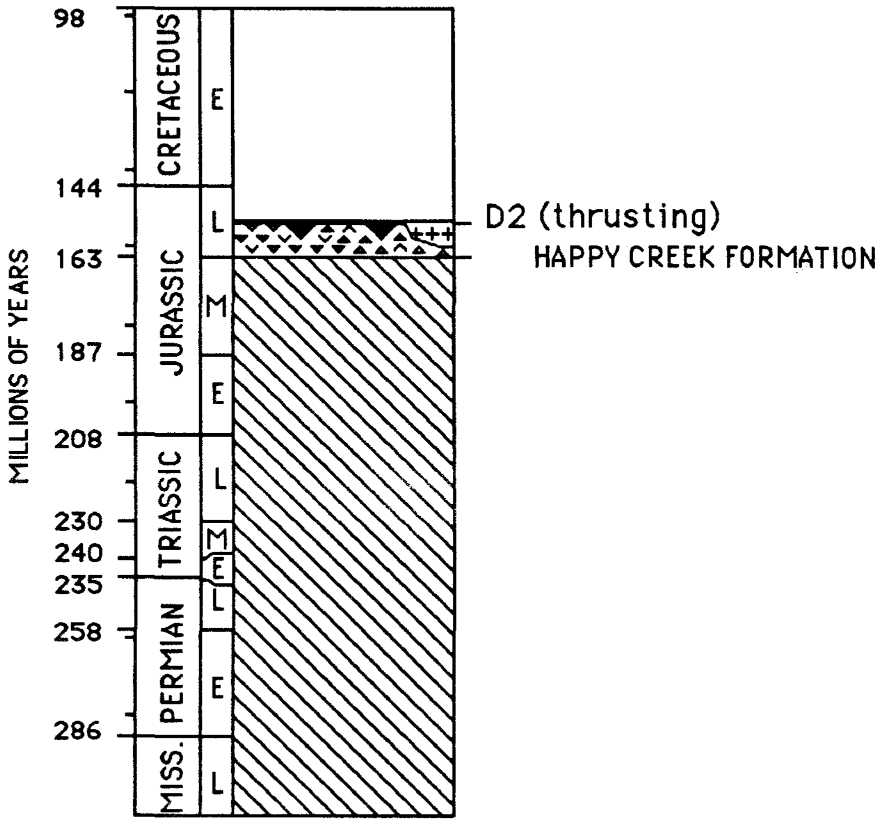


REFERENCES

GIANELLA AND LARSON, 1960
SILBERLING AND OTHERS, 1987

FIG.85

TECTONOSTRATIGAPHIC SUMMARY OF THE PUEBLO RANGE



REFERENCES

ROBACK (1988)

conditions), with the imposition of a strong NE-trending, SE-dipping dynamothermal fabric. Fabrics indicate westward vergence during thrusting. Intrusives in the eastern high-grade belt are foliated but also concordant to the foliation, implying syntectonic intrusion. Roback concludes the thrusting took place around 155 m.y. ago, with some uncertainity due to limited age data (discordant U/Pb zircon and Rb/Sr whole-rock geochronology) (fig. 85).

