# TECTONICS AND MAGMATISM IN THE WESTERN U.S. CORDILLERA - AN OVERVIEW

**(lightly revised in 2012 from a 1993(?) field trip guide for the International Conference on Geochronology and Cosmochemistry, Berkeley, led by E. Miller, S. Wyld, P. Gans, J. Wright, S. Forrest and A. Snoke; includes only the east-central Nevada part of the trip, not the Raft River or Ruby Range part of that trip)**

## INTRODUCTION

The broad mountainous region of western North America is known as the "Cordillera"[1], an orogenic belt that extends from South America (the Andean Cordillera) through Canada (Canadian Cordillera) and into Alaska (Fig. 1). The youthful topography of this impressive mountain belt is closely related to active crustal deformation as indicated by the current distribution of seismicity across the width of the orogenic belt (Figs. 1, 2A,B). The current plate tectonic setting and the dominant style of deformation vary along strike of the orogen. Folding and thrust faulting occur above active subduction zones in the Pacific Northwest of the U. S., in southern Alaska and the Aleutians. Strike-slip or transform motion occurs along the Queen Charlotte (Canada) and San Andreas Fault (California). Extension and rifting occur in the Basin and Range province of the western U.S. and along the length of Mexico's Gulf of California (Fig. 1). Variations in structural style are also apparent across strike of the orogen; for example, crustal shortening and strike-slip faulting in coastal California are concurrent with extensional faulting and basaltic volcanism in the adjacent, and inboard, Basin and Range province. This great diversity in structural style along and across strike of the Cordillera is likely to have characterized the past history of the orogenic belt for many hundreds of millions of years. The western margin of North America first formed by rifting in the latest Precambrian, its history has been shaped by paleo-Pacific-North America plate interactions since then, making it the longest lived orogenic belt known on earth (eg. Burchfiel et al., 1992). This long history of subduction has in turn led to the development of some of the best-known magmatic belts in the world.

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[1]Spanish for mountain or mountain chain

The Cordillera provides an excellent natural laboratory for the study of time-space distributions of magmatic activity, the relationship of magmatism to plate boundary tectonism, and the role and contribution of magmatism to the evolution of continental crust. In detail, however, the exact relationships between plate motions, magmatism and continental deformation still remain elusive. For example, plate tectonics is a simplification that applies well to oceanic lithosphere which is dense and strong and thus capable of transmitting stresses across great distances without undergoing significant internal deformation. This is not true of continents whose more quartzo-feldspathic composition and greater radiogenic heat flow make them inherently weaker. Displacements or strain within continental crust can accumulate at plate tectonic rates (1-15 cm/year) within narrow zones of deformation, or can take place more slowly (mm's to cm's/year) across broad zones of distributed deformation. Continents can accumulate large strains, can thicken over broad distances during crustal shortening and thin across broad regions during extension. These strain histories are usually partially preserved in the geologic record because of the inherent buoyancy of continental material which prevents it from being subducted into the mantle. Magmatism can have a profound influence on deformation of continental crust in that increased temperatures cause rheologic weakening of the crust. In the Cordillera, magmatism has played a key role in determining where and how deformation has occurred, and in part has dictated the magnitude of associated deformation. Specific understanding of the links between plate motions, plate boundary processes, magmatism and continental deformation is evolving slowly as this understanding grows with more detailed geologic and geochronologic studies, providing quantitative information on the time-scale of events, the rates of geologic processes, and our ability to compare timing of events across the Cordillera as a whole—the spatio-temporal evolution of both magmatism and deformation. Geophysical and petrologic/isotopic studies remain important and underutilized tools in our understanding of Cordilleran-wide events and how physical processes in the deeper crust and mantle are coupled to more easily studied magmatism and deformation at shallower levels of the crust.

## GEOLOGIC AND TECTONIC HISTORY

Studies of modern active plate margins have played an important role in interpreting the more fragmentary evidence for such activity in the geologic record of the Cordillera. Based on these comparisons, it appears that all plate tectonic styles and structural regimes known to us, with the exception of continent-continent collision, have had a role in the creation and evolution of the Cordillera.

## Late Precambrian to early Paleozoic

The initial formation of the Cordilleran margin occurred in the late Precambrian. The Windemere Supergroup is a thick succession of shelf-facies clastic rocks deposited between about 730 Ma and 550 Ma whose facies and isopachs define the newly rifted margin of western North America after the break-up of the Laurentia supercontinent (Hoffman, 1991). The Windemere Supergroup forms the lower clastic part of a 15 km thick, dominantly carbonate shelf succession whose isopachs and facies boundaries closely parallel the trend of the Cordillera. This shelf sequence (including the older and more localized basinal deposits of the Belt-Purcell Supergroup) is now spectacularly exposed in the eastern Cordilleran fold and thrust belt whose overall geometry and structure are controlled by the facies and thickness of this succession (Sevier belt and Canadian Rocky Mountains thrust belt, Fig. 3).

## Paleozoic

The Paleozoic history of the Cordillera has generally been described as one of continued passive margin sedimentation and little active tectonism. However, there are a multitude of tectonic fragments of island arc and back-arc basin successions of Cambrian to Triassic age embedded in the western Cordillera (Fig. 3). The common conception that these represent a collage of far-traveled terranes entirely exotic to the Cordillera ("suspect terranes") has been continously re-evaluated. Numerous studies indicate that at least some of these sequences developed adjacent to, but offshore of the western edge of the continental margin (e.g. Miller et al., 1992). Some of the terranes are now known to be more exotic than others in that they have geological and sedimentary ties to Baltica and the Arctic (e.g. Miller et al., 2011). Although these "exotic" fragments are likely displaced from their site of origin by rifting and strike-slip faulting, their presence nonetheless argues convincingly for a long history of subduction of paleo-Pacific crust beneath the western edge of the North American plate. Study of these accreted fragments suggest that during the late Paleozoic, western North America may have looked much like the SW Pacific today with its fringing arcs separated from the main Australasian continental shelves by back-arc basins (e.g. Miller et al., 1992). During the Paleozoic, the continentward portion of the western North American shelf experienced regional subsidence and uplift events but little deformation. Exceptions include deformation of continental margin sediments and intrusion of Latest Proterozoic to Late Devonian granites in Alaska and southern British Columbia, the closure of deep water, back-arc basins, and thrusting of basinal successions onto the shelf during the earliest Mississippian *Antler orogeny* (Roberts

Mountains thrust) and during the Permo-Triassic *Sonoma orogeny* (Golconda thrust) in the western U.S. part of the Cordillera (Fig. 3).

## Triassic and Jurassic

The bes-developed magmatic belts related to eastward subduction beneath western North America are of Mesozoic age. Arc magmatism of Triassic and Early Jurassic age (230-180Ma) is recorded by thick sequences of mafic to intermediate volcanic rock erupted in an island arc (Alaska, Canada and northwestern part of U.S.) to continental arc (southwestern U.S.) setting (Fig. 4). Tectonism in the overriding continental plate during this time-span was generally extensional in nature, leading to rifting and subsidence of parts of the arc and continental margin (Karish et al., 1987; Dilles and Wright, 1988; Saleeby et al., 1992).

The Middle to Late Jurassic brought a dramatic change in the nature of active tectonism along the entire length of the Cordilleran margin. This time-span is characterized by increased plutonism during the interval 180-150 Ma (Fig. 5) accompanied by significant crustal shortening. This shortening closed intra-arc and back-arc basins, accreted arc complexes to the North American continent, and fundamentally changed the paleogeography of the Cordillera (north of latitude 39°) from an Aleutian-type margin to an Andean-type margin. The Andean tectonic framework persisted throughout most of the latest Mesozoic and Cenozoic. In Canada, Middle Jurassic orogenesis (referred to as the Columbian orogeny) has been attributed to the collision of the Intermontane superterrane (composed of the Stikine, Cache Creek, Quesnel and Eastern terranes) (Monger et al., 1982). In the U.S. portion of the Cordillera, there is no evidence for collision, only eastward subduction of oceanic crust beneath the margin during this time span. Deformation in the back arc region at about 39° N in the Middle to Late Jurassic began first in the region of elevated heat flow within and behind the arc then migrated eastward with time towards the continental interior (Smith et al., 1993). In general, deformation appears to be localized adjacent to Jurassic plutonic complexes. Where well-dated, deformation and metamorphism is early to late Middle Jurassic (~181-165 Ma) along the eastern margin of the magmatic arc in western Nevada, Middle to early Late Jurassic (~172-161 Ma) in the back-arc region of central Nevada, and Late Jurassic (~160-144 Ma) in eastern Nevada and western Utah. A nearly identical pattern in the age of subduction-related plutonism across the transect is also seen, making a strong argument that the two are linked, and that heating by magmatism allowed the upper parts of the crust to deform locally during intrusion.

Jurassic orogenesis in the U.S. Cordillera was probably caused by rapid absolute westward motion of the North American plate (Fig. 6) (Engebretson et. al. 1985; Smith et al., 1993), which took place as the North Atlantic began to open. In effect, this westward plate motion caused the western margin of the continent to "collide" with its own arc(s) and subduction zone(s), leading to the accretion of outboard terranes. With continued westward motion, the overriding continent began to deform internally during an event which lasted broadly 30 Ma. During this time-span, deformation migrated eastward into the continent with time, in concert with the eastward migration of subduction-related magmatism.

## Cretaceous to Early Cenozoic.

A better record exists for Pacific plate motions with respect to North America during the Cretaceous to Early Cenozoic (in large part from magnetic anomalies on the ocean floor; Engebretson et al., 1985) and it is possible to draw some inferences about the links between orogenic and magmatic events and the history of subduction beneath the active margin (Figs 5, 6 and 7). There is a general lull in deformation and a lack of evidence for significant magmatism during the time interval ~140-125 Ma, which may correspond to a time of low orthogonal component of convergence along the Cordilleran margin and/or strike slip (Fig. 7) (Page and Engebretson, 1984). Particularly high rates of orthogonal convergence of the Farallon and Kula plates began again in the Late Cretaceous (beginning at about 120 Ma) to early Tertiary, in conjunction with an episode of rapid westward motion of North America (Fig. 6 and 7) (Engebretson et. al., 1985). During this time, major batholithic belts developed in the western North American Cordillera in concert with widespread crustal shortening (Figs 5, 6 and 8). Depending on the configuration of subducting oceanic plates, large components of margin-parallel strike-slip faulting are also implied for the Cretaceous (post 120 Ma) to early Tertiary (Engebretson et al., 1985). This component of motion would have, in general, displaced parts of the margin and its accreted terranes northwards towards Alaska along right-lateral strike-slip faults (Wyld et al., 2006). How much, where, and along what faults displacement occurred are still controversial questions.

In the western U.S., increased rates of subduction resulted in the emplacement of the bulk of the Sierra Nevada batholith (Fig. 8). On the western (forearc) side of the batholith, depressed geotherms caused by subduction led to high pressure-low temperature (blueschist) metamorphism within rocks now represented by the Franciscan Complex in the Coast Ranges, which are interpreted as a subduction zone assemblage. The intervening Great Valley basin (Figs. 2, 3) underwent a similar history of "refrigeration" during this timespan of subduction. Sediments deposited in this basin, buried as deep as 10 km, reached temperatures of only about 100°C, suggesting thermal gradients of 10°C/km or less (e.g. Dumitru et al., 1991). In contrast, heat flow in the arc and back arc region was high and magmatism was accompanied by crustal shortening. Deformation appears to have migrated eastward with time (Smith et al., 1993) and resulted in the well-known Cretaceous to (locally) early Tertiary Sevier foreland fold-and-thrust belt (Figs. 8, 9). The major thrust faults forming the Sevier thrust belt (Figs. 8, 9) displace stratified Paleozoic-Mesozoic shelf sediments eastward, and have a minimum total displacement of 100-200 km at the latitude of Nevada/Utah. Along most of its length, the eastern front of this thrust belt closely follows the transition from thin cratonic sections to thicker shelf sequences, indicating strong stratigraphic control on the overall geometry of the belt. In Southern California and Arizona, the thrust belt follows the eastern edge of the magmatic belt, but cuts across the craton/shelf transition and involves crystalline basement, arguing that the main control on the location of shortening-related structures in this area is thermal weakening of the crust by magmatism (Burchfiel et al., 1992). Based on geologic and geochronologic studies within tilted crustal sections west of the thrust belt in Nevada and Utah, we know that deeper parts of the crust lying between the thrust belt and the magmatic arc were hot and mobile and underwent thickening by folding and ductile flow (Fig. 9) (Miller and Gans, 1989).

Cretaceous magmatism in the backarc region exhibits a broad range in terms of age and compositional diversity. Ages range from 120-70 Ma and generally young towards the east. Compositions range from hornblende-bearing granodiorite to muscovite-bearing granites, and, in general, older plutons have less of a crustal signature, whereas younger ones a greater crustal signature (Barton, 1990). Jurassic and Early Cretaceous plutons (ca 170-100 Ma) emplaced into the back arc region prior to inception of crustal thickening consisted primarily of a compositionally expanded (gabbro to biotite granite) metaluminous suite. The Sr and Nd isotopic systematics of this plutonic suite indicate the presence of a significant mantle component (Fig. 10a) (Wright and Wooden, 1991). Following crustal thickening, the character of plutonism changed dramatically and a more compositionally restricted suite of peraluminous two-mica ± garnet granites was emplaced in the region just to the west of the main thrust belt (Fig. 9), over the time interval of approximately 83-75 Ma. Sr and Nd isotopic systematics of this plutonic suite are dramatically distinct from the Jurassic-Early Cretaceous suite and are consistent with the interpretation that they were derived almost entirely from partial melting of crustal rocks and metasedimentary strata (Figure 10) (Wright and Wooden, 1991). The isotopic data do not support a model of increasing crustal contributions to magmatism through time but appear instead to indicate two isotopically distinct groups of plutons. This is consistent with a model in which the younger suite of plutons represent crustal melts caused by crustal thickening with little or no mixing of mantle-derived magmas with these melts. There is a similar, but slightly older, suite of granites developed further to the west that appear to be spatially associated with the lesser-displacement Eureka thrust belt (Smith et al., 1993) (Fig 8).

Because the Cretaceous orogen described above was reworked by Cenozoic faulting and extension across the Great Basin region, the amount of crustal thickening that occurred during the Mesozoic is not certain. Was the western U.S. like the Tibetan Plateau at the end of the Cretaceous, underlain by 70-80 km thick crust? Or was crustal thickening more modest as evidenced by the ~50 km thick crustal root beneath the Canadian Cordillera today? This is a still controversial and debated question. If the crust was thick and hot, thus rheologically weak, extensional structures of Cretaceous and Early Tertiary age should have developed (as in the high Himalaya, e.g. Burchfiel et al., 1992). There are some structures of this nature and age reported in the literature ( e.g. Hodges and Walker, 1992; Wells et al., 2012 ) but they are few and their significance and exact age are not certain. None are shown in regional compilations (eg. Burchfiel et. al., 1992) (Fig 11).

Magmatism in the western U.S. portion of the Cretaceous magmatic belt ended abruptly at about 80-85 Ma, although subduction continued and in fact accelerated, achieving convergence rates of ~15cm/yr ( Engebretson et. al. 1985 ) (Figs. 6,7). To the north of the Snake River Plain and to the south of Las Vegas, subduction-related magmatism continued uninterrupted into the Paleocene and was accompanied by crustal shortening and thrust faulting (Fig. 11). These north to south differences have been attributed to segmentation of the subducting slab, with extremely shallow angle subduction beneath only the western U.S. (Great Basin) portion of the belt. This hypothesis is supported by evidence for rapid cooling of the Sierra Nevada batholith as it moved into a "forearc" position, thermally speaking. As the crust of the arc and backarc was "refrigerated" it regained its rheologic strength and ability to transmit stresses for greater distances (Dumitru et al., 1991). During this time, deformation stepped far inboard to Utah, Colorado and Wyoming, where crustal-penetrating reverse faults caused uplift of the Rocky Mountains during the latest Cretaceous to Eocene *Laramide orogeny* (Fig. 3, Fig. 11). Few or no active structures are documented in the Great Basin region, supporting the suggestion that, even though it may have been underlain by thickened crust, it may have been cool enough to have behaved mostly rigidly during this time span or that its behavior was controlled by the shallowly dipping slab (Fig. 11). The uplift of structures in the Rocky Mountains was contemporaneous with continued shortening in the foreland thrust belt in Arizona and Mexico to the south and in Montana and British Columbia to the north (Fig. 11).

## Cenozoic

Plate motions between the oceanic Kula and Farallon plates and North America changed again at the end of the Paleocene and the component of orthogonal convergence diminished rapidly (Figs. 6, 7) ( Engebretson et. al., 1985). In the western U.S., it is hypothesized that the shallowly-dipping slab either fell away into the mantle or gradually "decomposed" due to conductive heating, allowing hotter and deeper mantle to rise/upwell beneath the continent (Severinghouse and Atwater, 1990). Decompression melting of upwelling asthenospheric mantle into this region generated basalts which heated the base of the thickened continental crust, causing extensive assimilation and melting of crustal rocks. This magma mixing ultimately led to eruption of large volumes of intermediate to silicic volcanic rocks (Fig. 4). The hybridization of crustal and mantle sources during this time span is particularly obvious when isotopic signatures of these rocks are compared to that of earlier magmatism (Fig. 10a and b); they display the full range of compositions and isotopic values. Volcanism migrated progressively into the area of previous flat slab subduction, both southeastward from the Pacific NW and northward from Mexico (Fig. 12). The large input of heat into the thick crust caused rheological weakening, triggering an early phase of localized extensional thinning and flow of the deep crust. Rapid large-magnitude extension and diapiric rise of deep rocks occurred locally in the core complexes. When migrating volcanism merged at the latitude of Las Vegas at about 17-20 Ma (Fig. 12), the process of slab decomposition was presumably complete and the Basin and Range began to extend as a whole (Fig. 2), acquiring its present-day configuration of mountains and valleys. This broad zone of continental extension (Fig. 2) wraps around the southern end of the unextended but (thermally) elevated Colorado Plateau and projects as a finger northwards along the Rio Grande Rift on the eastern side of the Plateau. In the western U.S. it appears to have developed within and is restricted to the region of earlier inferred shallow slab subduction (e.g. compare Figs. 2 and Fig. 11) suggesting a genetic relationship. The western boundary of the Basin and Range is represented by the unextended Sierra Nevada crustal block with its thicker crustal root. To the west lies the virtually undeformed Jurassic-Cretaceous Great Valley sequence, underlain in part by oceanic crust refrigerated during Mesozoic subduction (Fig. 2).