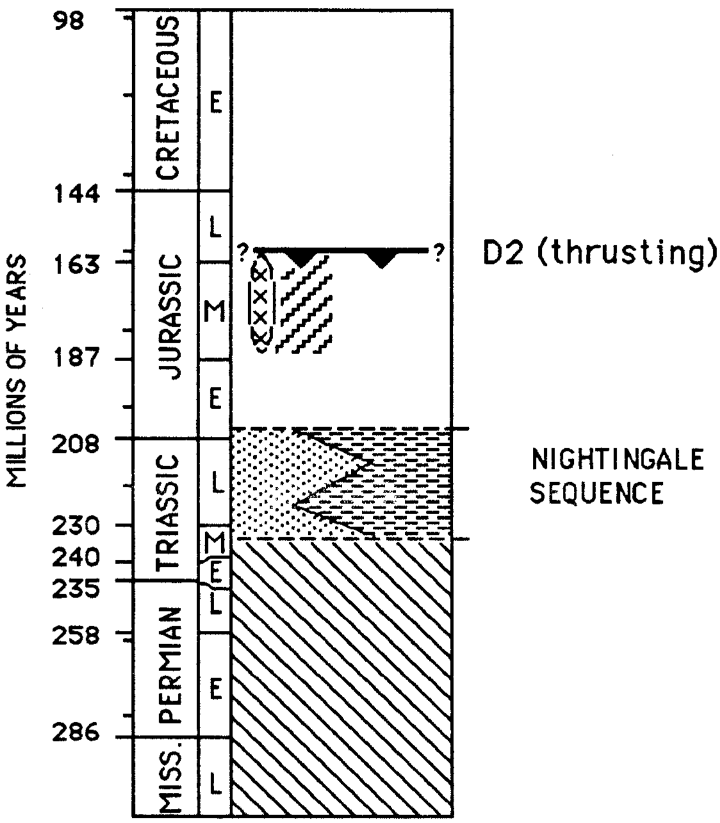
The geology has strong resemblance to the Jackson Mountains, in timing, stratigraphy and structural history. The volcanic pile, by general age and composition, is part of the arc system of the Happy Creek Formation. A member-rank correlation does not seem justified, however, as the sequence in the Pueblo Range is slightly younger and slightly more silicic than that in the Jackson Mountains. This silicic volcanic activity might correlate best actually to the Jackson Tuff Bed and other thin silicic tuffs in member I (the basal King Lear Formation). In that case, volcanism had ceased and orogensis begun in the Jacksons slightly earlier than in the Pueblo Range. This also implies the Jackson Tuff and member I and lower J in the Jackson Mountains are 155 to 160 m.y. in age. Based on current data, the deformational event in the Pueblo Range is also slightly younger (circa 155 m.y.) than in the Jacksons (where some thrusts are plugged by a 162 m.y. stock, although the stratigraphy indicates ongoing orogenic uplift after that), but shares a common orientation and nature; I infer it to also be part of the D2 compressive orogeny. In both places, compressional orogeny and arc volcanism overlapped in time and space.

Fox Range The Fox Range lies at the southern end of the Black Rock Desert, and has been mapped by Kirkham (1982). The oldest exposed rocks are Early Mesozoic metasediments assigned to the Nightingale Sequence, perhaps 2000 to 3000 m thick with neither top nor bottom exposed (fig. 86). Kirkham's lower member (metamorphosed to medium-grade) consists mostly of biotite schist, with thick quartzite and marble beds. This upper member is less metamorphosed, and contains phyllite, thin-bedded quartzite, and impure limestone. The carbonates are thinly to massively bedded (up to 100 m) lenses. The lower member is interpreted by Kirkham as shallow marine, near-shore deposits with clean quartz arenites and limestones with wood fossils(?). The upper member is concluded to have been laid down in a reducing basin environment (I do not find the evidence quoted for this convincing - the presence of carbonaceous material in the limestones, and of pyrite in the finer-grained facies). In the Lake Range, volcanic beds are also found in the upper member. There is a paucity of age data on the sequence in the Fox Range, and in fact the age is based entirely on correlation to the Auld Lang Syne Group to the northwest; the two members may be lateral facies equivalents. The sequence in the Fox Range is intruded by undated quartz diorite (pre-thrust and presumed Middle to Late Jurassic), and post-thrusting cross-cutting granodiorite and quartz diorite (presumably about 90 m.y., although protoclastic deformation may indicate a synorogenic age).

The upper member was placed over the lower along a low-angle thrust fault of uncertain sense (fold asymmetry may indicate westward vergence), in an unusual case of lower grade rocks put over higher grade. The main metamorphic episode is dynamothermal, and in the lower plate and lower

FIG. 86

TECTONOSTRATIGAPHIC SUMMARY OF THE FOX RANGE



REFERENCES

KIRKHAM (1982)

member is related to quartz diorite intrusive heating, syndeformational. Lower grade regional static metamorphism postdates the thrusting and characterizes the upper plate. High-angle Mesozoic faulting also is present, in N-S and E-W families, and is inferred to be part of the D1 wrench system.