Basaltic and bimodal volcanism characterizes the young history of the Basin and Range province. Volcanism and seismicity are diffuse across this broad zone of continental extension and thermal springs abound. One of the most impressive physiographic features related to young volcanism is the Snake River plain depression, believed to represent the Miocene to Recent track of a mantle hotspot which now resides beneath Yellowstone (Fig. 2B). The present Basin and Range province, together with associated extension in the Rio Grande Rift and that occurring north of the Snake River Plain reflect a total of 1.2X to 1.5X of E-W extension that began in the early-mid Tertiary and continues today. The modern, regularly spaced basin-and-range physiogeography that lends the province its name is simply the surficial manifestation of the dominant and youngest system of major normal faults that began their motion in the Miocene.

Given the long history of plate interaction along the western margin of North America, it may seem surprising that the actual limits and present topography of the western U.S. Cordillera are largely dictated by the youngest events to affect the belt. For example, the Basin and Range province includes all or parts of the Mesozoic magmatic arc, back arc thrust belt, older Paleozoic allochthons and sutures, and is also underlain by the Precambrian rifted margin of western North America. Despite the diversity of tectonic elements displayed across the Basin and Range, the crust is uniformly 25-32 km thick and much of it stands >1km above sea level, reflecting an anomalously thin and hot mantle lithosphere. The Moho across this broad (600 km) extensional province is exceptionally flat, despite differences in the history of older shortening and younger extension (McCarthy and Thompson, 1988), and implies that the lower crust was capable of undergoing large-scale flow during extensional deformation (Gans, 1987). Thus, it seems clear that the present-day structure of most of the crust and perhaps the entire lithosphere across this region reflects only the youngest events to affect this long-lived orogenic belt. This would imply that if the upper 5-10 km of the crust were removed by erosion, we would probably see very little evidence for the previous 600 m.y. history of this orogenic belt. If these upper levels of the crust were removed, we speculate that Tertiary granites and Tertiary high grade metamorphic complexes (sillimanite gneiss domes and granulite facies rocks) would constitute a greater portion of the exposed geology.

In California, the current relative motion between the Pacific and North American plates is partitioned into strike-slip displacement along the San Andreas fault and into folding and thrusting related to shortening perpendicular to the San Andreas transform plate boundary (as reflected by the recent Loma Prieta and Northridge earthquakes). Part of that plate margin strike-slip displacement (about 1.2 cm/year) is currently taking place along the eastern side of the Sierra, within the Basin and Range province itself (e.g. Faulds and Henry, 2008).

## CLOSING COMMENTS

This brief discussion of the geologic and tectonic evolution of the Cordillera permit us to make several generalizations about the evolution of such orogenic belts.

1. Mountain building and orogeny (that is, thickening of continental crust) are not necessarily the result of subduction and collision of allochthonous crustal fragments (terranes) along an active continental margin. Subduction occurred for long spans of time during the history of the Cordilleran margin, but, like the SW Pacific, led mostly to rifting and back-arc basin development. True mountain building in the Cordillera appears to have occurred during finite time intervals of rapid relative convergence, magmatism, and rapid absolute westward motion of North America with respect to its western subducting plate boundary.

2. The Cordillera has long been cited as a classic example of continental growth by the lateral accretion of allochthonous terranes (Coney et al., 1982). However, this mechanism is probably *not* the most fundamental or important process of crustal growth, unless it involves the addition of mature island arcs. Rifting, with formation of rift basins on existing continental shelves, along continental slopes, and within mature island arcs, coupled with the subsequent filling of these basins by thick prisms of sediment have together contributed significantly to the formation of many terranes now incorporated in the Cordillera. Extensional thinning and reworking of continental crust or previously thickened orogenic crust, when accompanied by magmatic additions from the mantle, can serve to redistribute and remobilize crust across great portions of an orogen, the amount of extension often equaling or exceeding estimates of crustal shortening within the same belts. The intrusion of mafic magmas into lower crust and removal of more easily melted components by hybridisation and melting during extension serve to stabilize continental crust after episodes of crustal thickening. The best example of this process is the reworking of the entire western continent during Cenozoic extension. These considerations suggest we should not use area-type calculations to determine growth rates for continental crust, as estimates of lateral crustal growth must take into consideration processes that occur in the third dimension as well.

3. Magmatism is sometimes viewed as a process that is largely independent of deformation in mountain belts. The Cordillera provides excellent examples that magmatism is, instead, intricately tied to deformation of continental crust, in that heating causes rheologic weakening, and permits deformation to occur. The rise of magmas help advect heat to higher levels of the crust, permitting continents to undergo large-scale deformation, whether related to shortening and thickening or stretching and thinning. This is manifested by the increasingly better documented eastward migration of magmatism and deformation in the Mesozoic as well as the space-time relation between magmatism and extensional tectonism in the Cenozoic of the western U.S. Cordillera.

**ICOG 8 FIELD TRIP # 10**

# GRANITES AND CORE COMPLEXES OF THE GREAT BASIN

# (NOT INCLUDING RAFT AND RUBY)

**FIELD TRIP LEADERS:**

**Elizabeth L. Miller, Phillip B. Gans, Jim Wright and Sandra Wyld**

**Roadlog revised 4/2012 by E.L. Miller**

## DAY 1 -

Leaving Stanford, we first cross the eastern side of the Coast Ranges underlain by the Franciscan Complex and enter into the Great Valley, which, together with the Sierra Nevada batholith, constitute California's three most famous lithotectonic assemblages, the "Holy Trinity" generated during Mesozoic Andean subduction. Stops are possible stops in the Sierra Nevada in order to look at granitic rocks of the batholith above the shores of Donner Lake. Most of the day will be spent driving across the Basin and Range provincee

## CALIFORNIA'S HOLY TRINITY

### Franciscan Complex

The Franciscan complex is an accretionary prism formed during Late Jurassic to late Cenozoic subduction of Pacific basin oceanic plates beneath the western margin of the North American plate (e.g., Ernst, 1970). It consists mainly of highly deformed sediments and lesser mafic rocks subjected to low-grade, high-pressure–low-temperature (≈100°-350°C) metamorphism and has been divided into three belts that are generally younger and less strongly metamorphosed towards the west. The Eastern belt consists mainly of sediments and basaltic volcanics that were subducted to depths of 10 to 30 km and metamorphosed under high-P/T conditions (lawsonite and aragonite are widespread, with jadeitic pyroxene in a few areas). The Central belt is mainly a chaotic mud-matrix melange containing blocks of sandstone, volcanic rock, chert, and as well as most of the rare, exotic higher-grade blueschist, amphibolite, and eclogite blocks for which the Franciscan is famous. The Coastal belt is mainly Paleocene and Eocene in depositional age and only incipiently metamorphosed (e.g., Cloos, 1986; Blake et al., 1988). The Franciscan is notable among blueschist belts in that it never experienced any greenschist facies overprint.

Most of the high-grade blocks in the Franciscan were apparently mafic rocks metamorphosed to amphibolite facies before they were overprinted by eclogite and/or blueschist facies metamorphism and then incorporated into much lower grade melange matrix (e.g., Moore and Blake, 1989; Wakabayashi, 1990). $^{40}Ar/^{39}Ar$ plateau ages from hornblendes in seven amphibolite blocks are all within the range 162±3 (Ross and Sharp, 1988) and probably date amphibolite metamorphism. A U-Pb isochron age on a blueschist block is also 162±3 Ma (Mattinson, 1986). Blueschist white mica K-Ar and limited $^{40}Ar/^{39}Ar$ ages are younger, mostly 138-159 Ma (summarized in McDowell et al., 1984). These data may reflect cooling through the lower closure temperature of white mica. Alternately, some have speculated that amphibolite metamorphism occurred on the sole of a hot ophiolite that was obducted onto the plate margin and that this amphibolite-greenschist terrane was later transected by the Franciscan subduction zone and overprinted to eclogite and/or blueschist facies (Blake et al., 1988).

Extremely fine grained Pickett Peak terrane metasediments of the Eastern belt have yielded whole rock K-Ar and whole rock total fusion $^{40}Ar/^{39}Ar$ ages mostly from 110-143 Ma, with a tendency to cluster around 120-125 Ma (Lanphere et al., 1978; McDowell et al., 1984, and references therein). Some workers interpret the age of metamorphism as ≈115-125 Ma (e.g., Lanphere et al., 1978; Blake et al., 1988) and suggest accretion and metamorphism of the Pickett Peak was a distinct event younger than the metamorphism of the high grade blocks. Others suggest that fine-grained micas are quite susceptible to Ar loss and the ages set a minimum for time of metamorphism. These workers generally infer that metamorphism of the high-grade blocks and Pickett Peak terrane reflect the same metamorphic event during initial cooling of the subduction zone at the initiation of Franciscan subduction (e.g., Cloos, 1985; Wakabayashi, 1990).

Apatite fission track data record a complex and protracted unroofing history, with middle to Late Cretaceous, early to middle Tertiary, late Tertiary, and Quaternary episodes of exhumation in various areas (Dumitru, 1989).

## Great Valley Sequence

Late Jurassic to Recent sediments of the Great Valley Group were deposited in an elongate north-northwest trending basin flanking the Sierra Nevada arc. Representative sections are best exposed along the western side of the Great Valley where they are folded and tilted eastward along the underlying Coast Range "thrust". Within these sections, terrigenous sandstone, greywacke, mudstone and conglomerate of the Great Valley Group are locally demonstrated to depositionally overlie basalts and hemi pelagic sediments of the Middle to Late Jurassic Coast Range ophiolite. In the northern Coast Ranges an impressively thick (~15 km) Upper Jurassic to Upper Cretaceous section is exposed and semi continuous subsidence of the basin is inferred from these sections. Along the eastern or Sierra Nevada side of the Great Valley, these stratigraphic units thin to zero and unconformably overlie a diverse igneous and metamorphic basement terrane, part of the Sierra Nevada batholith and its country rocks. Given the uplift history of the

Franciscan Complex (in order to exhume its blueschist facies rocks) adjacent to the semi continuous history of subsidence of the Great Valley Basin, it is likely that the boundary between the two has had a more complex history of movement than that of just a simple thrust. Recent studies have suggested that the original thrust contact may have been modified by normal-offset faults (e.g. Harms et al., 1992), which is also indicated by map relations along parts of the contact where little metamorphosed Great Valley sediments are in fault contact with rocks having undergone a coeval high-pressure metamorphic history.

The Great Valley sequence was mostly sourced from the Sierra Nevada arc. The proportion of lithic detritus in the Great Valley Sequence decreases upwards through the section, while the proportions of feldspar and quartz increase due to progressive unroofing of the plutonic roots of the Sierra Nevada magmatic arc with time (Ingersoll et al., 1977).

The Great Valley Sequence is inferred to have accumulated in a fore-arc or outer-arc basin during development of the Sierra Nevada volcanic arc and intrusion of the Sierra Nevada batholith at depth. Sedimentation in the fore-arc basin continued into recent times, outlasting magmatism, which shut off abruptly at about 80 Ma in the Sierra Nevada.

## The Sierra Nevada Batholith

The Sierra Nevada batholith consists mostly of Cretaceous and lesser Jurassic batholithic rocks which are exposed for approximately 640 km along the trend of the Sierra Nevada. The Sierra Nevada forms part of an even greater continental margin subduction related magmatic belt extending from southern Alaska to Baja California (Figs. 3 and 3A).

The Sierra Nevada batholith intrudes and cuts a variety of older, variably deformed, northwest striking lithostratigraphic units or terranes. In southern California these consist of Late Precambrian and Paleozoic strata of the Cordilleran miogeosyncline (passive margin shelf strata), and the older Precambrian crystalline basement rocks upon which this section was deposited. At the latitude we will cross the batholith along Highway 80, it intrudes Paleozoic to early Mesozoic volcanic arc and associated sedimentary sequences deposited in a marine setting. Basically these relations imply that the Mesozoic arc may have looked similar to the Alaskan-Aleutian arc today, its southern part developed in a continental setting and its northern part in an offshore, oceanic arc setting.

The Sierra Nevada is a composite batholith consisting of many individual discrete plutons, comagmatic plutonic suites and dikes, largely ranging in composition from quartz-diorite to granite. Smaller masses of mafic plutonic rocks include diorite, quartz diorite and hornblende gabbro. Plutonic rocks of the Sierra Nevada batholith extend far to the east and south of the Sierra Nevada itself, into both the northern and southern Basin and Range province. The Sierra Nevada is simply a young mountain range formed during Basin and Range extension.

Initial $^{87}Sr/^{86}Sr$ values (Kistler et al., 1971; Kistler and Peterman, 1973), $K_2O$ content (Bateman and Dodge, 1970), composition of plutons (Moore, 1959), rare-earth patterns (Dodge and Mays, 1972), and Pb isotopes (Doe and Delevaux, 1973) suggest magmatic source regions that vary along and across strike of the batholith. An increasing number of U-Pb ages provide a good basis for specifying intrusive events or a temporal context for geochemical/isotopic variations within the batholith (Fig. 3B).

Triassic plutons (225-195 Ma) occur mainly on the eastern side of the Sierra Nevada. Jurassic plutons (195-140 Ma) are more voluminous and occur mostly on the eastern flank of the range, but also in the western foothills of the Sierra Nevada. The bulk of the batholith consists of Cretaceous plutons that intrude into the central part of the batholith. Cretaceous intrusions young eastward within the main batholith, and are on the average less mafic eastwards and with age (e.g. Chen and Moore, 1982). Cretaceous plutons range from about 140-80 Ma but few plutons were intruded in the time span 149 - 120 Ma which represented a general lull in magmatic activity. At the beginning of this lull, a regional dike swarm, the Independence dike swarm, was intruded parallel and along almost the entire length of the batholith from Arizona to northern California, dated as 148 Ma (Chen and Moore, 1979). The youngest part of the batholith consists of 94-83 Ma zoned and nested plutonic suites of the Tuolumne Intrusive Suite (Fig. 3B)

Magmatism in the Sierra Nevada ended abruptly between 85 and 80 Ma and its cessation is believed to be a consequence of the inception of shallow slab subduction beneath this portion of the margin, leading to relative "refrigeration" of arc and back-arc crust and migration of deformation far to the east to the Rocky Mountains of the Colorado Plateau (Dumitru et al., 1991).

Widespread magmatism began again in the Tertiary, when subduction slowed and the slab steepened and foundered into the mantle. Oligocene to Early Miocene rhyolitic ash-flow-tuffs were erupted from centres along and to the east of the present Sierran divide. This was then followed by a minor hiatus in magmatism lasting about 5-10 m.y. before widespread Middle to Late Miocene andesitic volcanism began. Scattered Pliocene and Pleistocene basalt and rhyolite mark a transition to bimodal volcanism at this time.

## ROAD LOG - BERKELEY TO DONNER SUMMIT

**0 miles:** You will drive north from Berkeley on Hwy 80 to Richmond (Fig. 15). You will cross the Hayward Fault, a major strand of the San Andreas Fault System, which bounds the coast ranges ont their west side, and then the Coast Ranges themselves which are here mostly underlain by folded Tertiary marine and non-marine sandstones.
**47 miles:** Vacaville exit. Here we head east across the flat and featureless Great Valley, underlain by Jurassic, Cretaceous and Tertiary fore-arc basin deposits.
**76 miles:** Highway 880, bypass Sacramento to the north.
**90 miles:** Drive through Quaternary gravels and braided stream deposits that are well exposed in roadcuts on the north side of the highway.

From Citrus Heights exits to Auburn exits we cross the Penryn dioritic pluton (Jurassic) in the Sierran foothills belt. At Auburn we pass into mostly Jurassic metasedimentary wall rocks of the Sierra Nevada. These are arc and intra-arc basin volcanic and sedimentary rocks as well as ultramafic rocks. Their relationship to the continental margin (exotic, allochthonous or autochthonous) is debated.

On the climb up to Donner Summit, between the Monte Vista and Dutch Flat exits we cross the Melones Fault zone which separates lower and upper Paleozoic strata (also wall rocks of the batholith from Mesozoic foothill belt above). To the north, the long ridgelines in the distance are the gently west-dipping peneplain surface of the Sierra Nevada, that has been tilted about 3° or less by Basin and Range faulting. Miocene and Pliocene volcanic rocks were deposited unconformably on granite, but are mostly preserved in, and outline, paleo valleys that existed at the time of their eruption. Source areas lay to the east of the present crest of the mountain.
**164 miles:** Here we cross into the main granites of the batholith. Near Lake Spaulding turnoff, metasedimentary wall and roof rocks of the batholith form part of a Triassic-Jurassic arc sequence that rests depositionally on yet older upper Paleozoic arc sequences.

Approaching Donner Summit, after Kingvale exit, take the Soda Springs exit and follow signs for Donner Pass Road east, the old highway, south of present Hwy 80.

Past Norden and Lake Van Norden, take the turnoff to Donner Lake overlook or observation point, our first stop.

**Stop 1.1** Donner Summit

Here we find good exposures of the Cretaceous Donner Summit pluton which has yielded a K-Ar biotite age of 97 Ma (Everden and Kistler, 1970). More detailed geochronology has not been done in this pluton. This is a favorite stop to look at relations bearing on the origin of mafic enclaves that are so common in many of the plutons of the Sierra Nevada. The outcrop contains large masses of enclave-rich granodiorite intruded (?) or surrounded by a more enclave-poor granodiorite. Enclaves are diorite porphyry with phenocrysts of hornblende and plagioclase. Many different origins have been proposed for these mafic inclusions. These include (1) the possiblity that the enclaves represent fragments of mafic dikes, disseminated and incorporated into the granodiorite before it had completely solidified, (2) possibly pieces of slightly more mafic, previously crystallized magma entrained and carried upward within the ascending magma (e.g. Dodge and Kistler, 1990), and (3) that they represent blobs of co-existing but immiscible magma. At this stop, all rock types are cut by more felsic, aplitic dikes.

From here we continue driving east along Donner Summit Rd. We can stop at fresh, blasted exposures less than a mile down the road to grab some good samples if you'd like. Just a mile or so from stop 1.1, our lunch site is at the western and of Donner Lake. Possible swimming site for members of the Polar Bear Club.

At the eastern end of the lake is Donner Memorial State Park, where we will stop for historical details regarding the unfortunate Donner party who tried to cross Donner Pass in a bad winter.

Take freeway on ramp to eastbound Hwy 80. On the north side of the highway, before we reach Truckee, one can see well-stratified mudflow/lahar deposits in high cliffs and roadcuts. They are part of the Miocene Kate Peak Formation, deposited upon an old erosional surface of Paleogene age. Note the very gentle west dip of units.

## EASTERN ESCARPMENT OF THE SIERRA NEVADA

As we descend from Donner Summit to Reno, Nevada, we cross from the Sierra Nevada block to the Basin and Range physiographic and structural province (Fig. 2A). The western boundary of the Basin and Range province consists of one or more major east-dipping normal faults that create an impressive topographic escarpment along the eastern front of the Sierra Nevada. This topographic escarpment is in part a function of this faulting and the westward tilting of the Sierra Nevada block, and in part related to the fact that the thickness of the crust changes from ~40km beneath the Sierra Nevada to 28-30 km beneath the Basin and Range province.

Movement on the normal faults in that part bound the Sierra Nevada began ~3 Ma and now include right-slip, normal and oblique-slip faults (Zoback, 1989). A number of faults are tectonically active as evidenced by: (a) recent seismicity - an active seismic zone follows the boundary between the Sierra Nevada and Basin and Range, along which several magnitude 5.6 or greater events have occurred this century, in particular the 1872 Owens Valley earthquake of magnitude 7.5 to 7.8 (Rogers et. al., 1991); (b) youthful fault scarp and geomorphic features; (c) Quaternary volcanism of basaltic and bimodal nature, including the famous Long Valley volcanic centre - source to the voluminous Bishop Tuff (Field Trip #11), and the Owens-Mono Domes. Compilation of GPS strain measurements indicate that 1.2 cm of the 4.8 cm of right lateral motion between the Pacific and North American plates is taking place along the eastern side of the Sierra Nevada, along strike-slip faults of the East Sierra Shear Zone (Fig. 2B).

## THE BASIN AND RANGE PROVINCE

As we drive eastwards from Reno we have now entered the Great Basin region of the Basin and Range physiographic province. The Great Basin is a broad, elevated area characterized by internal drainage and which includes most of Nevada, the western part of Utah, and parts of southeastern California, southernmost Oregon and Idaho (Figs. 2A,B). The Great Basin is more or less coextensive with the northern half of the Basin and Range province.

The Basin and Range province of the Western U.S. Cordillera is a classic example of a broad, active, continental rift, with exceptional exposures of extensional structures. The province is characterized at the surface by northerly trending normal fault blocks related to ~ E-W extension (Fig. 2A,B), and at depth by geophysical observations of subhorizontal reflectivity and anisotropy of the middle to lower crust and uniform (28-32 km) Moho depth (e.g., Klemperer et al., 1986; Hauser et al., 1987; Catchings and Mooney, 1991; Catchings, 1992; Gashawbeza et al., 2008). The Basin and Range, including regions of extension which continue north into Canada, is linked to one of the largest and most prominent magmatic provinces of the Cordillera that evolved over a time span of ca. 55 Ma (e.g. Armstrong and Ward, 1991; Christiansen et al., 1992; Christiansen et al., 2002) (Fig.12). Prior to inception of this magmatism, the Cordilleran margin transitioned from high ocean-continent convergence rates to slower rates in the Cenozoic at ~ 50 Ma (e.g. Page and Engebretson,1984; Engebretson et al., 1985) and from ~80 - ~50 Ma, the future site of the Basin and Range is thought to have been underlain by a shallowly-dipping segment of the subducting Farallon slab (e.g. Atwater, 1989; Humphreys, 1995; 2009; Humphreys et al., 2003). Eocene-Oligocene magmatism in the northern Basin and Range (Fig. 12) is represented mostly by (ultimately) mantle derived calc-alkaline volcanic and plutonic rocks (e.g. Armstrong and Ward, 1991; Christiansen et al., 1992; Best and Christiansen, 1991; Christiansen and McCurry, 2008, Gans et al., 1989; Feeley and Grunder, 1991; Grunder, 1993) and constitutes part of the "ignimbrite flare-up". The initiation of magmatism began at ~ 55 Ma in southernmost Canada and Washington and migrated south, reaching southern Nevada at about ~ 21 Ma (Armstrong and Ward, 1991; Christiansen et al., 1992) (Fig. 1A). Stable isotope paleoaltimetry studies (e.g. Mulch and Chamberlain, 2007) indicate that topography likely migrated with magmatism (e.g. Corson-Kent et al., 2006; Mix et al., 2011). This unusual southerly "sweep" of magmatism (and topography) is believed to be related to the delamination or foundering of the shallowly dipping portion of the subducted Farallon slab, replaced by upwelling mantle asthenosphere (e.g. Atwater, 1989; Armstrong and Ward, 1991; Christiansen et al., 1992; Humphreys, 1995, 2009; Humphreys et al., 2003). In the northern Basin and Range, this magmatism appears to have erupted across, and certainly left behind, a rather featureless volcanic-covered plateau with a continental divide near the central part of the province as documented by a widespread unconformity developed at the base of volcanic strata, preserved in almost every tilted fault block or mountain range of the province (Van Buer et al., 2009; Colgan and Henry, 2009) and wide E-W trending paleo-valleys that channeled ash-flow tuffs, allowing them to flow over 100 km from their source (Henry, 2008; Henry et al. 2011) or spread out across the landscape (Best et al., 2009). Reconstruction of the rock units beneath the unconformity coupled with detrital zircon (DZ) data from Late Cretaceous and Tertiary sediments support the idea of a plateau with little of the structural relief of today's Basin and Range (Van Buer et al., 2009; Colgan and Henry, 2009; Druchke et al., 2011). The absolute elevation of this plateau is more controversial (e.g. DeCelles et al., 2004; Camillery et al., 1997; Mulch et al., 2006; Wells and Hoisch (2008), Druchke et al. (2011) (thick crust, high elevations) vs. Gans et al., 1987; Lerch et al., 2007; Colgan and Henry, 2009; Mix et al., 2011 (thinner crust, moderate elevations)). Although it has been widely reported that the onset of extension and formation of core complexes was contemporaneous with the ignimbrite sweep in the northern Basin and Range (e.g. Christiansen et al., 1992; Miller et al., 1983; Gans et al., 1989; Axen et al.,1993; Axen, 2008; Wernicke, 1992; Dickinson, 2009; Druchke et al., 2011), it is evident, based on the above data, that the Tertiary magmatic and deformational histories recorded in deep crustal rocks within core complexes may not have had much expression at the surface of the earth in terms of developing topographic relief and forming basin-bounding faults. These relations pose significant questions about the history of Tertiary magmatism, metamorphism and deformation documented within core complexes: What was the overall nature of these deep crustal events and processes, how much extension do they represent, and how exactly was this extension accomplished without effects on the surface of the earth? (e.g. Colgan and Henry, 2009; Henry et al., 2011).

The onset of faulting that formed the northern Basin and Range (as we know it today) has been steadily better-documented based on cross-cutting relations, the age and stratigraphy of syn-extensional basin fill deposits, and by low-temperature thermochronology where footwall cooling histories are used as a proxy for the age of fault slip. The inception of faulting in the central part of the province occurred between ca. 20 and 15 Ma, with rapid slip on faults at about 17-16 Ma (e.g. Colgan and Henry, 2009; Colgan et al., 2010; Miller et al., 1999; Stockli, 1999 and 2005; Fosdick and Colgan, 2008); its inception is younger (at ~ 15-12 Ma) in the northwestern (Colgan et al., 2006) and northeastern (Egger et al., 2003; Wells et al., 2000) parts of the province. The youngest faults along the boundaries of the province contribute to the broadening of the province today (e.g. Stockli, 2005; Surpless et al., 2002) (Fig. 1A). The onset of rapid Miocene fault slip in the central part of the Basin and Range was coeval with large volume eruption of mafic magma represented by the Columbia River and Steens basalts at ~17-16 Ma (e.g. Christiansen et al., 2002 and references therein) (Fig. 1A). Associated silicic caldera-forming eruptions along the north side of the province progress both northeastward towards Yellowstone along the Snake River Plain trend and towards the NW into Oregon from the McDermitt Caldera beginning at ~16.5 Ma (Fig. 1A) (e.g. Christiansen et al., 2002). As discussed by Colgan and Henry (2009), the final removal of the inferred Laramide shallow slab (Armstrong and Ward's 1991, "burning of the lithospheric bridge"), the inception of Snake River Plain magmatism, and the westward motion of the plate boundary with development of the San Andreas transform fault, are approximately coeval and potential explanations for the onset of Basin and Range faulting.

When the magmatic and extensional deformation histories of core complexes in the Cordillera are compared, some important variations along the strike of the orogen become apparent (Fig. 1B). North of the Snake River Plain, where crustal shortening (and anatexis) continued to ~ 60 Ma, core complexes preserve evidence for continued partial melting and rapid, diapiric rise of deeper crustal rocks during the onset of magmatism in the early Eocene (Vanderhaeghe et al., 1999; 2003; Foster and Raza, 2002; Whitney et al., 2004; Kruckenberg et al., 2008), and by 30 Ma ago, these once deep-seated rocks were exhumed through apatite fission track annealing temperatures (~ 100°C) (e.g., Lorenack et al., 2001; Foster and Raza, 2002; Vanderhaeghe et al., 2003) (Fig. 1B). Subduction was re-established close to its current position in the Cascade region of the Pacific NW by ~ 40 Ma (e.g. du Bray et al., 2006; Egger and Miller, 2011; Colgan et al., 2010) (Figs. 1A and B). In contrast, core complexes south of the Snake River Plain are characterized by a more prolonged and/or episodic development. The documented time lag between the cessation of crustal shortening in the Sevier belt (Late Cretaceous, e.g. DeCelles and Coogan, 2006) and the age of earliest reported Tertiary magmatism and extension increases southward in the core complexes (Fig. 1B). There is also a time lag between reported metamorphism and deformation in the metamorphic core of the complexes as compared to their age of exhumation based on apatite fission track data and onset of sedimentation in adjacent basins, ranging from at least ~ 7 to as much as ~17 Ma (e.g., Miller et al., 1999; Wells et al., 2000; Howard, 2003; Egger et al., 2003; Wagner and Johnson, 2006; Colgan and Henry, 2009; Colgan et al., 2010) (Fig. 1B). The time lag between the end of shortening in the Sevier Belt and beginning of extensional deformation in core complexes (Fig. 1B) makes it difficult to relate deep-seated Tertiary deformation in core complexes to the earlier history of crustal shortening because existing age brackets indicate that it is closer in age to the onset of Cenozoic magmatism (e.g. Gans et al., 1989; Strickland, 2010; Strickland et al., 2011; in press) (Fig. 1B). The suggestion that Tertiary deformation in core complexes is not related to earlier crustal shortening is countered by a significant number of papers that argue that large magnitude extension was continuous or episodic in core complexes *during* crustal shortening and *prior to* the onset of Tertiary magmatism. These claims are based on a variety of considerations including ages and metamorphic pressures reported from Cretaceous to

Eocene mineral assemblages and $^{40}Ar/^{39}Ar$ ages interpreted as exhumation ages (e.g. Cooper et al., 2010; Hodges and Walker, 1992; Lewis et al., 1999; Wells, 1997; Wells et al., 1990; Wells and Hoisch, 2008) but are countered by recent work in the Funeral Mountains (Mattinson et al., 2007) and Albion-Raft River Range (Strickland 2010; Strickland et al., 2011; in press). Continuing controversy thus exists regarding the Mesozoic vs. Cenozoic age of deformational fabrics in core complexes.

**ROADLOG - RENO TO AUSTIN SUMMIT, TOIYABE RANGE** Continue driving east thru Reno, and stay on Hwy 80, which follows the Truckee River through Cenozoic volcanic rocks (Fig. 16).

**252 miles:** Take Alt 95 to Hwy 50 east at Fernley. Continue eastward on Hwy 50 ("The Loneliest Highway in the U.S.A.") along the Truckee River until its demise in the Carson Sink region.

**279 miles:** Fallon. Hwy 50 continues across the Carson Sink and then across Salt Wells Basin. Sand dunes and Pleistocene lake shore-lines on hills underlain by Cenozoic volcanic rocks can be seen to the north of the highway.

Sand Springs Range, exposures of Mesozoic sediments intruded by Cretaceous granitoids (to south of road) all overlain by Tertiary volcanic rocks (north of road).

Descend the range and cross Fairview Valley. Chalk Mountain to the north is Mesozoic marbles (white) intruded by granite.

Westgate, view north of road is of Mesozoic sediments overlain by Cenozoic volcanic rocks.

**336 miles:** Eastgate, Cenozoic volcanic rocks as far as you can see everywhere.

**345 miles:** Carrol Summit, more Cenozoic volcanic rocks. Descend the range and cross Smith Creek Valley.

**370 miles:** Railroad Pass in North Shoshone Range, more Tertiary volcanic rocks. Descend the range and then cross Reese River Valley. The impressive range to the east is the Toiyabe Range, in the heart of the northern Basin and Range, where we will stay the night.

**392 miles:** Town of Austin on the west flank of the Toiyabe Range, is sited in the Jurassic Austin Granite which includes a variety of older rocks including lower Palaeozoic allochthonous rocks of the Roberts Mountain Allochthon.

# MESOZOIC MAGMATIC, METAMORPHIC AND DEFORMATIONAL HISTORY OF THE CENTRAL GREAT BASIN

The Sierra Nevada batholith comprises the many, more widely dispersed, plutons to the east of the Sierra Nevada, which are scattered as far east as the Nevada-Utah state line. These constitute a volumetrically lesser part of the batholith, and many of these plutons were intruded at depth in the thick shelf sequence and exposed at the surface today because of Basin and Range faulting.

Our route takes us from Reno through Austin and Eureka to Ely (Fig. 17) across a series of Mesozoic structures developed in the back-arc region including the Luning-Fencemaker thrust belt, and the Eureka thrust belt, both in the "hinterland" of the Sevier thrust belt (Fig. 17). These are developed across older Paleozoic features including the upper Paleozoic Golconda allochthon, the lower Paleozoic Roberts Mountain allochthon, and its foredeep basin, the Antler foredeep basin (Fig. 3).

Basin and Range faulting and tilting has exposed variable crustal levels across this region, making it an ideal region for the study of the relation of Mesozoic magmatism, metamorphism and deformation at different structural levels. We have carried out detailed work with these goals in mind in the Yerington region (Dilles and Wright, 1988), Pine Forest Range of northwest Nevada (Wyld, 1991), the Toiyabe and Shoshone Ranges in central Nevada (Smith, 1989), the Schell Creek, Snake, and Deep Creek Ranges, and the Kern Mountains in eastern Nevada and Utah (Miller et al., 1988; Miller and Gans, 1989) (Fig. 17).

These data permit general discussion of the space-time patterns of magmatism, deformation and sedimentation across a transect of the Mesozoic arc and back-arc region between latitude 38°30' N and 42° N (Fig. 1), and allow general conclusions to be drawn about the mechanisms of Cordilleran evolution as outlined in Smith et al. (1993). Pertinent data for this transect are compiled on palinspastic (Basin and Range extension removed) strip maps (Fig. 18) and shown on a time-longitude graph (Fig. 19). The general conclusions from this database are summarized below, after Smith et al. (1993).

## Progression of Timing

Where the timing of deformation and metamorphism are well-constrained, we can universally resolve two distinct episodes of activity, both of which were accompanied by magmatism. In addition, the timing of these episodes varies systematically with position inboard of the continental margin across what is now the Great Basin.

The first episode of deformation and metamorphism occurred in the Middle to Late Jurassic. Across the transect, there is a consistent pattern in timing, with deformation and metamorphism starting and ending earlier to the west than to the east (Fig. 19). Thus, deformation and metamorphism is early to late Middle Jurassic (~180-165 Ma) in the magmatic arc of western Nevada, Middle to early Late Jurassic (~172-161 Ma) in the back-arc region of central Nevada, and Late Jurassic (~160-144 Ma) in eastern Nevada and western Utah. A nearly identical pattern in the age of plutonism across the transect is also seen, although plutonism in some areas appears to be somewhat longer lived than deformation (Fig. 19): plutons are late Early to late Middle Jurassic (~190-165 Ma) in western Nevada, Middle to Late Jurassic (~172-154 Ma) in central Nevada, and Late Jurassic (~165-144 Ma) in eastern Nevada and western Utah.

The second episode of deformation and metamorphism occured in the mid- to Late Cretaceous across most of our transect although deformation also continued into the very earliest Tertiary in the northern parts of the Sevier belt. As in the Jurassic, a consistent pattern of increasingly younger ages of cessation of Cretaceous deformation is seen across our transect from west to east (Fig. 19). The timing of inception of Cretaceous deformation, however, exhibits a less consistent pattern (Fig. 19). Thus, deformation occurred in the late Early Cretaceous (~108-100 Ma) in magmatic arc assemblages of western Nevada, in the late Early to middle Late Cretaceous (~100-77 Ma) in the back-arc region of central Nevada, in the middle Late Cretaceous (≥ 82-72 Ma) in eastern Nevada, and from late Early Cretaceous (~110-100 Ma) to Paleocene in the Sevier belt of western Utah. A similar pattern in ages of Cretaceous plutons is also seen across the transect although, again, plutonism spanned a longer period of time in some areas than does deformation (Fig. 19): plutons are ~108-100 Ma in western Nevada, ~100-77 Ma in central Nevada and ~125-75 Ma in eastern Nevada.