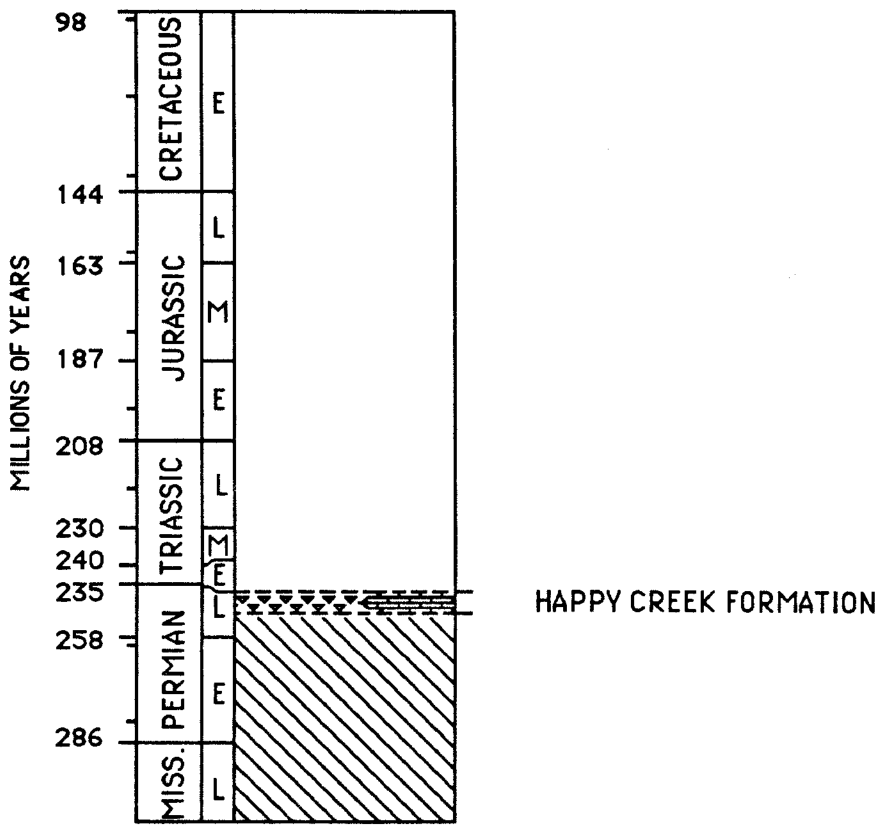
Granite Range This range, just north of the Fox Range across the Black Rock Desert, contains a sequence of volcanic flows and limestone, metamorphosed and intruded by an early Late Cretaceous granodiorite batholith (Bonham, 1969) (fig. 87). The volcanics are vesicular basaltic to andesitic flows, flow breccias and pyroclastics and containing hornblende, biotite and andesine. The limestones are fossiliferous calcarenites, and contain lenses of conglomerate with limestone, quartzite and volcanic clasts. A Triassic ammonite has come from marine sediments in the sequence.

This sequence I correlate to the basal Happy Creek Formation, on the basis of the Triassic age and the stratigraphy (particularly the distinctive combination of interbedded andesitic flows and calcarenitic limestones).

Other Exposures in the Black Rock Desert The Selenite Range contains meta-andesite, meta-dacite, breccias and tuffs, interbedded with chert, limestone and shale (Johnson, 1977). Based on correlation to Black Rock, this sequence has been considered to be Permian in age, and indeed this set of facies does resemble most the upper member of the McGill Canyon Formation (as seen especially in the Pine Forest Range), with which I

FIG.87

TECTONOSTRATIGAPHIC SUMMARY OF THE GRANITE RANGE



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BONHAM (1969)

tenatively correlate them. They are intruded and contact metamorphosed by granodiorite.

The southwest Calico Mountains contain pelitic and cherty facies with turbidites of both siliceous and volcanic provenance (Silberling et al., 1987), and tenatively correlated with the McGill Canyon Formation. In the western Calicos is a small exposure of phyllite, slate and quartzite of Triassic and Jurassic age (?) (Willden, 1964), assigned here to the Bliss Canyon Formation/Nightingale Sequence.

The Nightingale Range is composed of the Nightingale Sequence, which has been placed in the Auld Lang Syne Group by Johnson (1977). The Auld Lang Syne Group and its relationship to the Bliss Canyon Formation will be discussed at a later point.