These data collectively indicate that a generally eastward migrating deformational and magmatic front characterized both Jurassic and Cretaceous orogenesis in the western U.S. Cordillera (Fig. 19). The shifting locus of deformation and magmatism is particularly clear for the Jurassic episode when both processes began and ended progressively earlier from west to east. A broadly similar pattern appears to have affected the region during the Cretaceous, as indicated by the fact that deformation and plutonism ceased at progressively younger ages from west to east. The data from the

Cretaceous, however, suggest that deformation began in the arc and then spread rapidly across the hinterland to the Sevier belt, occurring almost simultaneously across much of the transect at about 100 Ma. An apparent exception to this pattern is evident in eastern Nevada where radiometric age data suggest that deformation and metamorphism peaked as late as 82 Ma. There does, however, appear to have been some earlier deformation in this area based on the development of local syn-tectonic basins in the Eureka belt in the late Early Cretaceous. Regardless of the exact pattern of deformation across the region, it is clear that large areas of the arc and back-arc were deforming simultaneously during both episodes of orogenesis and that both episodes spanned lengthy intervals of time, on the order of 40 Ma.

### Relation Between Magmatism and Deformation

In areas where Mesozoic ductile deformation and metamorphism are not clearly regional in extent at the present levels of exposure (e.g. the arc in the Cretaceous, the Toiyabe Range and environs east to western Utah in the Jurassic), ductile structures generally appear to be limited to the metamorphic aureoles of plutons and metamorphism and strain are typically observed to be upgraded towards pluton margins. As noted previously, these relations argue strongly for pluton emplacement synchronous with metamorphism and fabric development, a conclusion that is supported by available geochronologic data. In all areas that have been studied in detail, structures have a regionally consistent orientation despite being only locally developed around plutons. These data collectively suggest that the plutons intruded a crustal section subjected to regional stress but that at these crustal levels, ductile strain occurred only in the thermal aureoles of syntectonic plutons.

In areas where ductile deformation and metamorphism are pervasive and widespread on a regional scale at present levels of exposure (the arc in the Jurassic and the hinterland in the Cretaceous), many of the same relations to plutons are observed despite the more regional extent and character of deformation and metamorphism. Thus, regional metamorphism increases in grade and strain is intensified towards plutons; geochronologic data generally indicate similar ages for both plutons and metamorphic fabric development.

The relations summarized above are observed in all areas of ductile deformation regardless of the varying ages of deformation and plutonism across the transect. Thus it appears that the voluminous and eastward-migrating magmatic front that characterized each episode of orogenesis played a vital role in the initiation and localization of deformation and in the resultant regional patterns of strain and metamorphism. Petrologic and isotopic data indicate that initially, in the Jurassic to Early Cretaceous, plutons in the arc and back-arc regions were metaluminous and derived from an asthenospheric source (arc region; Farmer and DePaolo, 1983) and a subcontinental lithospheric mantle source (back-arc region; Wright and Wooden, 1991). Patterns of Mesozoic deformation and metamorphism in the arc and back-arc regions were thus initially dictated or influenced by a flux of mantle-derived heat. This situation changed in the Late Cretaceous when crustal thickening in the hinterland of the Sevier belt led to the widespread production of peraluminous magmas by crustal melting (e.g., Farmer and DePaolo, 1983; Wright and Wooden, 1991). This additional local influx of magma helped transport heat to higher crustal levels, probably resulting in the widening of the region of ductile deformation and increasing the time span of active deformation. We thus suggest that the feedback between crustal thickening and magmatism in the hinterland prolonged deformation in this area and acted to blur systematic progressions in the timing of events across this part of the orogenic belt.

## Driving Forces

The timing of these two major events sheds light on the nature of the driving forces behind Cordilleran orogeny. Both Jurassic and Cretaceous events correlate very closely with periods of rapid northwestwards motion of North America within an absolute reference frame as well as high orthogonal convergence rates along the western margin of North America (Fig. 19; Gordon et al., 1984; Page and Engebretson, 1984; Engebretson et al., 1985). Although the data discussed above clearly indicate diachroneity of deformation across the width of the orogen, the regular progression eastward of deformation together with the prolonged time span of both the Jurassic and Cretaceous events argue strongly that relatively constant plate motions over periods of tens of millions of years must be the prime driving mechanism for deformation. Abrupt changes in plate motion (May and Butler, 1986; May et al., 1989) are important in that they probably dictate the inception and cessation of such deformational events. In contrast, the collision of arc terranes with the western margin of North America, which has often been proposed as the major driving force for continental margin orogeny (e.g. Monger et al., 1982; Schweickert et al., 1984; Day et al., 1985; Edelman, 1991), would most likely produce brief and localized events only along the western margin of the continent. We thus suggest that rapid absolute plate motion coupled with high orthogonal relative convergence rates along the western margin of North America in the Middle-Late Jurassic and mid-Late Cretaceous resulted in long timespans where magmatism was fairly voluminous and the overiding continental plate was under regional compression; deformation occuring during these events produced the two distinct and protracted deformational events that are now documented in the western U.S. Cordillera.

**Stop 1.2 Austin Summit**

Drive through Austin and pull off Highway at Austin Summit where there is plenty of room to park, for a look at the Jurassic Austin Granite and discussion of the magmatic, metamorphic, and deformational history of the Toiyabe Range and environs.

In the Toiyabe and Toquima Ranges of central Nevada (Fig. 20), several Middle to Late Jurassic plutons intrude a pre-Mesozoic structural sequence consisting of autochthonous Paleozoic continental margin sediments and overlying allochthonous Paleozoic deep marine sediments. These plutons have K/Ar ages ranging from 154 to 172 Ma (Krueger and Schilling, 1971; Silberman and McKee, 1971). The nature of deformation and metamorphism around the Jurassic plutons has been studied in detail only in the central Toiyabe Range near the Austin pluton (Smith, 1989), dated at 161 Ma by the K/Ar method (Krueger and Schilling, 1971); U/Pb analysis of zircon is also permissive of an ~160 Ma age, but the data are complex and the zircons are discordant (J. E. Wright, unpublished data). The pluton has a 2-5 km wide metamorphic aureole within which a metamorphic foliation is developed that dips steeply southward, approximately parallel to the intrusive contact, changing from a spaced slaty or solution cleavage at the outer edges of the aureole to a penetrative schistosity immediately adjacent to the pluton. These relations, coupled with textural observations on the growth of contact metamorphic porphyroblasts with respect to metamorphic fabric development, indicate synchronous intrusion, metamorphism and deformation during the early Late Jurassic (Smith, 1989). There is no evidence of Middle to Late Jurassic deformation in the Toiyabe Range outside of the Austin pluton aureole (Smith, 1989), implying that deformation and metamorphism were localized because of heating by the intrusion. The 172 Ma (K/Ar) Grass Valley pluton, 35 km northeast of Austin, also has a 1-2 km wide envelope of foliated country rocks (D.L. Smith, unpublished data) and published descriptions of the two Middle Jurassic plutons in the Toquima Range suggest that they too, have only narrow aureoles (Stewart et al., 1977; Kleinhampl and Ziony, 1986). Deformation in this area thus appears to have developed only locally, in the thermally softened contact zones of syntectonic plutons.

Continue on another 4 miles to Bob Scott's Summit, turn off to the left into the campground (Bob Scott Campground) where we have reserved a group campsite.

## DAY 2 -

In the morning we have a long but scenic drive through typical Basin and Range country to Ely, Nevada. On the way we will have a brief stop in Eureka, an historic mining town for photos and a half hour stroll. In Ely we will refuel, then drive over the Schell Creek Range to the Snake Range where we have five stops planned. The first will be an overview stop and discussion of the Tertiary history of this mountain range. The next two stops involve hiking. Time permitting we will do both of these, to see the highly deformed Jurassic Old Mans Pluton and second a hike up through cliffs of mylonitic quartzite and schists in Hendry's Creek. From here we will drive south to our camp in the Great Basin National Park, Southern Snake Range.

## ROAD LOG - BOB SCOTT CAMPGROUND TO SNAKE RANGE

Reset odometer to **0.0**.

Coming down from Bob Scott Summit, we are entirely in Jurassic granite to the valley floor. Cross Big Smoky Valley and head east across the Simson Range (Hickison Summit - Cenozoic vocanic rocks) and across Monitor and Antelope Valley (look for herds of wild horses).

**Stop 2.1** Eureka
**60 miles:** Enter Eureka. An historic mining town where we will stop for a break.
East from Eureka, we drive through a variety of Lower Paleozoic stratified carbonate shelf deposits as well as Tertiary sediments (soft, light outcrops). Palaeozoic strata here are involved in a series of folds and thrust faults, part of the Eureka thrust belt (Figs. 17, 18).
**72 miles:** Start descending into the next valley, the Newark Valley.
**82 miles:** Pancake Summit.

**93 miles:** Antelope Summit, White Pine Range. Stratified Mississippian clastic and upper Palaeozoic shelf-facies carbonates deposited in foredeep basin of Antler thrust belt. Come down off range and cross Jakes Valley.
**117 miles:** Robison Summit, Butte Mountains, underlain by Cenozoic volcanic rocks. Drive through gorge - upper Paleozoic Ely Limestone, and Permian Arcturus Formation form most of the exposures in this canyon. Small Cretaceous stocks 120 - 110 Ma are found just before Ely, at the Ruth Copper pit on the south side of the road.
**137 miles:** Enter Ely and continue to the east side of town, gas up at Chevron. Drive out of town, following the signs for Highway 50 East, and head southeast towards Baker and Delta. across Steptoe Valley.
**162 miles:** Conners Pass, Schell Creek Range. Exposures in this range are mostly Lower Paleozoic strata until we descend to Spring Valley. As you approach and cross Spring Valley there are good views east of the northern and southern Snake Ranges. The giant snowcapped peak in the southern Snake Range is Wheeler Peak, underlain by Cambrian Prospect Mountain Quartzite.
**183 miles:** Sacramento Pass.

## GENERAL OVERVIEW OF THE GEOLOGY OF THE N. SNAKE RANGE

### Introduction

The Snake Range, in eastern White Pine County, Nevada, is a 150-km-long, north-south trending mountain range in the northern Basin and Range province. Sacramento Pass divides the range into two main parts, the northern and southern Snake Range (Figs. 21, 22).

The northern Snake Range includes the Mount Moriah Wilderness, established in 1989, and lies north of the Great Basin National Park, established in 1986 in the southern Snake Range. The steep-walled canyons and rugged ridgelines of the northern Snake Range provide access to the unusual flat-topped "Table" at 11,000 feet and Mount Moriah at 12,067 feet, forming some of the most scenic hiking country in the Basin and Range. The broad, arch-like physiography of the northern Snake Range is shaped by its geology, which is unique compared to other mountain ranges in the region. The northern Snake Range is now considered a classic example of a Cenozoic "metamorphic core complex" (e.g., Coney, 1979). The most prominent structural feature of the range is the northern Snake Range décollement (NSRD), a low-angle fault that juxtaposes an upper plate of complexly normal-faulted Paleozoic and Tertiary strata against a lower plate of ductilely attenuated metasedimentary and igneous rocks (Fig. 21). The NSRD defines a north-south trending asymmetric dome with about 1.6 km of structural relief . The age, origin, and tectonic significance of the NSRD have been topics of continuing debate since the fault was first described by Hazzard et al. (1953) and Misch (1960). Although the origin of the NSRD and core complex detachment faults in general remain controversial, there commonly is agreement that these complexes provide excellent exposure of both brittle and ductile structures formed as a result of large-magnitude crustal extension.

## Geologic Setting and Previous Work

Both the northern and southern Snake Range are underlain primarily by Late Precambrian to Permian miogeoclinal shelf strata deposited along the subsiding western continental margin of North America. Miogeoclinal strata in the footwall of the NSRD range in age from Late Precambrian to Ordovician and generally define a broad, north-south trending antiform. These rock units are relatively unfaulted but record a polyphase history of ductile deformation, metamorphism, and intrusion. The hanging wall or upper plate of the NSRD includes Middle Cambrian to Permian miogeoclinal rocks, as well as Tertiary sedimentary and volcanic rocks. In striking contrast to the lower plate, these rocks are little metamorphosed but highly faulted and tilted by multiple generations of normal faults.

In many ways, the evolution of ideas concerning the origin of the NSRD charts the progress of our understanding of the extensional history of the Cenozoic Basin and Range province as a whole. Until the early 1970's, most low-angle faults like the NSRD in the western U.S. were mapped as thrust faults. The Snake Range décollement was first described by Hazzard et al. (1953). A more detailed study of the geology of the northern Snake Range by Misch (1960) and Misch and Hazzard (1962) followed as part of a regional survey of the geology of eastern Nevada. They noted that the major structure in the northern Snake Range was a low-angle "décollement" that separated an autochthon of strongly metamorphosed rocks from an allochthon of faulted and folded, but largely unmetamorphosed, carbonate rocks. Nelson (1966, 1969) mapped the northern end of the range as part of a regional mapping project that included the Kern Mountains and southern Deep Creek Range. He mapped the décollement of Misch (1960) and identified lower-plate schists and marbles as metamorphosed equivalents of Precambrian and Cambrian-Ordovician miogeoclinal units and upper plate rocks as Cambrian to Permian carbonate rocks and younger Tertiary rocks. Nelson proposed that rocks in the lower plate (his “autochthon”) recorded three deformational events, the youngest of which involved the generation of mylonites and formation of folds that he assigned to the Mesozoic. He believed that these structures were associated with east-directed thrusting along the décollement horizon. The position of the Snake Range in the hinterland of the Cretaceous Sevier orogenic belt led these and later workers to relate the low-angle "décollement faulting" in the northern Snake Range to Mesozoic thin-skinned thrust faulting further east, where the NSRD represented the basal shearing-off plane for these thrusts (Misch, 1960; Miller, 1966). However, Hose and Danes (1973) and Hintze (1978) recognized that the deformation in the upper plate was dominated by normal faulting and attenuation of the stratigraphic section rather than by shortening and thickening of the rock column. This led them to propose a model wherein the Snake Range and environs represented an uplifted hinterland, where extension was linked to coeval shortening in the foreland via a basal detachment fault now exposed as the NSRD. Armstrong (1972) was the first to suggest that many of these faults might be Tertiary rather than Mesozoic in age, and therefore unrelated to Mesozoic thrust faulting. Armstrong specifically cited geochronologic and stratigraphic relations from the southern Snake Range as some of his principal evidence for a Cenozoic age for the Snake Range decollement. Coney (1974) suggested that at least some of the lower plate deformation in the northern Snake Range might be related to the Snake Range décollement. He studied folds in marble mylonite beneath the NSRD and proposed “quaquaversal” or radial sliding of upper plate rocks off the northern Snake Range.

Hose and Blake (1976) compiled a 1:250,000 scale geologic map of White Pine County, Nevada, which included the first compiled geologic map of the northern Snake Range based largely on reconnaissance mapping by Hose. They mapped the low-angle NSRD in its entirety, which they described as separating a lower plate of undifferentiated Lower Cambrian quartzite and pelite and Middle Cambrian marbles from an upper plate of complexly faulted Middle Cambrian to Permian carbonate rocks. Hose and Blake (1976) proposed that two metamorphic and deformational events were recorded in lower plate rocks, a post mid-Jurassic to pre-early Eocene intrusive and high-grade metamorphic event followed closely by a low-grade metamorphic event associated with the development of a strong penetrative foliation and WNW-ESE trending mineral elongation lineation. This was followed by post-early Eocene movement along the NSRD and associated faulting in the upper plate. As part of the Wilderness RARE II study, additional mapping of the southern part of the northern Snake Range was carried out at a scale of 1: 62,500 (Hose, 1981). This publication pointed out the magnitude of structural thinning of upper plate units by normal faulting. In an influential paper by Wernicke (1981) on the geometry and kinematic significance of extensional detachment faults, the Snake Range décollement was utilized as a key example of a large-displacement, eastward-rooting low-angle normal fault.

Studies in the northern Snake Range based at Stanford University began in 1981 and are still ongoing. Miller et al. (1983) and Gans and Miller (1983) suggested that the basic structural relationships in the northern Snake Range were best explained as extensional in origin and Cenozoic in age. For the first time, the Late Precambrian lower plate units in the central and southern part of the range were identified and correlated, bringing to light the incredible tectonic stretching or attenuation of these rock units by ductile deformational processes. The upper plate units were shown to have been affected by multiple generations of predominantly east-dipping normal faults, and it was pointed out that over much of the range there appeared to be near stratigraphic continuity between the oldest units present above and youngest units present below the décollement. These and other relations led Miller et al. (1983) to question the need for significant displacement on the decollement and to propose instead that the NSRD originated as a subhorizontal ductile-brittle transition zone between a brittlely extending upper plate and a ductilely stretching lower plate. This interpretation was challenged by Bartley and Wernicke (1984), who specifically proposed that the NSRD represented a low-angle normal fault or shear zone with 60 km or more displacement that brought lower plate rocks up and out from under a thrust plate in the Sevier belt to the east. In their model, the near continuity or lack of stratal ommission between upper and lower plate rock units cited by Miller et al. (1983) was strictly fortuitous. Gans and Miller (1985) responded to this alternative interpretation by citing additional regional stratigraphic and structural relations that created difficulties with their proposed model.

Further studies in the northern Snake Range expanded our geologic mapping and utilized structural and kinematic analyses, seismic reflection profiling, metamorphic petrology, and extensive geochronology and thermochronology in order to help resolve issues as to the amount of displacement and initial angle of the NSRD, as well as the age of lower plate deformation and its geometric and kinematic relationship to the evolving NSRD (Rowles, 1982; Gans and Miller, 1983; Miller et al., 1983; Grier, 1983; Gans et al., 1985; Geving, 1987; Lee et al., 1987; Miller et al., 1987; Miller et al., 1988; Miller et al., 1989; Gans et al., 1989; Huggins, 1990; Lee, 1990; Lee and Sutter, 1991). Our geologic mapping at scales of 1:12,000 and 1:24,000 over an eleven-year period (1981-1992) was the first detailed mapping to be completed in the range. During the first half of this project, mapping was carried out on 1:16,000 black and white and 1:24,000 color aerial photographs with compilation onto orthophotoquadrangles because topographic maps were not yet available for the region.

Our studies have shown that lower plate rocks consist of metamorphosed Late Precambrian to Lower Cambrian quartzite and pelite and Middle Cambrian to Ordovician marbles that correlate in a straight-forward fashion to less deformed and metamorphosed sections in the adjacent Schell Creek, Deep Creek, and southern Snake Ranges. Jurassic and Cretaceous granitic plutons and Tertiary dike swarms intrude lower plate units. Lower plate rocks record at least three metamorphic and deformational events. The first metamorphic event, of Jurassic age, is best preserved along the southern flank of the northern Snake Range. Here, Late Precambrian and Lower Cambrian quartzites and metapelites have been intruded and contact metamorphosed by a mid-Jurassic plutonic complex. Structural fabrics associated with this event are strongly overprinted by superimposed Cretaceous and Cenozoic fabrics. The second metamorphic event, of Late Cretaceous age, affected a much broader region of the lower plate. A series of mineral-in isograds mapped along the eastern side of the range indicates that the grade of metamorphism increases from greenschist to amphibolite facies from south to north and with structural depth in the succession (Geving, 1987; Huggins, 1990). A Late Cretaceous pegmatite and aplite dike swarm was intruded during this metamorphic event, which has been dated as approximately 82-78 Ma (Huggins and Wright, 1989; Huggins, 1990). Structural fabrics associated with this metamorphic event have also been strongly overprinted by Cenozoic fabrics, making their analysis and interpretation difficult. However, on the northwestern flank of the range, Tertiary strain decreases and eventually dies out. Here, west-dipping foliations, minor thrust faults, and a map-scale fold now inferred to be of Cretaceous age (P. B. Gans, 1992, unpublished data) are preserved. Lower to upper greenschist-facies metamorphism of Oligocene(?) to Miocene age (Lee and Sutter, 1991) strongly affected much of the lower plate, causing retrogression of older mid-Jurassic and Late Cretaceous metamorphic assemblages. The Tertiary metamorphic event was accompanied by vertical thinning and horizontal stretching, resulting in a subhorizontal, bedding-parallel mylonitic foliation and WNW-ESE trending stretching lineation. This foliation is axial planar to isoclinal, recumbent folds in the northern half of the range. A strong gradient in the amount of deformation or strain of lower plate rock units developed during this youngest event. Strain increases dramatically from west to east across the range. Mesoscopic, microstructural, and petrofabric studies on lower plate rocks were utilized by Lee et al. (1987) to modify the in-situ pure shear model proposed by Miller et al. (1983). Lee et al. (1987) proposed a strain path whereby pure and simple shear (top to the east) acted in unison and in sequence in the lower plate and that this strain was intimately tied to the evolution of the NSRD, ultimately leading to slip along this surface in the brittle regime. Structural studies by Gaudemer and Tapponnier (1987) were used to promote a model whereby lower plate deformation occurred entirely by simple shear. As pointed out by Lee et al. (1987), the question of simple versus pure shear remains controversial, as most structural and petrographic observations used to resolve these questions do not yield unique interpretations. In sum, important questions still remain regarding the exact age of development of lower plate fabrics, whether they are developed as a consequence of pure and/or simple shear, and what the exact kinematic relation is between these fabrics and the evolving NSRD.

In the overlying upper plate, unmetamorphosed to weakly metamorphosed Middle Cambrian to Permian carbonate rocks have been attenuated by at least two sets of imbricate normal faults whose exact ages remain poorly constrained but are at least in part Oligocene-Miocene in age. In the Sacramento Pass area and at the northern end of the northern Snake Range, Tertiary alluvial fan conglomerate, lacustrine deposits, and volcanic rocks are cut by normal faults that merge along strike with the NSRD, demonstrating a Tertiary age for much if not all of the extensional faulting (Gans et al., 1989). Parts of the upper plate of the NSRD are characterized by a systematic history of east-directed normal faulting that resulted in successive northwestward tilting about a common axis in response to WNW-ESE extension, parallel to that recorded by the ductile deformational fabrics in the lower plate (Miller et al., 1983). However, the amount of strain indicated by these faults and the amount of rotation related to faulting varies across the range, as does the direction of tilting or rotation. Normal-sense movement along the NSRD during the time span of upper plate faulting is indicated, as only a few mapped faults actually cut and offset the NSRD. Motion along the NSRD is believed to have been down-to-the east and resulted in the juxtaposition of the less metamorphosed and mostly younger upper plate rocks down upon the more highly metamorphosed and generally older lower plate rocks. An important exception to this occurs in the northern part of the range where Middle Cambrian and younger rocks of the upper plate routinely overlie Late Cambrian and locally Ordovician strata in a lower plate position.

Although we previously postulated that most of the movement on the NSRD was Oligocene to early Miocene (e.g. Miller et al., 1983; Gans and Miller, 1983; Gans et al., 1985), we now know that motion on the eastern part of this fault system is middle Miocene in age. This conclusion is based on an apatite fission track study (Miller et al., 1989, 1990) that suggests that lower plate rocks were rapidly cooled and exhumed between 20-15 Ma and by the fact that Miocene or younger sedimentary sequences along the northern, eastern, and southern flanks of the range are cut and tilted by a set of normal faults that we infer either to cut or to sole into the NSRD in the subsurface (Miller et al., 1989, 1990)(Gans et al., 1989).

We now think that the combined data on the deformational history of upper and lower plate rocks indicate that the NSRD is a composite structure rather than a single fault; thus it was never simultaneously active over its entire mapped extent. East-dipping normal faults along the eastern flank of the range were likely active in mid-Miocene and younger times and perhaps originated as steep faults that have since been rotated to present low dips. If the high strain rocks and mylonites of the lower plate represent more than one Tertiary event, our studies to date have not been able to resolve these with confidence. Limiting factors include the resolution of available geochronologic techniques, given the complex thermal and deformational histories of the minerals available for dating and the inability of structural studies to distinguish more than one superimposed event if these are developed at low angles to one another. Clearly, the NSRD represents the end result of an involved history of extensional strain at both ductile and brittle levels of the crust and that more than one episode of faulting is responsible for the relative uplift and exposure of ductile extensional fabrics in the range. However, more sophisticated studies and modeling of existing data are necessary in order to fully understand the kinematic history and mechanics of deformation and faulting.

6.4 mi after Sacramento Pass summit, turn left onto graded dirt road (beneath power line), reset odometer. Follow power line for ~ 0.5 mi, turn left onto dirt track. As you drive north on this road, the low hills on the right expose the basal unconformity for the Tertiary section. Here, Pennsylvanian Ely Limestone (light tan knobby hills) is overlain successively by a thin ( x m) interval of reddish, well rounded conglomerate (Tc1), ~30 m of highly altered and oxidized bio-hornblende dacite lava flows (brick-red resistant rib of rocks), and > 500 m of light tan to buff, well bedded lacustrine limestone ( Fig. 25). This section (described in detail by Grier (1983)) has been tilted to the northwest and is repeated on a series of top to the east normal faults. We will examine components of this section and discuss its age and tectonic significance later in the day.

## GEOLOGY OF THE SACRAMENTO PASS AREA

Sacramento Pass is a topographically low region between the greater elevations of the northern and southern Snake Ranges, east central Nevada (Fig. 22). Sacramento Pass is underlain by generally non-resistant Cenozoic strata which were deposited in a fault-bound depression between the two parts of the range, informally referred to here as the Sacramento Pass section. The stratigraphy and sedimentology of the Sacramento pass section, described in detail by Grier (1983, 1984) provides insight into the history of Basin and Range faulting which led to the uplift of rocks in the surrounding mountains and the development of the present-day topography of the region.

The Tertiary Sacramento Pass section is at least 1500 m thick and includes volcanic rocks, lacustrine limestone, and alluvial fan deposits. In addition, impressive horizons of monolithologic breccia and large coherent blocks of Paleozoic strata occur within the section and are interpreted as landslide deposits (Grier, 1984). The base of the Tertiary section rests disconformably upon upper Paleozoic strata. The best exposures of the basal unconformity and the lower units in the Tertiary Sacramento Pass section are found just north of Weaver Creek (Fig 1). Here, earlier conglomerate of unknown age, containing clasts derived exclusively from uppermost Paleozoic strata, rests unconformably above the Pennsylvanian Ely Limestone. The basal conglomerate is overlain by volcanic rocks which include latite flows and rhyolite tuff dated by Hose and Blake (1976) as Oligocene. These units, in turn, are conformably overlain by lacustrine limestone that grades upwards and interfingers laterally with alluvial fan deposits of probable Miocene and younger age.

The alluvial fan deposits of the Sacramento Pass section contain varied clast types attesting to uplift of surrounding ranges (Grier 1983, 1984). The section is now tilted moderately to the west and repeated by a series of normal faults which produce extension in a WNW-ESE direction (Fig. 1). Tilted Tertiary strata are in normal fault contact with bedrock of the southern and northern Snake Range.

A Miocene or younger age for most of the Sacramento Pass section is supported by fission track ages of apatite from cobbles of granite and metasedimentary rocks in the conglomerate that range from 15-20 Ma. These ages attest to the very young age of continued faulting which uplifted the Snake Range and cut and tilted the Tertiary section itself.

Stop 2.2 **Sacramento Pass**

**2.0 miles:** Park on saddle. Overview of the geology of the northern Snake Range and the Sacramento Pass region. From this vantage point, we have an excellent view of some of the important geologic features of the northern Snake Range. Looking to the north, there are low buff coloured hills of Tertiary section in foreground. This section is at least 2.5 km thick, dips predominantly to the west or northwest, and has been repeated on a series of at least 3 major east-dipping faults (Fig. 22). Further to the north, behind the Tertiary section are more resistant grey, white , and brown Paleozoic limestone and dolomite. These represent thin fault slivers of the upper plate of the northern Snake Range Decollement (NSRD) and lie between the Tertiary sections to the south and lower plate rocks to the north. The lower plate rocks underlie the forested slopes on the southern flank of the northern Snake Range and here consist mainly of different phases of the Jurassic composite Silver Creek granite. Dipping grey cliffs on the skyline of the northern Snake Range (above the forested slopes) are middle Cambrian to Ordovician limestones in the upper plate of the NSRD. The NSRD is visible to the NNE as a ledge of marble mylonite - basically a highly deformed roof pendant of the Silver Creek Granite. It is nearly flat lying at the crest of the range, and then rolls over towards the south such that at the southern flank of the range it dips steeply southward beneath where we are standing. Continue north on dirt road.

**2.7 miles:** Nice view of the middle part of the Sacramento Pass section. Interfingering lacustrine limestone (light colored) and coarse conglomerate (pinkish tan) with occasional large slide blocks of Paleozoic limestone and dolomite (dark brown and grey).

**3.1 miles:** Intersect better graded dirt road (Silver Creek Road), reset odometer to **0.0** and turn left.

Now driving up section through the upper (conglomeratic) part of the Sacramento Pass Tertiary section

**Stop 2.3** Sacramento Pass Area

**1.5 miles:** Cross east-dipping fault - exposed in saddle on right. This is one of several large east-dipping faults in the Sacramento Pass area, each fault dips east 20-50° and has 1-2 km of top-to-the-east displacement. Here, the fault juxtaposes conglomerate from the upper part of the Tc section against lacustrine limestone w/ slide blocks from the lower/middle part of the Tertiary section. We will stop for a quick examination of one of the large slide block/megabreccia sheets. Continue driving west on Silver Creek Road

**2.0 miles:** Intersection with Miller Basin road. Turn left and head back out to highway 50

**5.2 miles:** Intersection with highway 50, turn left (east), reset odometer to **0.0**

**0.2 miles:** Rest area on right

**3.9 miles:** Turnoff to Strawberry Creek.

**7.1 miles:** "Y intersection" w/ Nowhere Cafe on the left. Stay on left fork - Hwy 50 towards Delta - and turn left ~ 150 yards past the Y intersection on a good graded dirt road, head north up Snake Valley. Reset odometer to **0.0**

**2.1 miles:** Silver Creek Ranch. Drive slowly through the ranch please.

**5.2 miles:** Turn left at tire.

**7.0 miles:** Cowpond on left, take road to right. Spectacular view of the NSRD as we drive into our next stop. Marble mylonite beneath the NSRD forms a conspicuous ledge halfway up the mountain. Tilted and faulted Paleozoic strata above the marble tectonite ledge are clearly truncated by the NSRD. The dark rocks beneath the marble mylonite are biotite granite of the Silver Creek Pluton.

**8.5 miles:** Sharp right turn.

**9.1 miles:** Cattle pond, Y in road, bear left. As we drive into Old Man Canyon, the low hills on either side of the drainage consist of upper plate Devonian to Pennsylvanian carbonates. The decollement dips beneath us.

**Stop 2.4** Old Mans Canyon

**10.7 miles** Park. The NSRD is the major mapped structure in this part of the northern Snake Range and exposures of this impressive fault system with its underlying marble mylonites in the Old Mans Canyon area constitute its type area as described by Misch (1960) and Misch and Hazzard (1962)(Fig. 24). It is nearly flat-lying where it is exposed at higher elevations in the northern half of the quadrangle but descends steeply southward and eastward to the southeastern flank of the range, defining the southern and eastern side of its broad domical structure (Fig. 24). The complex system of east-west-striking faults along the southern edge of the range may be younger than (and thus cut) the NSRD mapped at higher elevations. Conglomerate units that are cut by this system of faults are mid-Miocene and younger, placing an older age limit on these faults.

All of the lower plate rock units in the Old Mans Canyon region have been deformed together in the Cenozoic and possess a pervasive sub-horizontal mylonitic or gneissic fabric and a WNW-ESE trending elongation lineation. We will hike through the deformed Old Mans Pluton up to the NSRD. Towards the decollement, both the pluton and pendants of marble become mylonitized. You can actually put your finger on the decollement surface at this stop, but it is a fairly subtle fault. Cambrian Pole Canyon Limestone above the decollement is recrystallized and brecciated whereas the same limestone beneath it is ductilely deformed.

The large gneissic pluton of Jurassic age that underlies the Old Mans Canyon region is a composite pluton and consists of two distinct bodies informally referred to as the Old Mans pluton and the Silver Creek granite. The Old Mans pluton is the older and easternmost of the two and grades from a hornblende-bearing diorite in its easternmost exposures to a biotite tonalite and lesser granite on the west. Abundant pegmatite and aplite dikes, which emanate from the top of the younger Silver Creek granite, intrude overlying and adjacent tonalite of the Old Mans pluton. U-Pb ages on several fractions of zircon from 3 localities in this composite pluton indicate a Jurassic age of intrusion, approximately 155±5 Ma ago (J. E. Wright, unpublished data). A screen of Late Precambrian to Cambrian age metasedimentary rocks separate the two main phases of the pluton and is well-exposed along the eastern side of the Silver Creek drainage. These rocks record two superimposed metamorphic events that are spatially related to the two adjacent plutons (Miller et al., 1988). Muscovite, biotite, staurolite, garnet, and andalusite grew during the first metamorphic event ($M_1$), which was associated with the intrusion of the Old Mans pluton. In most places these minerals were severely retrograded during the second metamorphic event ($M_2$), which was synchronous with the development of a much stronger foliation. Muscovite, biotite, staurolite, andalusite, and cordierite grew during the second metamorphic event. Andalusite and cordierite are developed in close proximity to the margin of the Silver Creek Granite. Rare kyanite is present in the more distal aureole of the younger Silver Creek granite and occurs as small, euhedral crystals growing in retrograded $M_1$ andalusite and in $M_2$ cordierite. Late stage growth of neoblastic kyanite suggests cooling at constant pressures at or below the aluminum silicate triple point. Both the pluton and the inferred Jurassic-age fabrics developed in these rocks are cut by a younger, sub-horizontal fabric of Cenozoic age.

**ROAD LOG CONTINUED**

Retrace your route, driving back out to the main Snake Valley Road.
Reset odometers to **0.0**, as you turn left (north) on Snake Valley Road. As we drive north along the east flank of the Snake Range, the rugged brown hills you see in the foreground are upper plate klippen of mid-Palaeozoic dolomite and Ordovician Eureka Quartzite.

Behind the klippen are light gray-yellow, smooth slopes underlain by stretched lower plate Prospect Mountain Quartzite. The NSRD projects above the smooth hills and beneath the vegetated cliffs of upper plate rocks in the higher ridgelines.

**1.0 miles:** Welcome to Utah

**2.2 miles:** Main road veers right, take left fork for 100 yards and then bear left into Hatch Rock Quarry.

**5.2 miles:** Take left fork. Cavernous, brecciated Ordovician-Silurian dolomite in the upper plate forms the ridge line to the south. The SRD is defined by a thin ledge at 10:00. The light colored slopes below the thin ledge and straight ahead are stretched, lower plate Prospect Mountain Quartzite, the decollement dips about 15° towards us so that it barely clips the top of the hill at 2:00 where there is a klippe of Devonian Guilmette Limestone, and projects under brecciated Cambrian limestone to our right.

**5.3 miles:** Take right fork.

**5.6 miles:** Take left fork, follow signs towards trailhead. As we drive into Hendrys Creek, the smooth light colored slopes and ridges on either side of the road are underlain by mylonitic Proterozoic and Lower Cambrian quartzite and schist - part of the lower plate of the northern Snake Range Decollement (NSRD). Very dark colored, more rugged looking hills and ridges on either side are highly faulted and brecciated klippen of Middle Cambrian to Devonian limestone and dolomite of the upper plate. The NSRD is well exposed at the base of these klippen - often as a distinct ledge. The NSRD dips gently eastward about 15° beneath Snake Valley, where it has been imaged seismically (Allmendinger et al., 1983; McCarthy, 1986) and can be projected just above and parallel to the dip-slope ridge crests of quartzite to the exposures of the NSRD at higher elevations in the range.