The eastern edge of the Cretaceous portion of the Sierra Nevada batholith cuts pretty much across the middle of the terrane (Smith et. al., 1971). The plutons, scattered in the east, become increasingly common westward and begin to obscure the earlier relationships. In the northeast corner of the terrane, the more diffuse Jurassic batholith is preserved, as smaller (shallower?) bodies commonly intruding their own volcanic pile (the Happy Creek Formation) and the strata underneath that pile.

Summary of Late Paleozoic and Early Mesozoic Geologic History of the Black Rock Desert

Within this province, which is a composite of the previously defined Black Rock Terrane (Late Paleozoic to Late Triassic basinal) and Jackson Terrane (Late Triassic to Middle Jurassic volcanic) of Silberling et al.

(1987), a highly internally consistent picture has emerged.

The Mississippian to early Early Permian lower member of the McGill Canyon Formation is characterized by basinal hemipelagic and turbidite deposits, from an orogenic source characterized by Devonian age and older marine meta-sediments. Volcanism was not a significant factor. In late Early Permian time in the upper member, andesitic volcanism became important and regional shoaling and offlap took place with the development of local shallow marine carbonate mounds and platforms of the carbonate member. Some of these carbonate successions slid down into the nearby deeper water environments, which apparently lay to the east and south of the shallower marine accumulations. Shallow marine clastic and carbonate platform deposition coninued into the early Late Permian, with volcanism waning as a factor. The formation extends throughout the province.

Rapid subsidence and the deposition of starved basin hemipelagic chert and argillite marks the late Late Permian through the early Middle Triassic, in the Quinn River Formation. Whether this unit is present in the southern part of the Black Rock Desert is not clear. Minor Middle Triassic volcanism far away is marked by thin silicic tuffs at Quinn River (perhaps related to the Koipato Rhyolite and related units, and serving as a tie between these otherwise highly disparate units). There is no evidence in this area for the Sonoma Orogeny, other than very indirectly in the cessation of volcanism, the lack of clastic influx into the basin, and the rapid subsidence.

Basinal deposition was renewed in the late Middle Triassic with the Bliss Canyon Formation and Nightingale Sequence(?). The basin was anoxic but not starved, and offlap and filling of the basin took place from the west. The shelf margin facies include shallow marine clastics and the development of carbonate platform and local mounds, from the late Middle Triassic to middle Norian. Volcanism was again not a factor. The basin lay east of the arc margin, and probably continued into the Dun Glenn and Auld Lang Syne Groups further east (see later discussion).

Quite suddenly in the middle to late Norian, arc volcanism was turned on again with a distinctive augite-phyric andesite sequence throughout the province. Shallow marine to subaerial andesitic volcanism and uplift continued in the Jackson Mountains and Pueblo Range through the Early and Middle Jurassic. Presumed Jurassic diorite intrusives throughout the area also imply ongoing arc volcanism, even where the bedded units are not exposed. Meanwhile, the Pine Forest Range by the Early Jurassic had subsided to basinal depths and the volcanics higher in the section interfinger with and underlie a turbidite succession. Significant local unroofing and development of relief took place just before that, with large proximal debris flows being shed into that basin. The Happy Creek Formation may have been a large volcanic constructional pile built up as an island on the western edge of and within an Early Mesozoic marine basin.

Borderlands-style wrench tectonics (event D1, and probably sinsitral) in the Middle Jurassic is present in the Pine Forest and Jackson Mountains, as well perhaps as the Fox Range. Just after this phase, the marine basin and volcanic pile were imbricated and uplifted during northwest-southeast regional compressional orogeny (event D2), with complex thrusting directed to the east and west. The King Lear Formation (found only in the Jacksons) is the subaerial molasse shed by this orogeny, filling and overlapping the wrench basins formed by D1. Arc volcanic activity continued through this period but died off in the Late Jurassic or earliest Cretaceous, picking up again in the late Early Cretaceous (in the volcanic record of the King Lear Formation) and Late Cretaceous (as the scattered plutons of the Sierra Nevada batholith). A second thrust event (D3) is documented in the Jacksons in the Aptian of the King Lear Formation. The regional extent of these two events (particularly D3) is not clear, as often the Mesozoic compressional structures within the terrane cannot be assigned accurate ages or phases, and reports on detailed mapping through much of the southern part of the province are almost non-existent.

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