**6.1 miles:** Cattle gate. The cliffs to the left and in front of us expose impressively thinned Prospect Mountain Quartzite (uppermost quartzite unit). Quartzite and schist units beneath the Prospect Mountain belong to the Precambrian McCoy Creek Group. The distinctive steel-blue/grey schist beneath the Prospect Mountain is the Osceola Argillite, the upper-most unit of the McCoy Creek Group.

**14.6 miles:** Park at trailhead.

**Stop 2.5** Hendry's Creek

Moderate hike ~ 2hrs From here, we will hike up the trail a short ways, cross the stream and then hike up the north side of the canyon, through the highly attenuated upper Precambrian-Lower Cambrian sequence of alternating quartzite and schist, to the top of the ridge for an overlook

## GEOLOGY OF THE HENDRY'S CREEK AREA

Hendry's Creek provides access to spectacular exposures of flaggy quartz mylonites and interbedded schist that display a wide variety of ductile-to-brittle mesoscopic and microscopic structures, now considered typical of rocks having undergone large amounts of (extensional) strain (Gans and Miller, 1983; Miller et al., 1987; Lee et al., 1987). All rocks in this drainage possess a very prominent foliation subparallel to bedding and a well-developed WNW-ESE trending mineral elongation or stretching lineation. Lower plate rock units are so attenuated that their present thicknesses are less than one tenth of their original stratigraphic thicknesses. The flaggy and conspicuously jointed mylonitic quartzites are quarried for building purposes and provide a uniquely aesthetic decorative stone of many uses.

Upon careful examination, it can be seen that the most obvious deformational fabrics in the schist and quartzite of Hendry's Creek postdate the growth of most metamorphic minerals in these rocks. This is evidenced by the conspicuous chlorite pressure shadows developed around garnets in a direction parallel to the stretching lineation, by pulled apart segmented staurolite and biotite porphyroblasts, and by well-developed sets of extensional crenulation cleavage in the schists that cut and rotate coarse-grained micas. From these observations it can be concluded that the strain that resulted in the extreme attenuation of lower plate units occurred at lower metamorphic grade, probably entirely at lower greenschist-facies conditions (Gans and Miller, 1983; Lee et al., 1987). Although the high strain mylonitic foliation in lower plate rocks is generally developed subparallel to original bedding and unit contacts, an exception to this occurs further up Hendry's Creek, where some of the deepest exposed levels of the lower plate are involved in a series of open to isoclinal recumbent folds whose axes trend parallel to the lower plate stretching direction. Similar folds in the Hampton Creek drainage to the north have been ascribed by Rowles (1982) to a component of constrictional strain operating during the overall attenuation and stretching of the units. The sub-horizontal mylonitic foliation is axial planar to these folds. In quartzite units, original bedding is folded by these structures, but in the intervening schists, the planar surfaces that define the folds are an older schistosity that is generally tightly crenulated.

Peak metamorphism has been dated as Late Cretaceous in age (Lee and Fischer, 1985; Huggins and Wright, 1989; Huggins, 1990) and resulted in the growth of coarse-grained garnet and staurolite in schist units of the McCoy Creek Group. This metamorphic event is responsible for the series of mineral-in isograds mapped along the eastern side of the range; metamorphism increases northward and with depth in the stratigraphic succession (Geving, 1987; Huggins, 1990). Deformational fabrics produced during this event are strongly overprinted by Tertiary fabrics, thus cannot be measured and analyzed. Tertiary strain that resulted in the extreme attenuation of the units in Hendry's Creek occurred at lower metamorphic grade, collapsing older metamorphic isograds (Geving, 1987; Miller et al., 1988).

Return to vehicles and then drive back out to main Snake Valley road, turn right and return to Baker, make a right and take paved road (488) ~4.5 mi to Lehman Caves National Monument campground for the night.

## DAY 3

Today we will hike up to look at deformed Cretaceous granites and pegmatites in lower plate rocks of NSRD in Smith Creek, north of Hendry's Creek in the northern Snake Range. Then up to Marble Wash to look at mylonitic marbles and deformed Cenozoic mafic dikes in the lower plate. From here we continue north to the Kern Mountains where we will see deformation of Cenozoic dikes and their country rocks in the aureole of a Cenozoic granite. Our final stop will be in Blue Mass Canyon in the Tungstonia granite, a Cretaceous two mica granite in the western Kern Mountains. We will camp here tonight (not an improved campground).

## ROAD LOG - LEHMAN CREEK CAMPGROUND TO NORTHERN SNAKE RANGE, KERN MOUNTAINS, AND SCHELL CREEK RANGE

Leave campground, and head north to Hwy 50. Head north again along Snake Valley Rd. Reset odometer to **0.0** at Hendry's Creek turnoff/Snake Valley Road intersection.

**4.1 miles:** Sign - turnoff to Hampton Creek. Continue straight.

**10.0 miles:** Unmarked turnoff to Smith Creek. Turn left and drive up 7 miles to the end of the road/trailhead near the mouth of Smith Creek. Park at trailhead.

**Stop 3.1** Smith Creek
Moderate hike (~2 mi roundtrip). Here we will examine some dramatic exposures of both the upper plate and lower plate of the northern Snake Range. As we hike up Smith Creek, the towering cliffs on either side consist largely of middle and upper Cambrian limestones of the upper plate. You will be able to see many, relatively small displacement normal faults and get a feel for the structural style of the upper plate. The NSRD here dips 10-15° to the east and is in the shallow subsurface beneath you. As you hike into the range, you are crossing into progressively deeper structural levels of the "dome" such that we will cross the NSRD, and enter into tlower plate rocks. Hike up Smith Creek approx 1 mile to where the old road ends in a large cottonwood grove. Just before you reach the cottonwood grove, veer off to the right and climb up a few hundred yards on the left (south) side of Smith Creek Canyon for a spectacular view of one of the major upper plate normal faults - exposed on the north wall of the canyon. There are also many smaller faults. The big fault is one of the down to the west faults - antithetic to the overall sense of slip on the NSRD- and here places upper Cambrian and Ordovician limestone against the lower part of the Middle Cambrian limestone. The NSRD is also well exposed here as it "Vs" down into Smith Creek. Continue upstream, through the cottonwood grove to the next big clearing. Deeply incised gorges into the lower plate expose the same lower plate metasedimentary sequence as we saw in Hendrys Creek - but here they have been invaded by very large amounts (50%) of leucogranite dikes and sills - thought to be mainly Late Cretaceous in age. Return to cars and drive back to North Snake Valley Rd., turn left and head north.

## THE GEOLOGY OF THE SMITH CREEK AREA

### The Northern Snake Range Décollement

In the vicinity of the Smith Creek drainage, the NSRD is spectacularly exposed at higher elevations as a nearly flat, exhumed fault surface at an elevation of ~11,000 feet, the highest exposure of the NSRD within the northern Snake Range. Toward Hendry's and Hampton Creeks, the NSRD dips gently (5-8°) and uniformly to the east. To the north-northeast, the NSRD dips 10-15° into the Smith Creek area, defining a map-scale, WNW-ESE trending trough-like depression or mullion, as the NSRD drops in elevation from 11,000 feet to < 7,000-8000 feet in the Smith Creek area. The axis of the trough is subparallel to the inferred WNW-ESE extension direction in both the lower and upper plates, and is located in a region where the polarity of upper plate normal faulting changes across Smith Creek from predominantly down-to-the-west to the north to predominantly down-to-the-east to the south. The trough may be the related to the formation of a transform fault, no longer exposed, in the upper plate that accommodated the change in polarity of normal faulting, or it may represent a primary structure, such as a fault mullion, that developed parallel to the movement direction along the NSRD.

Unmetamorphosed Middle Cambrian to Devonian rocks of the upper plate underlie much of the rugged, high, ridges to the north of Smith Creek and constitute perhaps the best exposures of upper plate extensional fault complexes in the northern Snake Range. Metamorphosed and strongly deformed Eocambrian metasedimentary rocks and Cretaceous plutonic rocks crop out in lower plate windows in the deeply incised headwaters of Smith Creek, and adjacent Horse Canyon, and Little Horse Canyon and have a well developed mylonitic foliation and WNW-ESE stretching lineation. Tilted Miocene sedimentary rocks are exposed locally along the eastern flank of the range and yield important insights on the youngest faulting responsible for uplift of the range.

## Upper Plate Structure

The structural style of upper plate rocks in the Smith Creek drainage region is similar to that of the northern Snake Range in general (Gans and Miller, 1983), with some important variations on the theme. All the upper plate units are cut by at least two and locally three superposed generations of faults. These faults all appear to be normal faults, as they consistently place younger rocks on older rocks and structurally thin the stratigraphic section to a small fraction of its original thickness. Faults are present at all scales, with displacements ranging from centimeters to kilometers. Only faults with offsets in excess of 50-100 m are shown on our maps, but these appear to account for most of the upper plate strain. Upper plate units generally young westward and upward, from predominantly massive Middle Cambrian rocks on the eastern flank of the range to Ordovician to Devonian strata along the high ridges and peaks to the north, south and west of Smith Creek. A composite stratigraphic thickness of more than 5 kilometers is represented by the rocks of the upper plate, but nowhere is there more than a kilometer or so of relatively unfaulted stratigraphic section, and the maximum structural thickness of the upper plate is generally even less than that.

Upper plate strata in and around Smith Creek can be divided into two structural domains that are separated by an east-west trending boundary approximately coincident with Smith Creek. South of Smith Creek, upper plate units dip mainly to the WNW and are cut by two and locally three generations of normal faults. These faults have predominantly top-to-the-east or -southeast displacement. North of Smith Creek, units dip mainly to the southeast and are cut by faults with predominantly top-to-the-northwest displacement. The boundary between these opposing tilt domains is poorly defined, but appears to be a kilometer-wide zone, rather than a discrete transfer fault. Within this zone, bedding attitudes are commonly south-dipping but extremely variable, and faults range from steep to low-angle, with varying senses of displacement. If there is a systematic structural style or sequence of faulting within this zone, it is not clearly apparent.

In the southern domain, bedding dips mainly to the northwest, but dips range from vertical to sub horizontal. Much of this variability appears to be a consequence of local drag in the vicinity of faults. The more intact sections dip ~40-60° to the northwest. Older (Middle Cambrian) units of the upper plate appear to dip less steeply than younger strata. It is not clear whether this is because the older rocks are actually less faulted and tilted or whether they have been affected by normal drag on the underlying NSRD. Along the eastern flank of the range, bedding locally rolls over from gently northwest-dipping to gently east-dipping, a feature we attribute to normal drag along a major east-dipping normal fault in this vicinity (see below).

The oldest faults in the southern domain dip 20-30° westward and are at relatively low angles to bedding. This generation of faults has been identified with confidence only in Upper Cambrian and younger rocks. Top-to-the-east or -southeast displacement can be inferred from the fact that these faults omit section and generally dip less steeply than bedding. The next younger set of faults is sub horizontal, at high angles to bedding in Upper Cambrian and younger units of their hanging walls but generally at lower angles to bedding in the Middle Cambrian footwall rocks. Slickensides and stratigraphic offsets indicate top-to-the-southeast offsets of at least a kilometer on some of these. The youngest normal faults dip moderately eastward and are at high angles to bedding in both hanging walls and footwalls. These youngest faults may in part be synchronous with and related to the major east-dipping fault along the eastern flank of the range.

The northern (east-tilted) domain of the upper plate in the Smith Creek region consists largely of Middle Cambrian rocks but also includes Upper Cambrian to Pennsylvanian rocks in the Mount Moriah quadrangle to the west. Bedding dips are variable but average ~30-40° to the southeast and east. Most of the larger faults strike northeast, dip moderately northwest, and have clear top-to-the-northwest displacements as indicated by offset markers, drag features, etc. Many such faults are spectacularly exposed on the northern wall of Smith Creek Canyon. Because of the lack of distinctive marker beds within the thick Middle Cambrian section, it is generally not possible to determine exact offsets on many of these faults or to resolve different generations of faults, as in the southern domain. Nevertheless, at least two generations appear to be present, an older, more gently dipping set of normal faults and a younger more steeply northwest dipping set. Several faults of the younger set, including the major west-dipping fault that juxtaposes Upper Cambrian and younger rocks against Middle Cambrian rocks in the northwest corner of the quadrangle, sole into the NSRD.

A major east-dipping fault is exposed along the eastern flank of the northern Snake Range, approximately coincident with the eastern limits of bedrock exposure. This fault dips 10-30° eastward and is inferred to have at least several kilometers of offset, as it juxtaposes tilted Miocene conglomerates in the hanging wall against faulted and tilted Middle Cambrian and younger rocks in the footwall. The Miocene rocks in the hanging wall now dip ~20-30° westward and are interpreted to represent the syntectonic basin fill that accumulated during uplift of the range and movement on this fault. The range-front fault is clearly younger than much of the faulting within the range, including the NSRD, because it cuts previously faulted and tilted rocks in the footwall; it continues uninterrupted across the domain boundary defined by the upper plate fault systems; and, most importantly, it cuts conglomerates that contain clasts derived from the whole spectrum of upper plate units as well as from mylonitic, lower-plate marbles, quartzite and orthogneiss. The source of these lower-plate clasts can be confidently assigned to the headwaters of Smith Creek in the Mount Moriah quadrangle to the west based on the distinctive clasts of Cretaceous orthogneiss and pegmatite exposed in this drainage. The conglomerate is inferred to be Middle Miocene or younger because the clasts derived from the lower-plate yield fission-track ages of 20-15 Ma and record the rapid cooling and exhumation of the lower plate (Miller et al., 1990).

The succession of faulting and tilting events in the Smith Creek region is interpreted to reflect progressive extension of large magnitude. The direction of extension documented by the older sets of faults (i.e., the faults within the range and restricted to the upper plate of the NSRD) is ~N45W-S45E on the basis of slip vectors on many of the faults and the average direction of tilting in both the northern and southern domains. The direction of movement on the younger range-front fault appears to be more easterly. The ages of the various generations of faults are not well enough constrained to evaluate whether they formed in response to one progressive extensional event or whether they reflect distinct faulting episodes, perhaps widely separated in time. We interpret the northwest-dipping faults in the northern domain to be broadly synchronous with at least the first two generations of southeast-dipping faults in the southern domain because they document similar magnitudes and directions of extension, and there does not appear to be a consistent crosscutting relationship between faults of the two domains. Thus, the poorly defined northwest-trending boundary between the two structural domains is interpreted to be an accommodation zone that approximately parallels the extension direction and separates opposing, coeval tilt domains. The only constraint on the older generations of faults is that they are post-Paleozoic, although we suspect they must predate or be synchronous with rapid, early to mid-Miocene cooling and exhumation of the lower plate (Miller et al., 1990). The youngest range-front fault must have continued to move during and after this mid-Miocene cooling, as it records the erosional breaching of the lower plate.

## Lower Plate Metamorphism and Structure

Eocambrian metasedimentary rocks and Cretaceous plutonic rock of the lower plate are exposed at low elevations on the flank of the range and in the bottom of Smith Creek and adjacent Horse and Little Horse canyons. Metasedimentary rocks include quartzite and schist correlated with the three upper units of the Precambrian McCoy Creek Group of Misch and Hazzard (1962), and the Lower Cambrian Prospect Mountain Quartzite and Pioche Shale, and marble correlated with lower Middle Cambrian Eldorado Limestone. Igneous rocks include a large biotite tonalite/granodiorite orthogneiss body known to be Early Cretaceous (Miller et al., 1988) and small leucogranite aplite and pegmatite bodies of Late Cretaceous age (Huggins and Wright, 1989). The scattered exposures of the Early Cretaceous orthogneiss (Kbg, also informally called the orthogneiss of Horse Canyon) all appear to be part of a single sill-like intrusive body, several hundred meters thick that extends at least 10-15 km in a WNW direction and several kilometers in a northeast direction. Its sill-like geometry is particularly obvious in the deeply incised headwaters of Horse Canyon and Little Horse Canyon, where both the upper and lower contacts are well exposed and dip very gently to the east or southeast. In detail, the intrusive contacts of the orthogneiss systematically cut up section toward the north or northeast, and we suspect that much of its present sill-like geometry and WNW elongation is a reflection of the large magnitude of superimposed Tertiary strain.

Lower plate units are locally pervasively injected by dikes and sills of leucogranite, aplite, and pegmatite. At least some of these are Late Cretaceous in age and broadly synchronous with amphibolite-facies metamorphism (Huggins, 1990). These were described as "lit-par-lit" injections by Misch (1960), but they were clearly a network of dikes and sills that have been strained into sill-like masses.

The highest grade metamorphic assemblages occur in the Smith Creek area, where peak metamorphic mineral assemblages include biotite + clinozoisite + muscovite/phlogopite + quartz ± plagioclase ± calcite ± sphene in the Upper Cambrian calc-schists and muscovite + quartz + biotite + plagioclase + garnet ± staurolite ± kyanite in Unit 2 of the McCoy Creek Group. Peak metamorphic temperatures of 600-650°C, based on mineral assemblages and garnet-biotite geothermometry, were reached in the Smith Creek area (Huggins, 1990). U-Pb analyses of metamorphic monazite from the schist units yield ages of 78 Ma, interpreted as the age of metamorphism (Huggins, 1990). Since it is likely that emplacement of the 82 Ma pegmatites was concurrent with peak metamorphism, we interpret the data to indicate that peak metamorphic conditions were reached between 78-82 Ma (Huggins, 1990). Deformation occurred during this metamorphic event, as evidenced by curved inclusion trails within metamorphic porphyroblasts. However, strong overprinting by a younger, Tertiary deformation prevents detailed study of the geometry and orientation of these fabrics.

Overprinting this Cretaceous metamorphic event is a middle to upper greenschist facies metamorphic and mylonitic deformational event that ductilely thinned and stretched lower plate rocks and collapsed older metamorphic isograds (Geving, 1987; Miller et al., 1989). As with the older event, grade of metamorphism increases with depth and from south to north and resulted in new growth of coarse-grained plagioclase, chlorite, and muscovite, and biotite growth at deeper levels in the Smith Creek region. Older metamorphic porphyroblasts are partially to completely replaced by sericite ± chlorite ± biotite. New(?) fibrous kyanite was observed along the edges of staurolite and garnet porphyroblasts and is aligned parallel to the predominant foliation. The predominant foliation is a mylonitic foliation defined by coarse-grained chlorite, muscovite, biotite and recrystallized quartz grains in the schist units, dynamically recrystallized, flattened and elongate quartz grains in the quartzites, and flattened and elongate calcite and color banding in the marbles. The mylonitic foliation is subparallel to contacts of metasedimentary units, and subparallel to the overlying NSRD. Within the mylonitic foliation is a WNW-ESE trending mineral elongation lineation, defined by stretched quartz and calcite grains, aligned older metamorphic porphyroblasts, and pressure shadows on older porphyroblasts. Mesoscopic asymmetric kinematic indicators, such as S/C foliations in the orthogneiss, extensional crenulation cleavages in the schist units, and shear bands indicate a component of top-to-the-east sense of shear during formation of the mylonitic foliation and stretching lineation. Strain associated with the mylonitic deformational event increases from west to east, as clearly evidenced by the dramatic thinning of lower plate units. The thickness of the Prospect Mountain Quartzite decreases from about 20% of its original stratigraphic thickness in the region of The Table to about 6% at the junction of Deep Canyon and Smith Creek.

The high-strain fabric in the lower plate of this quadrangle presumably correlates with the younger ($S_2$) fabric elsewhere in the range, as it clearly postdates peak metamorphism (metamorphic minerals are pulled apart, with new growth of chlorite-white mica in pressure shadows) and occurred at temperatures where calcite and quartz were behaving ductilely. However it should be emphasized that the absolute age(s) of the high-strain deformation remains poorly constrained. Miller et al. (1983) inferred it to be largely Oligocene and synchronous with extensional faulting in the upper plate, which at the time was thought to be mainly Oligocene (Gans and Miller, 1983). Lee and Sutter (1991) carried out an $^{40}Ar/^{39}Ar$ study of the lower plate and also concluded that the mylonitic deformation was largely Oligocene on the basis of ~37 Ma ages on rhyolite porphyry dikes that appeared to predate the deformation and 24-25 Ma ages of metamorphic micas interpreted to reflect post-deformational cooling of the lower plate. However, given all of the uncertainties (e.g., correlation of fabrics, temperature(s) of deformation, mica closure temperatures), the data are still permissive of the high strain deformation on the eastern side of the range being mostly, if not entirely Miocene (post-24 Ma), Oligocene (37-24 Ma), or pre-37 Ma. The only firm minimum and maximum age brackets are provided by the age of peak metamorphism (latest Cretaceous) and the final cooling and exhumation of the lower plate at approximately 20-15 Ma (Miller et al., 1990).

Return to main Snake Valley road, turn left (north), reset odometer to **0.0**

**9.5 miles:** Turn left at minor intersection and angle northwest towards Warm Springs Ranch by large stand of trees
**10.6 miles:** turn left by telephone pole and continue bearing left by blue trailer home on right. Pass old corrals on left
**11.0 miles:** Road crosses irrigation ditch (just past corrals) - take the first "soft" left on a dirt track that appears headed straight back towards the northern Snake Range. Follow this road southwest towards the mouth of Marble Wash.
**16.0 miles:** Y in road, bear right. Follow road down into Marble Wash

**Stop 3.2** Marble Wash
Here we will make several stops to examine superb examples of marble mylonites, boudinaged mafic dikes and sills, calc schists, and examples of large magnitude heterogenous strain on all scales. In addition, we will hike up through the lower plate sequence and put our fingers on terrific exposures of the NSRD.

## STRUCTURAL AND STRATIGRAPHIC RELATIONSHIPS IN THE MARBLE WASH AREA

Isolated remnants or klippe of the upper plate of the NSRD are present above vast tracks of subhorizontally foliated marbles of the lower plate in the Marble Wash area. These rocks record the polyphase deformational-metamorphic-igneous history typical of rocks elsewhere in the northern Snake Range, which in the upper plate includes at least two generations of normal faults, and in the lower plate includes both Cretaceous and Tertiary metamorphism and ductile deformation. The NSRD dips gently northeastward off this flank of the range and is inferred to lie in the shallow subsurface beneath the alluvium to the east of the range. A broad, gently east-dipping pediment surface of Quaternary alluvium (Qol) that slopes down and merges with the highest Lake Bonneville beach terraces (Qbs). This pediment is deeply incised (up to 40 m) by the modern drainages (e.g., Marble Wash and Swan Creek), reflecting the progressive drop in base level for the Quaternary catchment basin.

### Lower Plate Rocks

Middle and Upper Cambrian rocks of the lower plate are spectacularly exposed in Marble Wash. They include calcite marble, dolomitic marble, and calc-schist. Scarce sill-like-bodies of diorite also occur in the lower plate. Metamorphic minerals present include high-Mg biotite, phlogopite, muscovite, quartz, calcite, dolomite, plagioclase, tremolite/actinolite, clinozoisite, and sphene. In the calc-schists, a common assemblage is phlogopite or biotite ±muscovite±quartz±clinozoisite±plagioclase±calcite. These minerals and the absence of index minerals such as diopside, wollastonite, garnet, and scapolite suggest that the rocks last equilibrated at upper greenschist facies conditions. The age of the peak metamorphism in this area has not been dated directly but is thought to be late Cretaceous based on U-Pb monazite ages of high grade pelites farther south in the range (Lee and Fischer, 1985, Wright and Huggins, 1989).

The lower plate units are generally right side up and in correct stratigraphic order but are now marble and calc-schist tectonites with a predominant gently east-dipping mylonitic foliation and a well developed east-southeast trending mineral elongation lineation. The penetrative strain has attenuated the stratigraphic section to a small fraction of its original thickness, such that complete sections of Dunderberg Shale are often only a few meters thick, as opposed to an original stratigraphic thickness of 60-100 meters. Elsewhere along the eastern flank of the range, complete sections of lower plate units are commonly thinned to ~10% of their original thickness, largely by plastic flow (Miller et al., 1983; Lee et al., 1987). In detail, the character, orientation, and magnitude of lower plate fabrics in this quadrangle are highly variable and complex, a consequence of both heterogeneous strain and a polyphase history. Intraformational folding on scales ranging from centimeters to hundreds of meters is widespread, particularly in lower plate marbles. These folds are mainly isoclinal recumbent folds with hinge lines that parallel the mineral elongation lineation, but other orientations and more open varieties also exist. Boudinage of more resistant layers (mafic sills, dolomitic marble, calc-schist) on scales ranging from centimeters to hundreds of meters is spectacularly developed and ubiquitous (Gaudemer and Tapponnier, 1987; Lee, 1990). Stunning examples of such boudinage can be observed on the walls of canyons in the lower plate. Here, low strain lozenges of dolomitic marble, calc-schist, and diorite are set in a swirly, banded matrix of blue and white calcite marble mylonite. Indeed, much of the complexity to lower plate structural fabrics, including variations in fold geometry and orientation, multiple foliations, and conflicting shear sense indicators, appears to be a direct consequence of the heterogeneous strain and quasi-turbulent flow field around more resistant blocks. For example, shear bands, trains of asymmetric folds, and multiple transposition foliations within the banded marble mylonites are best developed on the flanks of large (10-100 m) lenticular bodies of dolomite and calc-schist and typically verge in opposite directions on either side of the same boudin

.

Drive back out (retracing path) to Snake Valley Road. Turn left (north). Set odometer to zero.

**10.6 miles:** Intersection of Snake Valley Road and Pleasant Valley Road. Turn west (left) on Pleasant Valley Road. As we drive along, view northward of fault-bounded Deep Creek Range with the Tertiary Ibapah Granite along the skyline. View southward of Mt. Moriah, northern Snake Range and The Table, basically the exhumed surface of the Snake Range decollement. Gently west tilted Tertiary conglomerates underlie Pleasant Valley and rest on Tertiary volcanic section exposed at the very eastern and of the valley. To the north in the Deep Creek Range, the overturned Water Canyon anticline involving Precambrian to Cambrian age rocks is visible.

**19.4 miles:** Skinner canyon turnoff, Skinner Canyon, eastern Kern Mountains.

**Stop 3.3** Skinner Canyon

This stop will involve a short hike up the mouth of Skinner Canyon to see a variety of highly deformed metasedimentary and igneous rocks which occur as a roof pendant and as wall rocks to the Oligocene Skinner Canyon pluton. Particularly interesting are the map relations which demonstrate the relative timing of events regarding the intrusion of the pluton with respect to deformation of the country rocks. In addition we have also been able to place fairly precise brackets on the absolute timing of events and this data will be discussed at this stop. Based on map relations and the timing of events, we think that the Skinner Canyon granite represents an excellent example of a pluton that was emplaced in the crust syntectonically with extension.

## OVERVIEW OF THE KERN MOUNTAINS-DEEP CREEK RANGE

The Kern Mountains and Deep Creek Range lie north of the Snake Range and appear to constitute a single, west-tilted structural block. The geology of these ranges has some important similarities to that of the Snake Range, but also some important differences. On the eastern flanks of the Deep Creek Range and Kern Mountains, polymetamorphosed and ductilely deformed metasedimentary and granitic rocks are separated from unmetamorphosed but highly faulted upper Paleozoic and Tertiary rocks by a major, east-dipping low-angle normal fault that closely resembles the northern Snake Range Decollement. This fault has tens of kilometers of top-to-the east displacement and appears to omit kilometers of structural section (Gans et al., 1986). A narrow ridge of faulted Paleozoic strata can be followed continuously from the upper plate of the northern Snake Range to the eastern Kern Mountains where it lies in the upper plate of this low angle fault. Thus, we believe that much of the Kern Mts-Deep Creek Range block represents the northward continuation of the footwall to the NSRD. However, unlike in the northern Snake Range, much of the footwall in these ranges is unmetamorphosed, lacks any penetrative ductile fabric, and shows clear evidence for significant Tertiary tilting of what was once an intact upper crustal section.

The southern Deep Creek Range is underlain by a west-dipping, 13 km thick miogeoclinal succession that ranges from the Upper Precambrian on the east to Upper Permian on the west (Rodgers, 1987). This section is in turn overlain by Eocene volcanic rocks that dip 30-50° to the west, indicating that the entire range was tilted westward in the Tertiary (Fig. 30). Metamorphic grade increases systematically downsection, from totally unmetamorphosed on the west (Conodont Alteration Indices of 1-1.5) to upper greenschist facies, locally amphibolite facies on the east. The metamorphic history is complex, with evidence for both Cretaceous and Tertiary dynamothermal metamorphism in different areas, steep metamorphic field gradients, and evidence for a close association of metamorphism with plutonic activity (Rodgers, 1987). To the north, the lower part of this succession is intruded by a large biotite granite, known informally as the Ibapah Granite, that has been dated by U-Pb zircon at 39 Ma (Miller et al., 1988, Rodgers, 1987) and at 36 Ma by $^{40}Ar/^{39}Ar$. In the southernmost part of the range, the lower part of the section is folded into a large recumbent anticline - the Water Canyon Anticline- that closes to the west and plunges gently to the north. This fold has a subhorizontal to north-dipping axial surface, such that its overturned limb continues into the northeastern Kern Mountains, where it becomes metamorphosed to amphibolite facies and is intruded by granites of both Cretaceous and Tertiary age (Fig. 30).

Most of the Kern Mountains is underlain by granite. The western two thirds consists of the distinctive 2-mica Tungstonia Granite that contains spectacularly large phenocrysts of muscovite up to 5 cms in diameter (Best et al, 1974). This granite has yielded highly discordant U-Pb ages on three fractions of zircon that define a chord with a lower intercept of 75± 9 Ma (Lee et al., 1986) and $^{40}Ar/^{39}Ar$ muscovite ages as old as 69.5 Ma (Gans et al, 1986). This granite is considered a classic example of the Cordilleran muscovite granites (often referred to as S-type granites) that are largely late Cretaceous to early Tertiary in age (Miller and Bradfish, 1980). These granites typically define a belt inboard and are slightly younger than the main part of the Mesozoic magmatic arc and are compositionally distinct in their strongly peraluminous compositions. As such, they are thought to be largely the result of crustal melting associated with intracontinental thickening during the Late Cretaceous to Early Tertiary Sevier and Laramide orogenies.

The eastern third of the Kern Mountains is underlain by a biotite granite, the Skinner Canyon Granite, that has yielded a concordant U-Pb zircon age of 35±1 Ma (Miller et al., 1988) and a $^{40}Ar/^{39}Ar$ plateau age of 36 Ma on metamorphic hornblende from the sillimanite grade aureole (P. B. Gans, unpublished data).

The metasedimentary wall rocks of these plutons exhibit a polyphase deformational history, in part related to the emplacement of the granites. Throughout the western part of the Kern Mountains, both the outer part of the Tungstonia Granite and the adjacent country rocks have a very well developed high-temperature gneissic fabric that parallels the contacts of the granite. This fabric appears to represent a pure flattening that was developed during the emplacement and final crystallization of the granite. Toward the east, this Cretaceous synplutonic fabric is overprinted by a penetrative, subhorizontal to gently east-dipping foliation with a well-developed ESE stretching lineation. This fabric is particularly strongly developed in the metasedimentary rocks and appears to be associated with the development of a major east-dipping, low-angle fault that bounds the range, as stretching lineations and fault striations are parallel and indicate top to the ESE sense of shear. Rhyolite and diorite dikes that have been dated at 39 Ma in the western part of the range are mylonitized by this younger fabric in the east, thereby demonstrating a post Eocene age for the fabric. Relations between the 36 Ma Skinner Canyon Granite and the younger fabric are ambiguous. The main bodies of the Skinner Canyon Granite are undeformed, whereas leucocratic dikes and sills associated with the granite are strongly boudinaged by this deformation. Either the Skinner Canyon granite was broadly syntectonic with this deformation, or the deformation is entirely younger than the 35±1 Ma granite, but rheologic contrasts dictated that strain was restricted to the wall rocks.

A detailed $^{40}Ar/^{39}Ar$ and apatite fission track thermochronologic transect across the Kern Mountains revealed important differences in the thermal history of the western part of the range relative to the eastern part. Muscovite ages within the Tungstonia Granite systematically decrease towards the west from 69.5 to 36 Ma, a trend we interpret to reflect progressive degassing by the Skinner Canyon Granite, despite the fact that most of these spectra are relatively flat. Potassium feldspar ages show similar trends, but on the eastern side of the range, show a prolonged, post- 35 Ma cooling history. Apatite FT ages are as old as 38 Ma on the west side of the range and cluster around 15-18 Ma on the east. Thus, the data are compatible with the evidence for westward tilting of the range, but do not place firm brackets on the timing of this tilting.

A more revealing thermochronologic study was carried out across the Ibapah Granite in the Deep Creek range to the north (Gans et al, 1991). $^{40}Ar/^{39}Ar$ incremental heating analyses of muscovite, biotite, and K-feldspar together with zircon and apatite fission track analyses revealed a simple pattern of monotonically decreasing ages from the western crest of the range to the eastern flank - downward through an inferred structural section of about 7 km. These age profiles together with consideration of the expected cooling history within an extensional tilt block suggest that the main episode of unroofing and westward tilting associated with rapid slip on the major range bounding fault (i.e., the low-angle normal fault) began 18 Ma and was largely completed by about 14-15 Ma. An earlier episode of rapid slip may have occurred 36-35 Ma but is not required by the data. A handout includes additional data on the thermochronology of the Deep Creek Range and can be found at the end of the Figures file.

## GEOLOGY OF SKINNER CANYON

Gently dipping units at the mouth of Skinner Canyon represent portions of the Middle Cambrian section ($Cm_1$, $Cm_2$, $Cm_3$) as determined by their stratigraphic position above the Pioche Shale and Prospect Mountain Quartzite (exposed to the south of the mouth of Skinner Canyon), and beneath the Ordovician Pogonip Group marbles to the west, exposed higher in the canyon (Fig. 31). The Paleozoic units have been metamorphosed to amphibolite facies such that they now consist of coarse grained marbles and calc-silicates. Other rock types include dark-coloured diorite and lighter, reddish-weathering muscovite porphyry (patterned on map). All of the metasedimentary rocks as well as the diorite and muscovite porphyry bodies are highly deformed and exhibit high strain to mylonitic type fabrics and textures in hand specimen and thin section. Foliation is gently dipping and parallel to (transposed) unit boundaries. Lineations associated with this foliation trend W-NW to E-SE and give consistent top-to-the-east shear sense. Foliation development appears to have occurred at amphibolite facies metamorphic conditions, as indicated by growth of aligned metamorphic hornblende in the diorite dikes and the presence of syntectonic sillimanite and andalusite in more pelitic horizons in the Pogonip Group.

Higher up the canyon, and closer to the contact with the Skinner Canyon granite, tabular masses of aplite and granite intrude the section. These masses are deformed along their margins and boudinaged within the country rock foliation. They clearly share at least a portion of the deformational history of the country rocks. In general, the intrusive contact of the Skinner Canyon granite is complex and cross-cutting, and although dikes and offshoots are often highly deformed, the main body of the granite itself is coarse, equigranular and undeformed. The deformation of the aplite dikes occurs at fairly low temperatures, so is interpreted as late.

These relations are taken together to suggest that the intrusion of the pluton was broadly synchronous with wall-rock metamorphism and deformation, yet most of its crystallization must have occured after deformation had ceased at these specific levels of the crust. It is also possible that deformation may have continued or was renewed at a younger time beneath the fault system but that the high feldspar content of the main part of the granite precluded it from being deformed during this event.

Hornblende-bearing diorite and muscovite porphyry bodies mapped in the Skinner Canyon area represent deformed equivalents of an unusual dike swarm exposed in the western part of the Kern Mountains, where they are not deformed or metamorphosed. Here muscovite from muscovite porphyry dikes has been dated at 38 Ma ($^{40}Ar/^{39}Ar$) (Gans et al., unpublished data). In the aureole of the Skinner Canyon Granite, metamorphic hornblende from associated diorite dikes yields an $^{40}Ar/^{39}Ar$ plateau age of 35±1 Ma (Gans et al., unpublished data). U-Pb dating of zircon from the Skinner Canyon granite yields an age of 36 Ma (Miller et al.,1988). Lamprophyre dikes which cross-cut all fabrics further to the west yield $^{40}Ar/^{39}Ar$ (hbl) ages of 23 Ma (Gans et al., unpublished data).

The metamorphic complex is cut by the low-angle fault system that wraps around the eastern end of the range, and underlies gently tilted Miocene conglomerates in Pleasant Valley. Apatite fission-track cooling ages from the Skinner Canyon Granite and metasedimentary rocks on the eastern flank of the range are 15-18.Ma. Muscovite $^{40}Ar/^{39}Ar$ ages from this part of the range are 28-32 Ma. K-feldspars tend to yield pronounced age gradients from 17 to 28 Ma. Thus, the final uplift and unroofing of this part of the range did not occur until the mid-Miocene, much like the northern Snake Range.

## ROAD LOG-SKINNER CANYON TO BLUE MASS

Retrace tracks and go back out to Pleasant Valley Road.

**0.0 miles:** Reset odometer to 0.0 miles at mouth of Skinner Canyon, before heading west to Blue Mass Canyon.

Driving west along Pleasant Valley, good views to south of Skinner Canyon Granite. Further west, the Skinner Canyon is intrusive into the Cretaceous Tungstonia granite (Fig. 30).

**10.1 miles:** Blue Mass turnoff, turn left and drive into the Kern Mountains.

**12.7 miles:** Abandoned ranch house, park.

**Stop 3.4** Cretaceous Tungstonia granite

Here you will get to examine one of the classic muscovite granites that lies within a belt of late Cretaceous to early Tertiary "S-type" granites east of the main Mesozoic magmatic arc. These granites are all strongly peraluminous, have isotopic signatures that indicate they were derived largely or entirely from crustal melting, and are coeval with the latter stages of a major crustal thickening event (Miller and Bradfish, 1980; Miller et al, 1988; Wright and Wooden, 1991).

The Tungstonia Granite is a late Cretacous (~75 Ma) granite to granodiorite (~70% $SiO_2$, plag > Kspar) that is remarkable for the size and abundance of phenocrystic muscovite it contains (Best, 1974). Individual books of muscovite are commonly 3-5 cm in diameter and are riddled with tiny biotite inclusions. The granite is isotropic in its interior, but becomes strongly foliated parallel to its margins - a fabric we interpret as protoclastic. Here we are camped near its western limit at a position that we believe is close to the original roof of the pluton. A shallow level of emplacement for this end of the pluton is suggested by the narrow, low Temp contact aureole into rocks that had previously never been heated above about 150° C (Gans et al, 1986)

We will camp here tonight.

## DAY 4

From our campsite at Blue Mass, we drive to the Schell Creek Range to the west to look at Eocene-Oligocene syn-extensional volcanic rocks. Snow conditions permitting, we will continue over Kalamazoo Pass (great views) to Highway 93 and north to Wells, Nevada (Fig. 34). Alternatively, we will drive back out to Spring Valley and north over Schellbourne Pass to Highway 93 and north to Wells. Rendevous with Art Snoke in Wells City Park for lunch, gas, etc. for a short overview of the geology of the

East Humboldt Range. We will spend the afternoon visiting sites in the East Humboldt Range, low-angle normal faults, and mylonites. Depending on snow conditions,we will see rocks in the core of the range at Angel Lake cirque (2-3hr hike). We will camp tonight at the Angel Creek campground in the East Humboldt Range. Handouts on field trip stops will be given to you at this time.

## ROAD LOG- BLUE MASS TO WELLS, NEVADA

Drive back out to Pleasant Valley Road, retracing route.
Reset odometer to **0.0** turn left.
**7.3 miles:** Intersection, veer left.
**15.1 miles:** Tippett an abandoned dude ranch. Still a great place to come whoop it up on Friday nights!
**26.1 miles:** Roads join. As you continue driving, pass volcanic rocks on right. that are part of the 40-35 Ma andesites in the region
**31.4 miles:** Intersection. Take road to N. Spring Valley Road/Hwy 50 or Schellbourne....
**39.5 miles:** Intersection .
**76.1 miles:** Turnoff to Mike Springs. Kern Mountains and Deep Creek Range to north, northern Snake Range to the south.

Crossing Spring Valley we are driving on the highest beach terrace of the Pleistocene lake that filled the area during the last glaciation. From here have a superb view southward of Spring Valley - a typical half-graben in the Basin and Range province, and the topographic escarpment of the eastern side of the Schell Creek Range formed by movement on the range-bounding Schell Creek fault. This fault and the subsurface geometry of the basin was nicely imaged by a seismic line described by Gans et al. (1985). The Schell Creek fault dips eastward ~45° beneath Spring Valley and is estimated to have 8-10 km of displacement.
**85.5 miles:** North Spring Valley Road. Turn south or left.
**85.9 miles:** Beginning of pavement.
**87.4 miles:** Kalamazoo Creek turnoff, turn right, reset odometer to **0.0**.

We will now drive westward into the northern Schell Creek Range in order to examine some of the late Eocene-early Oligocene volcanic rocks that were erupted in part synchronous with extension. Like in the other ranges in east-central Nevada, the Paleozoic and Tertiary rocks in the Schell Creek Range are highly faulted and tilted. Tilts are generally to the west, such that the oldest rocks (Upper Precambrian and lower Cambrian) are exposed on the east flank of the range and progressively younger rocks are exposed to the west. A major low angle fault along the crest of the range separates a fairly intact Middle Cambrian and older section below from a much more faulted Upper Cambrian and younger section above and may represent the westward continuation of the NSRD.

**3.0 miles:** Massive cliffs and talus of quartzite of the Late Precambrian Mc Coy Creek Group.

**3.8 miles:** White cliffs of Middle Cambrian Raiff Limestone

**6.6 miles:** First hairpin turn, field trip stop to see massive cliffs of Kalamazoo Tuff. Refer to geologic map in Fig 33.

**Stop 3.5** The Kalamazoo Tuff in upper Kalamazoo Creek.

Here we will spend about 45 minutes examining the type section of the Kalamazoo Tuff (Gans et al, 1989). This 35 Ma tuff is the only regionally extensive unit in this part of White Pine County and thus serves as an important stratigraphic/time marker. Here, the unit is about 250 m thick and consists, in ascending order of (1) a thin, light tan, unwelded base, (2) a massive cliff-forming grey to black partially welded vitrophyre, (3) a slope forming interval of pinkish, densely welded, devitrified, and vapor phase altered tuff, and finally (4) a ledge forming, brown, densely welded upper section. The tuff is zoned upwards from a crystal-poor rhyolite at its base, containing ~ 5% crystals of san-plag-bio to a dacite with 25-30% crystals of plag-hbl-bio-cpx. The lower part of the tuff contains abundant lithic clasts.(largely of upper Paleozoic limestone) and fiamme up to 20 cm, wheras the upper portion lacks lithics and has only cm scale fiamme.

## VOLCANIC STRATIGRAPHY OF EAST-CENTRAL NEVADA

Tertiary rocks in east-central Nevada can be conveniently divided into three distinct groups. The oldest is late Eocene to early Oligocene in age and consists of local accumulations of andesitic to dacitic lavas, rhyolitic tuffs and lava flows, and intercalated lacustrine deposits and conglomerates. These older sequences are overlain by the more voluminous 35 Ma Kalamazoo volcanic rocks (Young, 1960); in ascending order, they include the Kalamazoo Tuff, hornblende dacite lavas, and the tuff of North Creek. The youngest group ranges in age from middle Oligocene to Miocene and consists of thick local accumulations of fanglomerate and lacustrine limestone, and rare interstratified tuffs. A composite stratigraphic column summarizing typical thicknesses, geochronologic data, compositions, and mineral modes from selected units within each of the three groups is illustrated in Figure 32.

The oldest volcanic rocks in the study area are exposed at widely separated localities in the northern Egan Range, the central and northernmost Schell Creek Range, and southwestern Deep Creek Range. They consist mostly of andesitic lavas and rhyolitic tuffs, and, for the most part represent small-volume eruptions from separate volcanic centres. In addition, most of the Tertiary granites in the study area are within the age range of this older suite of rocks and appear to represent the intrusive equivalents of the early rhyolites.

The name Kalamazoo Volcanics was first used by Young (1960) for a sequence of volcanic rocks exposed in the Kalamazoo Creek area of the north-central Schell Creek Range. Gans et al. (1989) redefined the term Kalamazoo volcanic rocks to encompass three areally extensive members; the Kalamazoo Tuff, Hornblende Dacite lavas, and an ash-flow tuff, informally called the tuff of North Creek. Complete sections of all three members are well exposed in the vicinity of Kalamazoo Summit in the northern Schell Creek Range.

The Kalamazoo Tuff is the most areally extensive ash-flow tuff in east-central Nevada. Its distinctive compositional and mineralogical zonation, well developed eutaxitic texture, and abundant carbonate lithic fragments distinguish it from all other eruptive units in east-central Nevada. Exposures of the Kalamazoo Tuff are concentrated in the north-central Schell Creek Range, but occur throughout an elliptically shaped area that extends 140 km from the Butte Mountains to the Confusion Range. It generally occurs at or near the base of exposed Tertiary sections, resting disconformably on either upper Paleozoic strata or the early Oligocene andesites, rhyolites and sedimentary sequences described above. Complete sections as thick as 350 m are well exposed in its type area in the head waters of Kalamazoo Creek (Fig. 33).

Crystal-rich hornblende dacite lavas, together with associated breccias and shallow intrusions, constitute the most voluminous rock types in the area. Their distribution resembles that of the Kalamazoo Tuff, with a similar east-west dimension but an even greater north-south extent. Although it is possible that the more peripheral exposures of hornblende dacite may include some unrelated lavas, their consistent stratigraphic position, together with limited geochemical and geochronological data, suggest that they are comagmatic. The hornblende dacite lavas typically overlie the

Kalamazoo Tuff or the older andesites and rhyolites, and are often the youngest eruptive unit exposed in any given area. Because their age, modal mineralogy, and composition closely resemble the last erupted portion of the Kalamazoo Tuff, we believe the two are genetically related.

The tuff of North Creek is the youngest and volumetrically least significant member of the Kalamazoo volcanic rocks. Its areal distribution and age overlap with that of the two older members, but its modal composition and geochemistry are distinct. The tuff is exposed in two principal areas: (1) east of North Creek at the north end of Duck Creek Valley in the Schell Creek Range; and (2) along the southern and eastern margins of the Kern Mountains. In many of the exposed sections, the tuff is separated from the underlying hornblende dacite lavas by a thin interval of pale-green, water-laid tuff and tuffaceous sandstone.

### Relative timing of extensional faulting and magmatism

The eruption of the older andesites and rhyolites heralded the onset of volcanism in east-central Nevada. They were erupted intermittently over the 5 m.y. interval prior to eruption of the more voluminous Kalamazoo volcanic rocks and may represent premonitory eruptions recording the coalescence of a batholith-sized magma chamber at shallow levels in the crust (e.g. Steven and Lipman, 1976). These older sequences are poorly preserved but were apparently derived from small eruptive centers scattered throughout the study area. None of the individual units appears to be very extensive; but the total volume of erupted material is almost certainly greater than 100 $km^3$. Their compositions are somewhat bimodal; rhyolite (commonly high silica rhyolite) and andesite or low-silica dacite predominate. The most mafic volcanic rocks were erupted from centers that presently lie peripheral to the main exposures of Kalamazoo volcanic rocks, suggesting that a higher proportion of low-density silicic melts accumulated in the central part of the study area and behaved as a density filter, impeding the ascent of more mafic magmas. The onset of extensional tectonism in east-central Nevada is inferred to have conincided with the eruption of the early andesites and rhyolites because: (1) intercalated conglomerates locally contain debris that records the denudation of deep structural levels in adjacent areas, and (2) feeders for some of these volcanic rocks locally cut the earliest extensional faults.

The Kalamazoo volcanic rocks compose the bulk of the erupted material, with an estimated original volume of approximately 1 000 $km^3$. All three members were erupted in less than 1 Ma. at about 35 Ma. Their source area is inferred to be either in the northern Schell Creek Range or buried beneath the northern part of Spring Valley. The Kalamazoo Tuff has a volume of approximately 300 $km^3$ and is zoned upward from a crystal poor rhyolite base to a crystal-rich hornblende dacite top. The similar composition and mineralogy of the last erupted part of the Kalamazoo Tuff and the overlying 500+ $km^3$ of hornblende dacite lavas suggest that the former may represent a compositionally zoned cap and the latter a sampling of the dominant volume of a dacitic magma body of batholithic dimensions. The age and distribution of the younger and smaller tuff of North Creek resemble those of the older two members, but it is more alkalic in bulk composition and contains abundant pyroxene rather than hornblende. Evidence that extensional faulting was contemporaneous with eruption of Kalamazoo volcanic rocks includes the following: (1) The Kalamazoo Tuff appears to overlap a greater amount of structural relief than do the older volcanic sequences. (2) Angular unconformities are locally developed between the hornblende dacite lavas and the tuff of North Creek. (3) A comagmatic hornblende dacite intrusion in the northernmost Snake Range appears to postdate some of the extensional faulting in that area.

Exposures of the younger sedimentary and volcanic sequences are widely separated but remarkably similar. Fanglomerate typically overlies and predominates over lacustrine deposits. Clast contents record the unroofing of some of the deepest exposed structural levels. Intercalated monolithologic slide masses suggest that steep topographic scarps existed. Dips typically decrease upward within the sequences indicating syndepositional growth faulting. Sparse radiometric age data from thin intercalated silicic tuffs suggest that the lower parts of these sequences are mid- to latest Oligocene, but their upper age limit is unconstrained. They typically overlap the older generations of normal faults unconformably but are involved in the youngest faulting. On both the northern and southern flanks of the northern Snake Range, faults that imbricate these sequences appear to sole into the Snake Range Decollement. We believe the younger sedimentary and volcanic sequences represent separate basins or half-grabens that first began to form during the middle to late Oligocene along the youngest generation of east-dipping normal faults currently exposed in the ranges. We have no upper age constraint on this episode of faulting and sedimentation and do not know whether it entirely preceded or was continuous with the formation of the present basins and ranges (see discussion in Gans et. al., 1985).

In many parts of the Basin and Range province, voluminous intermediate to silicic volcanism was followed by much smaller eruptions of basalt (±rhyolite). The only evidence for late-stage bimodal volcanism in east-central Nevada are rare rhyolitic tuffs (such as the 24-Ma tuff in the northernmost Snake Range) and rare 23 Ma ($^{40}Ar/^{39}Ar$, hbl) lamprophyre dikes in the eastern Kern Mountains. In addition, Hose and Blake (1976) reported a few 21 Ma olivine basalt flows in western White Pine County.

We have purposefully highlighted those field relations which suggest that volcanism and extensional volcanism in east-central Nevada were synchronous. However, in most areas there are neither major angular unconformities within the volcanic sections nor between the volcanic rocks and the underlying upper Paleozoic basement, which suggests that much of the faulting and tilting postdated Oligocene volcanism. The lack of more evidence for syndepositional tectonism within the volcanic sequences reflects the relatively short time interval (~1 My) during which most of the volcanic rocks were erupted, and the fact that that Tertiary rocks are generally best preserved in the less extended areas, having been largely eroded from the most highly extended and uplifted areas such as the Snake Range and the central part of the Schell Creek Range. Given the limited preservation of Tertiary rocks and the absence of dateable units within the upper parts of the Tertiary sequences, it is not possible to accurately assign amounts and rates of extension to various time intervals. From the available data we can say with confidence only that extensional faulting in east-central Nevada began at least 36 m.y. ago during some of the earliest volcanism, was ongoing at the time of the major outpourings of volcanic rocks at 35 Ma, and continued into the early Miocene for an unknown amount of time during deposition of the younger sedimentary and volcanic sequences. Locally there are Quaternary scarps on some of the major range-front faults.

If snow permits, we will continue driving up the steep, winding road to Kalamazoo Summit (~2-3 miles). Please drive very carefully. Stop briefly at the summit for an overview of Duck Creek Valley and geology of the northern Schell Creek Range.

Continue southwest down into Duck Creek Valley until you hit the pavement, turn left and follow highway out through Gallagher Gap. Nice sections of steeply tilted Ordovician to Devonian dolomites exposed on right side of road in Gallagher Gap, as well as exposures of the Ordovician Eureka Quartzite. Continue to T intersection with highway 93. Turn right and reset odomoter to **0.0**.

**21.0 miles:** Schellbourne Station - site of one of the original Pony Express stations.

**35.0 miles:** Lages Junction, continue beyond gas station and then turn left on U.S. 93 towards Wells and Currie.
**51.0 miles:** Currie Nevada. Good place for a pit stop
**113 miles:** Wells, Nevada

Alternate Route if pass is still under snow: Turn around and drive back to intersection with North Spring Valley Road. Go north towards Schellbourne Pass and then to Highway and north to Wells.

Wells city park for lunch at 12, 1:30pm latest (bathrooms, tables, water, shade, etc). Short overview of the geology of the East Humboldt Range, drive to first stop, overview of range and excellent exposure of low angle normal fault. Depending on snow situation proceed to Angel Lake cirque for exposures of the core of the range (2-3 hr hike), drive down the hill to Angel Creek campground for the night. (All field trip guide materials will be given to you at the stop.)

The following references are for the entire field trip. We have omitted the Ruby and Raft River part of the field trip in this version of text.

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